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Cenozoic Mountain Building and Climate Evolution

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Cenozoic Mountain Building and Climate Evolution[1]

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Abstract

Climate and land surfaces are intimately linked. This manuscript provides an introduction to mountain and climate interactions, with a particular focus on how temperature, atmospheric circulation and precipitation change with surface uplift. Although the timing and mechanisms of the South American Andes and North American Cordillera uplift are uncertain, both orogens have a first- order control on regional climate. With surface uplift, climate models generally suggest enhanced windward precipitation, orographic deflection of prevailing winds and decreased surface temperatures. However, the details of these responses are unique to each orogen and dependent on geographic position. In South America, Andean uplift led to a shift in prevailing winds from a Pacific source to an Atlantic source, development of a low-level jet along the eastern flank, increased moisture transport from the tropics and enhanced convective rainfall along the Andean lowlands and eastern flank. In North America, topographic blocking enhanced windward precipitation and initiated monsoonal conditions along the eastern flank, modified atmospheric circulation to create a pronounced trough-and-ridge pattern over western North America and formed a rain shadow over central North America. These changes add to the mechanistic understanding of regional Cenozoic climate and environmental change.

Keywords: surface uplift, regional climate, North American Cordillera, South American Andes

1. Introduction

Mountain ranges and high topography provide a fundamental control on Earth's climate. These features steer atmospheric circulation patterns, focus precipitation and aridity, and regulate surface temperatures. In this way, Cenozoic mountain building and surface uplift played an important role in the long-term evolution of both regional and global climate change. Here we present an introduction to how mountains and climate interact in order to better understand the significance of the development of the South American Andes and the North American Cordillera and associated uplift-induced climate change.

2. Mountain and Climate Interactions

Climate and land surfaces are intimately linked. In particular, the high topography of mountainous regions creates unique ecosystems that drive evolution and biodiversity, supply

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nutrients and sediments through weathering and erosion, store water as snow and ice, and dictate local and downstream weather and storms. Mountains typically have steep slopes and strong elevation gradients (e.g. biodiversity, temperature, precipitation) that create unique local and regional climate variability. Here we investigate how temperature, atmospheric circulation, and precipitation vary in mountainous regions.

2.1. Temperature

Mountainous regions exhibit some of the largest elevational temperature gradients on Earth. In the Colorado Rockies, average annual temperature varies more than 10 °C with elevation on the windward side of the range and up to 15 °C on the leeward side. In South America, average annual temperatures vary more than 20 °C up the Eastern Andean Cordillera and 5 °C along the Western Andean Cordillera.

The rate at which temperature of an air parcel decreases with height is the lapse rate. To first order, the lapse rate of a parcel is determined by adiabatic expansion and compression. As a parcel of air ascends, the surrounding atmospheric pressure decreases and the parcel expands adiabatically. Through this process, the work done to expand the parcel consumes internal energy, causing the air parcel to cool. Conversely, as a parcel descends, temperature increases by compressional warming. For an ascending or descending unsaturated (dry) particle, the average dry adiabatic lapse rate is 9.8 °C/km.

The lapse rate for a rising parcel over a high mountain is typically lower than the dry adiabatic lapse rate due to warming through latent heat release as vapor condenses during the precipitation of clouds, rain, hail and snow. The lapse rate for a saturated air parcel, the moist adiabatic lapse rate, is 6.5 °C/km. In reality the adiabatic lapse rate of air parcels across most mountain ranges is a continuum between the dry and moist adiabatic lapse rates, reflecting the initial relative humidity of the air parcel and the amount of cooling that it undergoes during ascent. Latent heating of a parcel may cause it to be more buoyant than surrounding cooler air, leading to instability and upward motion known as moist convection. For example, the presence of moisture and propensity for moist convection accounts for low temperature lapse rates in tropical regions in comparison to those in extratropical regions.

The dry and moist adiabatic end members belie the complexity of lapse rates in mountainous regions. Over many ranges, non-adiabatic processes cause lapse rates to deviate from these end members. For example, temperature lapse rates generally follow diurnal and seasonal heating patterns with higher values in the nighttime and winter due to stable stratification of the boundary layer and lower values in the daytime and summer due to convective mixing and the breakdown of stable stratification (Pepin and Losleben, 2002; Kattel et al., 2013). Air parcels are also heated and cooled through thermal emission and mixing and can vary both spatially and temporally. On the Bolivian Altiplano, seasonal temperature lapse rates range from 4.0 to 6.5 °C/km (Gonfiantini et al., 2001) while in the central Himalaya temperature lapse rates have an annual mean of 5.4 °C/km but vary seasonally from 4.3 to 6.1 °C/km (Kattel et al., 2013). In the Colorado Rockies, temperature lapse rates vary spatially from 1.0 to 6.3 °C/km with greater lapse rates at higher elevations (Pepin and Losleben, 2002). In this region, steepening of the temperature lapse rate may be explained by increased snow cover, enhanced airflow over the mountains, and changes in solar radiation (Pepin and Losleben, 2002). In the Washington Cascades, seasonal temperature lapse rates vary more on the leeward (eastern) side of the mountains than on the windward (western) side. In the lee of the Cascades lapse rates vary seasonally by 4 °C/km with the smallest values (2 °C/km) December-January and largest values (6 °C/km) May- July (Minder et al., 2010). On the windward side of the mountains, annual lapse rate variation is 1 °C/km (4-5 °C/km) (Minder et al., 2010). Greater variability of temperature lapse rates on in the lee of the Cascades is likely due to pooling and damming of cold air along the eastern flank (Minder et al., 2010).

Taken together, temperature lapse rates depend on parcel humidity, temperature, and mixing and vary with topography, climatology, latitude, season, relative humidity, atmospheric circulation, cloudiness, orientation (aspect), and vegetation (Rennick, 1977; Stone and Carlson, 1979; Laughlin, 1982). In mountainous regions, orographic orientation (aspect) affects the amount of incoming solar radiation and temperature. Southern and western facing slopes receive more solar radiation and typically have warmer temperatures than shaded northern and eastern facing slopes. Near surface stability and heating can also affect temperature lapse rates by influencing cloud cover. For example, under a stable cloudy sky, a reduction of daytime insolation and confinement of nighttime outgoing long-wave radiation may cause moister conditions and variations in temperature lapse rates (Stone and Carlson, 1979; Jeffery et al., 2012; Kattel et al., 2013). In mountainous regions, these lapse rates yield steep temperature gradients that drive pressure differences and atmospheric circulation and dictate both local and regional climate.

2.2. Atmospheric Circulation

Major mountain ranges redirect how and where air flows across the planet. On a planetary scale, orography induces large-scale waves that disrupt zonal (west-east) wind patterns and cause meridional (north-south) flow (Smith, 1979). Major mountain ranges also affect regional-scale circulations and produce distinct climate regimes such as the South Asian Monsoon or the interior prairie grasslands of North America (Ruddiman and Kutzbach, 1989; Broccoli and Manabe, 1992; Kutzbach et al., 1993).

Topography drives large-scale circulation patterns due to the mechanical blocking effects of mountain barriers (Galewsky, 2008). When winds encounter large landmasses, they flow up and over or deflect around the obstacle (Etling, 1989; Whiteman, 2000). In part, the path winds take depends on the kinetic energy of the system. If the flow has sufficient kinetic energy to rise against the force of gravity, it will do so (Galewsky, 2008; Galewsky, 2009). If not, the flow is deflected around the barrier (Whiteman, 2000).

Additional characteristics that determine whether airflow moves up and over or around an orogen include (1) the length, width, and height of the orographic barrier, (2) the orientation and spacing of the mountain range, and (3) the speed and stability of the approaching airflow, where stability is a measure of the potential for vertical motion (Justus, 1985; Whiteman, 2000). In the atmosphere, stable flow contains little vertical motion and is unlikely to rise while unstable flow can more easily rise up and over a barrier. Additionally, if a mountain range is lined with valleys or channels, winds may funnel through those gaps. For example, on the eastern Altiplano, the channeling of moist flow up river valleys causes cloud formation and precipitation near river headwaters (Giovannettone and Barros, 2008).

Winds tend to flow over mountain ranges when the barrier is long, the cross-barrier wind is strong, and approaching flow is unstable or near neutral (Whiteman, 2000). Alternatively, winds tend to be deflected around mountain ranges when barriers are narrow or high, convex on the windward side, cross-barrier winds are weak, and approaching flow is stable (Whiteman, 2000). In mountainous regions, deflected winds can also form a barrier jet, a jet-like wind current that forms when stable air approaches a barrier that flows parallel to the mountain range (Parrish, 1982). In the northern (southern) hemisphere, barrier jets turn towards the north (south). For example, the South American Low Level Jet (SALLJ), one of the dominant features of the continent's climatology, flows southward parallel to the Andes and is responsible for transporting considerable moisture from the Amazon to the La Plata Basin in southern South America (Vera et al., 2006). Moreover, the movement of winds over or around mountains is of distinct importance in understanding climate as these barriers locate regions of precipitation and aridity and regulate surface temperatures.

When air flows up and over a mountain, buoyancy and gravitational forces act upon vertically displaced air particles and downstream waves develop (Smith, 1979; Durran, 1990). These waves are known as mountain gravity waves, or lee waves, and produce distinct climate patterns (Durran, 2003).

Mountain waves propagate away from the orography that caused them and extend vertically and horizontally through the troposphere (Smith, 1979). Formation of these waves depends on (1) the speed, vertical temperature gradient, and stability of the flow as it approaches a mountain barrier; (2) the size and shape of the orographic barrier; and (3) the orientation of flow relative to that barrier (Smith, 1979; Durran, 2003). The largest mountain waves form in the lee of an orographic barrier when approaching flow is perpendicular to the barrier and that barrier is both steep and high (Whiteman, 2000; Durran, 2003). Weak approaching winds form only shallow waves downwind of an orographic barrier while moderate strength winds frequently overturn in the lee of mountains and form standing, or non-propagating, downstream eddies (Smith, 1979; Durran, 1990; Whiteman, 2000).

Orographically induced stationary waves, planetary-scale waves that remain nearly fixed in relation to the earth surface, have strong effects on regional climate through meridional surface winds (Smith, 1979). These waves advect temperature and moisture, direct preferred paths of storm tracks, and contribute to zonal circulation (Smith, 1979; Nigam and DeWeaver, 2003). For example, the wintertime jet stream in the western United States develops a pronounced trough and ridge pattern as it moves over the Rocky Mountains. Winds in this wave pattern advect northerly cold air over western North American, transport warm air from the Gulf of Mexico toward eastern North America, and steer storm tracks across the continent.

Although this discussion has focused exclusively on planetary-scale circulation, it is also important to consider small-scale orographically induced features that contribute to atmospheric circulation as well. Many of these characteristics are due to elevation differences and temperature contrasts that induce differential along-slope heating and cooling and vertical motion (Barry, 2013). The most common example, discussed briefly here, is valley circulation.

Mountain valley winds are dictated by an antitriptic wind component, which maintains a balance between the pressure gradient force and frictional force and is directed toward areas of low pressure, and a gravity wind component, which is directed downslope (Barry, 2013). Valley winds are also driven by a strong temperature control and display distinct diurnal and seasonal variation. Nighttime radiative cooling causes local down-valley movement while daytime warming causes up-valley circulation (Barry, 2008; Barry, 2013). A similar seasonal pattern describes general winter and summer valley circulation as well.

2.3. Precipitation and Aridity

One of the primary observable effects of orographically modified circulation are continental patterns of precipitation and aridity. Air undergoing forced ascent over an orographic barrier cools adiabatically, leading to condensation of clouds and precipitation along the mountain front (Roe, 2005; Hughes et al., 2009). As winds descend the leeward side of mountains, the air undergoes compression, warms, and experiences a reduction in relative humidity, leading to undersaturated conditions that produce little precipitation. This phenomenon, known as the rain shadow effect, demonstrates the potential for an orographic range to produce two distinct climates on the windward and leeward sides of a range.

The rain shadow effect describes climates where prevailing winds flow perpendicular to orographic barriers such as in the North American Cordillera. In the Rockies, moist cool climates (Pacific coast) dominate the windward flank of the region while interior leeward climates (Great Plains) are warmer and drier (Broccoli and Manabe, 1992). However, as is the case in the western Tibetan Plateau, when orographic barriers lie parallel to prevailing winds more nuanced mechanisms are needed to explain patterns of precipitation and climate (Broccoli and Manabe, 1992).

In the absence of orography, precipitation amounts decrease gradually with distance from moisture sources (Broccoli and Manabe, 1992). With orography, the magnitude of relief has a direct effect on the amount of precipitation and the development of climate features (Roe, 2005). Accordingly, higher mountain ranges produce stronger windward precipitation and more intense interior aridity than lower ranges. Additionally, due to rainout, condensation decreases exponentially with height such that regions with maximum precipitation are typically found at low elevation on the windward flank (Roe, 2005). Orography in tropical regions can also act as a dynamical forcing that destabilizes the atmosphere and produces convective rainfall. For example, both rainout and convective rainfall are common phenomenon in the central Andes where westward-flowing moist air from the Amazon region ascends the eastern flanks of the Andes (Garreaud et al., 2003; Insel et al., 2010, Barnes et al., 2012). The frequency and intensity of orographically induced convective rainfall varies on diurnal and seasonal timescales as a function of insolation, and is greatest during the afternoon and early evening and in summertime (Garreaud et al., 2003).

In addition to large-scale orographic phenomena, small-scale features such as mountain gaps affect climate as well. For example, the Columbia River Gorge provides a channel along which moist air flows across south central Idaho and onto the Yellowstone plateau (Stewart et al., 2002). As a result, regions such as Yellowstone National Park, which lies on the eastern edge of the Snake River Plain, have relatively high annual precipitation east of the Rocky Mountains. Similarly, valleys in the Andes provide a conduit for moisture to move among mountains and into regions one might otherwise assume to be dry (Bendix et al., 2006; Giovannettone and Barros, 2008). As expected, convective instability and precipitation in these regions have strong diurnal patterns (Bendix et al., 2006).

In mountainous environments, contrasts in elevation induce precipitation and aridity, regulate temperatures, and direct circulation patterns to develop and drive climate regimes. Next we examine a brief history of Cenozoic surface uplift and paleoaltimetry in the Americas as case studies of how orography contributed to past climate change.

3. Paleoaltimetry Techniques

Cenozoic climate change is defined by a transition from ice-free poles and warm conditions to ice-covered poles and cold conditions. While a reduction in greenhouse gases, carbon dioxide in particular, likely explains much of the Cenozoic global cooling, a complete mechanistic understanding of Cenozoic climate change, particularly regional changes, must consider additional climate forcings during this period. Here we outline how progressive surface uplift of the South American Andes and North American Cordillera affected climate on regional and possibly global scales.

3.1. Paleoaltimetry Proxies

Elevations of continental surfaces are among the most uncertain features of past geologic times. In the absence of direct measurements, proxies are used to infer past surface elevations. Several different paleoaltimetry proxies exist including the stable isotopic compositions of authigenic minerals and volcanic glass, the clumped isotopic composition of authigenic minerals, and paleo-fossil leaf traits, all of which preserve a signal of a physical characteristic (e.g. temperature, meteoric $\delta^{18}O$, meteoric δD, moist enthalpy) that varies as a known function of elevation. For example, the stable isotopic compositions ($\delta^{18}O$ and δD) of precipitation (meteoric waters) decrease upslope as the heavy isotopologue is preferentially lost through rainout. Most of these proxies were initially founded on the assumption that lapse rates were constant through time and equivalent to modern values (Koch, 1998; Poage and Chamberlain, 2001; Garzione et al., 2006). This assumption has recently been challenged and found to lead to potentially large biases on the order of kilometers in paleoaltimetry estimates (Rowley and Garzione, 2007; Ehlers and Poulsen, 2009; Poulsen et al., 2010; Insel et al., 2012; Feng et al., 2013, 2016; Fiorella et al., 2015). In fact, lapse rates of meteoric $\delta^{18}O$ vary with topography and climate (as described in Section 2.1) and almost certainly have not remained constant throughout the Cenozoic (Stone and Carlson, 1979; Poulsen et al., 2010; Poulsen and Jeffrey, 2011; Insel et al., 2012; Feng et al., 2013). To this end, as orogens evolve lapse rates may be affected by temperature, atmospheric CO_2 concentrations, precipitation type, moisture source, and air mass mixing, among others (Ehlers and Poulsen, 2009; Poulsen et al., 2010; Feng et al. 2013, 2016).

3.2. Climate Modeling

Climate modeling has proven to be a valuable complement to proxy studies for evaluating both past elevations and the climate response to surface uplift. Terrestrial proxy data preserve a signal that reflects some combination of elevation and climate, but as demonstrated by the lapse rate complexities discussed in Section 3.2, deconvoluting these signals can be challenging. Three-dimensional general circulation models (GCMs) can be used to directly investigate the climate response to orography and surface uplift and can provide direction for interpreting proxy signals (e.g. Ehlers and Poulsen, 2009; Poulsen et al., 2010).

Given that past continental surface elevations are not known with confidence, most GCM studies have taken a systematic approach to modifying topography in which surface elevations are progressively adjusted from modern values (e.g. Ruddiman and Kutzbach, 1989; Ehlers and Poulsen, 2009). This approach is somewhat arbitrary and need not be followed, particularly when presuppositions about past surface elevations exist (e.g. Feng et al., 2013).

GCMs have been used to explore the effects of mountains and mountain uplift on climate for nearly three decades (e.g. Ruddiman et al., 1989; Kutzbach et al., 1989; Ruddiman and Kutzbach, 1989). Early studies using atmosphere-only GCMs investigated uplift of major plateaus in western North America (Colorado Plateau, Basin and Range, and Rocky and Sierra Mountains) and south Asia (Tibetan Plateau and Himalaya Mountains). In addition to the diversion of the planetary wave pattern in the extratropical region of the northern hemisphere, these studies emphasized uplift-related regional changes in seasonal surface winds and precipitation due to enhanced seasonal cooling and heating during summer and winter.

Since this early seminal work, both the performance and capabilities of GCMs have improved substantially. Though a summary of these advances is beyond the scope of this review, increases in GCM spatial resolution, improvements in simulating the physics of convection and cloud formation, and the incorporation of water isotopes are worth mention for their benefits to the study of surface uplift and climate change. The first of these advances has led to significant improvements in the simulation of convection, clouds, and precipitation in the vicinity of high topography. Moreover, the reduction in spatial scale and improved resolution of topographic features in GCMs has allowed for the simulation of climate features nearer to the scale represented by the proxies. The second advance has allowed for straightforward comparisons between simulated meteoric stable isotope tracers and stable isotope proxies.

4. Surface Uplift and Climate Change

4.1. South American Andes

The Andes are the second largest topographic feature in the world and the only major orographic barrier in the southern hemisphere. The Andean Cordillera is 7000 km long and 100-300 km wide with a mean height over 4000 m and many 5000-6000 m peaks (Barry, 2008). The Andean Plateau, or Altiplano, which lies mainly in the Bolivian and Peruvian Andes, is approximately 400 km wide and has a mean elevation of almost 4000 m (Barry, 2008). Although the mechanism and timing of Andean surface uplift are uncertain, most agree that the Andes grew as a result of convergence of the Nazca and South American Plates during the Cenozoic (Gregory-Wodzicki, 2000). The Altiplano achieved its modern east-west width by 15-25 Ma, although its elevation at that time is unknown (Allmendinger et al., 1997; Isaks, 1988; Gregory-Wodzicki, 2000).

The two dominant, and contrasting, theories of Andean surface uplift are rapid recent plateau rise and slow and steady plateau rise (Barnes and Ehlers, 2009). Geologic evidence for rapid surface uplift has been proposed from paleobotany, soil carbonate $\delta^{18}O$, volcanic glass δD, and clumped isotope paleothermometry proxies and has been interpreted to indicate pulses of Andean surface uplift of 2.5 ± 1 km from 19 to 16 Ma and ~10 to 6 Ma (Gregory-Wodzicki, 2000; Ghosh et al., 2006; Garzione et al., 2006; Saylor and Horton, 2014). However, these rapid uplift estimates may be biased by neglecting climate change associated with surface uplift (Ehlers and Poulsen, 2009; Poulsen et al., 2010; Insel et al., 2012) and/or by neglecting evaporative enrichment of stable isotope records (Fiorella et al., 2015). Accounting for these biases largely removes the evidence for rapid uplift and lends support to models that favor a history of slow and steady Andean surface uplift, which has been protracted over ~40 Ma (e.g. Barnes and Ehlers, 2009; Insel et al., 2012).

Regardless of the exact timing of uplift, the rise of the Andes undoubtedly had a substantial impact on the regional climate and environment. Modern Andean climate is governed by blocked zonal flows that drive regional circulation and precipitation (Insel et al., 2010). Mechanical blocking deflects low-level Atlantic trade winds into a northerly barrier jet, the South American Low Level Jet (SALLJ), that flows along the eastern flank of the Andes (Campetella and Vera, 2002; Vera et al., 2006). Moisture transport across the Amazon Basin, development of the LLJ, and orographic lifting of easterly winds combine to yield high precipitation in this region (Lenters and Cook, 1995, Insel et al., 2010).

As the Andes are uplifted, paleoclimate models indicate that Altiplano temperature decreases, precipitation increases along the eastern flank, and zonal winds are blocked (Fig. 1) (Ehlers and Poulsen, 2009).

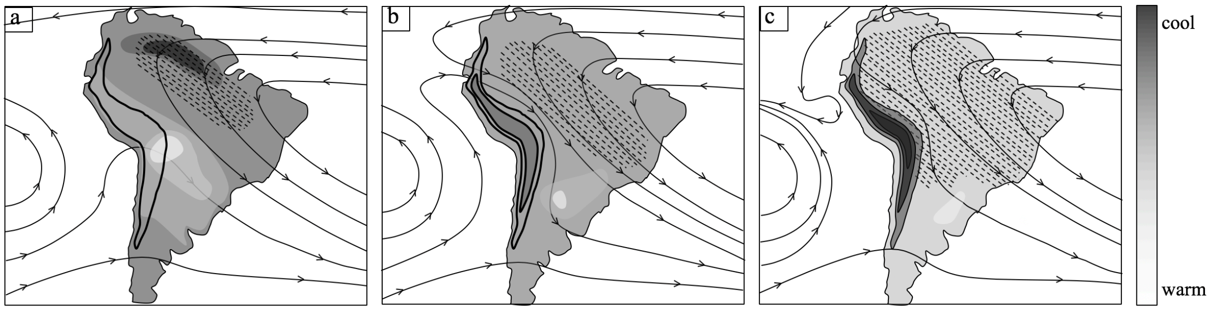

Figure 1. Schematic of South American climatology with low (A), medium (B), and high (C) surface uplift. Thin vectors represent surface winds. Diagonal hatching indicates regions with mean annual precipitation greater than 150 cm/yr. Gray shading represents temperature where dark gray indicates cooler temperatures and white indicates warmer temperatures.

As elevation increases from 200 to 4800 m, temperature on the Altiplano and other high elevation areas along the Andean Cordillera decreases more than 6 °C due to non-adiabatic cooling and increased latent and sensible heat loss (Ehlers and Poulsen, 2009).

Andean growth also leads to a shift in the source and direction of prevailing winds. Before the Andes uplifted, prevailing winds are westerly from the Pacific. As the Andes rise, westerly winds are blocked and zonal winds switch to an easterly Atlantic source (Ehlers and Poulsen, 2009, Poulsen et al., 2010). Additionally, with uplift mechanical blocking leads to the development of a low-level jet (SALLJ) along the eastern flank of the orogen (Insel et al., 2010).

Development of the SALLJ is integral to the evolution of uplift-induced precipitation patterns in the eastern Andes. Without the Andes, paleoclimate models indicate that the SALLJ is absent (Insel et al., 2010). Accordingly, low-level transport of moisture fades with distance from its source, convective processes and cumulus cloud formation are suppressed, and vertical velocities are reduced due to the lack of a lifting mechanism (Insel et al., 2010). Additionally, due to its subtropical location, the Altiplano region is hyperarid without SALLJ moisture transport (Fiorella et al., 2015). At 50-75% of modern height, the SALLJ is abruptly initiated, westerly flow is at least partially blocked, and moisture transport, humidity, and precipitation increase along the eastern flank and across the Altiplano (Ehlers and Poulsen, 2009; Insel et al., 2010; Fiorella et al., 2015). As Andean elevations increase to their modern height and the SALLJ strengthens, the eastern flank transitions from arid to humid and net evaporation decreases across the Altiplano (Ehlers and Poulsen, 2009; Insel et al., 2010; Fiorella et al., 2015). Under modern conditions, moisture transport from the Amazon Basin and resulting eastern flank precipitation are so high that the South American Monsoon develops along the eastern Bolivian Andes (Garreaud, 2003).

Upper level (200 mb) flow is dominated by the Bolivian High, an upper level anticyclone that develops during summer months. The Bolivian High affects low-level near-surface circulation (Garreaud, 1999; Lenters and Cook, 1999) but is not directly affected by mechanical forcing of Andean elevation (Insel et al., 2010). In its modern state, easterly upper level flow enhances precipitation over the Andes while westerly flow causes dry conditions (Insel et al., 2010). When the Andes are absent, the reduction of low-level moisture transport and eastward shift in the region of maximum latent heating and convergence causes the Bolivian High to shift eastward (Insel et al., 2010).

Importantly, the climate changes associated with surface uplift are non-linear (Ehlers and Poulsen, 2009; Garzione et al., 2014). For example, when the Andes are 0% and 25% of their modern height, winds are still predominantly southwesterly from the Pacific. As the Andes achieve 50% and 75% of their modern height, winds shift to flow from the northwest but still move over the orogen. However, at their modern height, prevailing winds flow from the east across the Amazon basin, are deflected southward due to blocking, and join with the south Atlantic convergence zone (Lenters and Cook, 1997; Ehlers and Poulsen, 2009). This stepwise change highlights the potential for non-linear uplift induced climate change and the importance of uplift/climate thresholds (Insel et al., 2012).

It is also important to note that the climatic effects of Andean uplift extended far beyond the Cordillera and its flanks. As Andean elevation increases, moisture transport along the SALLJ decreases Amazonian precipitation in the eastern basin (Insel et al., 2010), but increases precipitation in the central Amazon basin due to orographic blocking and rainout (Ehlers and Poulsen, 2009). Additionally, Andean uplift contributed to the development of new ecosystems, redirected erosion and river drainages, and led to the proliferation of new flora and fauna biodiversity in the Amazon basin (Hoorn et al., 2010). Along the west coast of South America, Andean uplift initiated shifts in SST gradient and circulation, which contributed to increased marine productivity and decreased the intensity and frequency of El Niño-Southern Oscillation (ENSO) and El Niño events (Feng and Poulsen, 2014).

Taken together, uplift of the Andes contributed to a wide array of regional climate, landscape, and biological effects along the Cordillera and elsewhere. A dual approach of model reconstructions and proxy records of surface uplift in the Andes Mountains explains up to 6 °C cooling during the Cenozoic, a general trend of increased precipitation along the eastern flank and Altiplano, and shift of prevailing winds from a westerly Pacific source to an easterly Atlantic one (Ehlers and Poulsen, 2009). Although the timing is debated, Andean surface uplift certainly had a first order impact on regional climate.

4.2. North American Cordillera

The North American Cordillera is a 6000 m long mountainous spine that passes through most of western North America. The modern North American Cordillera stretches from Alaska to Mexico and encompasses the Rockies, Cascade Range, Sierra Nevada Mountains, Columbia Plateau, and the Basin and Range province. This topography extends over 1000 km wide and has a first order control on regional climate.

As with the Andes, the mechanisms and timing of surface uplift along the North American Cordillera are still debated, although most agree that orogenic growth was largely initiated by subduction of the Farallon Plate below the North American Plate. Evolution of the North American Cordillera occurred in three distinct tectonic phases: Sevier orogeny (140 Ma to 50 Ma), a thin-skinned compressional deformation; Laramide orogeny (70 to 80 Ma through 35 to 55 Ma), a thick-skinned compressional deformation; and Basin and Range extension (17 Ma to present), which resulted in a reduction of mean elevation.

The two dominant and contrasting mechanistic hypotheses of North American Cordillera evolution are Cretaceous/early Paleogene gravitational plateau collapse and Cenozoic uplift (DeCelles, 2004). According to the Cretaceous gravitational collapse hypothesis, subduction of the Farallon Plate created a highland that later experienced gravitational strain and spread laterally through the Cenozoic (Hodges and Walker, 1992; DeCelles, 2004). Alternatively, the Cenozoic uplift hypothesis proposes that the North American Cordillera formed from early Cenozoic rapid north to south migration of high elevation topography (Mix et al., 2011; Chamberlain et al., 2012).

Further confounding the geologic and tectonic history of the North American Cordillera, Lechler et al. (2013) note that the complex topography in this region likely reflects a combination of gravitational plateau collapse, uplift, and extension processes. Additionally, reconstructions of North American paleoaltimetry are complicated because different methods yield kilometer-scale differences in elevation estimates. Stable isotope based paleoaltimetry estimates suggest that the North American Cordillera plateau was 3-4 km during the Eocene (Mix et al., 2011; Chamberlain et al., 2012, Feng et al., 2013) while fossil-leaf based paleoaltimetry estimates indicate peak elevations 2-3 km, depending on whether temperature and/or moist enthalpy were considered (Chase et al., 1998; Wolfe et al., 1998). Despite these incongruities, recent proxy-based paleoaltimetry reconstructions and climate modeling tend to favor the Cenozoic uplift theory (Mix et al., 2011; Chamberlain et al., 2012, Feng et al., 2013; Sewall and Fricke, 2013) and development of a late Eocene plateau (Fan et al., 2014a).

North American Cordillera paleoclimate models indicate cooler temperatures with surface uplift (Ruddiman and Kutzbach, 1989; Feng et al., 2013, Sewall and Fricke, 2013). With no topography, temperatures have a largely meridional gradient with warmer temperatures in the south and cooler temperatures in the north (Sewall and Fricke, 2013). With uplift, this gradient weakens and high elevation areas cool non-adiabatically through advective heat loss (Feng et al., 2013; Sewall and Fricke, 2013). Additionally, western flank temperatures cool due to the formation of large amplitude stationary mountain gravity waves that advect cold air from the north (Kutzbach et al., 1989; Ruddiman and Kutzbach, 1989).

Modern atmospheric circulation in the North American Cordillera is predominately driven by orographically directed flows and competition between air masses from the Pacific Ocean and the Gulf of Mexico. However, without high surface topography climate models indicate that westerly winds flow relatively unimpeded across North America (Kutzbach et al., 1989; Feng et al., 2013). With progressive uplift, westerlies are diverted around the Rocky Mountains (Kutzbach et al., 1989) and develop a pronounced trough and ridge pattern as they are deflected over and around the orogen (Fig. 2) (Ruddiman and Kutzbach, 1989).

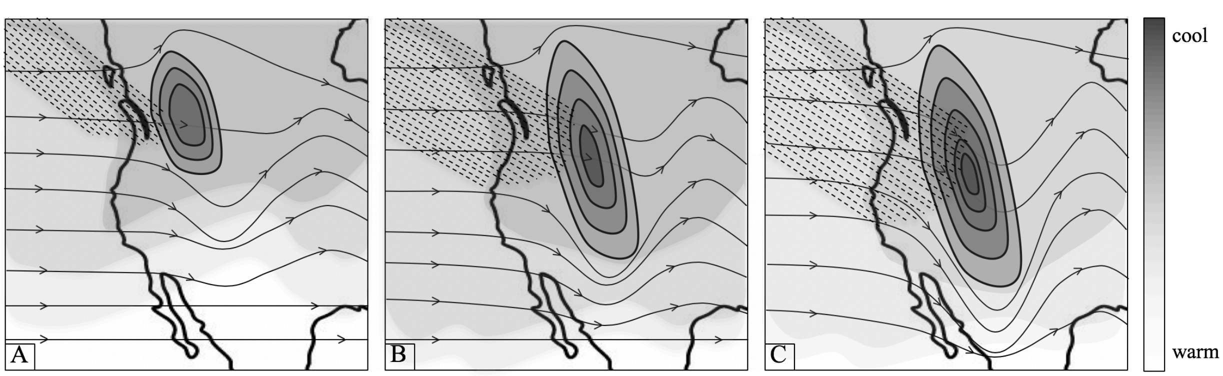

Figure 2. Schematic of North American climatology with low (A), medium (B), and high (C) surface uplift. Thin vectors represent surface winds. Diagonal hatching indicates regions with

mean annual precipitation greater than 150 cm/yr. Gray shading represents temperature where dark gray indicates cooler temperatures and white indicates warmer temperatures.

As the Cordillera evolves, deflection intensifies and a sharper trough emerges. At its modern height, southerly flow from the Gulf of Mexico transports moisture to the lee of the Colorado Rockies where a seasonal monsoon develops (Feng et al., 2013; Sewall and Fricke, 2013).

Beyond the North American Monsoon, changes in atmospheric circulation due to surface uplift lead to the development of distinct seasonal precipitation patterns. As surface uplift evolves and the amplitude of the stationary wave increases, more pronounced onshore surface flow and surface heating increases windward wintertime precipitation (Fig. 2) (Feng et al., 2013; Sewall and Fricke, 2013). At modern height, forced ascent of westerly airflow over orographic masses promotes diabatic cooling, condensation, and snowfall and causes strong winter windward precipitation (Chamberlain et al., 2012; Feng et al., 2013; Sewall and Fricke, 2013). Uplift causes the summertime leeward monsoon to intensify (Kutzbach et al., 1989) while also inducing a rain shadow that causes much of the interior continental United States to become drier (Ruddiman and Kutzbach, 1989; Broccoli and Manabe, 1992; Fan et al., 2014b).

As with the Andes, orography has a first order control on regional and local climate in North America. Modern hemisphere-scale stationary waves seasonally shift the position of the jet stream and dictate temperature patterns across the continent.

5. Conclusions

Cenozoic surface uplift has long been associated with climate change. Here we have presented a summary of uplift-induced temperature, precipitation, and atmospheric circulation changes in the Andes and North American Cordillera regions. Although the timing and mechanism of uplift in neither region is yet constrained, model results of systematic uplift generally suggest enhanced windward precipitation, orographic deflection of prevailing winds, and cooler temperatures. However, there are distinct differences in the response of uplift between these regions that are related to their geographic position. Topographic blocking at low latitudes by the Andes led to development of a low level jet, moisture transport from the tropics, and initiation of convective rainfall and cumulus cloud formation along the eastern flanks. Topographic blocking in middle latitudes by the North American Cordillera modified the stationary wave pattern creating a pronounced ridge and trough over western North America, formed a rainshadow downstream of high topography, and initiated monsoonal conditions along the eastern front of the Cordillera. These changes acted in concert with other factors to produce Cenozoic climate and environmental change.

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