

STABLE ISOTOPE PALEOALTIMETRY: A CRITICAL REVIEW

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ABSTRACT. Stable isotope paleoaltimetry makes use of systematic trends in the distribution and isotopic composition of modern precipitation with climate and topography, and of the potential to estimate the isotopic composition of paleoprecipitation from authigenic (*in-situ* formed) minerals. To illustrate the usefulness as well as potential limitations of this method, we review (1) processes controlling the isotopic composition of modern precipitation, (2) stable isotope data from modern precipitation across regions of high topography, and (3) stable isotope data from authigenic minerals that have been used to infer paleotopography. From this we conclude that stable isotope studies of authigenic minerals can permit useful inferences on paleotopography, with uncertainties that critically depend on a detailed understanding of local- to global-scale paleoclimate during the time interval of interest.

INTRODUCTION

Many recent studies of young and actively deforming orogens have shown that the effect of tectonically controlled mountain uplift on surface processes can be difficult to evaluate. Commonly thermochronologic methods (particularly fission-track and U-Th/He studies) are used to infer surface uplift, but these methods really constrain denudation histories, which may, in some cases, largely reflect climatic conditions. Similarly the timing of syn-orogenic deposition, while clearly linked to denudation and sediment transport, is not a direct constraint on the timing of surface uplift. As a result, our quantitative understanding of the interaction between tectonic deformation, mountain uplift and surface processes is limited, with the most poorly constrained parameter generally being the paleoelevation of most mountain belts. Therefore much attention has been focused on developments in the field of paleoaltimetry, including the use of basalt vesicularity (Sahagian and Maus, 1994; Sahagian and others, 2002a, 2002b), cosmogenic nuclides (Brook and others, 1995; Libarkin and others, 2002), paleobotany (Wolfe, 1993; Forest and others, 1995, 1999; Gregory-Wodzicki, 1997, 2002), and stable isotope studies of authigenic (*in-situ* formed) minerals both within and on the leeward side of orogens (see below for references). Of these methods, the one arguably best suited for quantitative analysis of the regional-scale effects of mountain uplift may be the study of the oxygen and/or hydrogen isotope composition of authigenic minerals. This approach makes use of systematic trends in the isotopic composition of precipitation with climate and topography, and of the potential to estimate the isotopic composition of paleoprecipitation from authigenic minerals. As discussed in more detail below, the main effect of topography on precipitation is progressive condensation and removal of precipitation from a moist air mass moving over high terrain. Because condensation partitions more of the “heavy” isotopes of oxygen and hydrogen (¹⁸O and ²H) into the condensed phase, an air mass not only retains lower amounts of water after passing over a mountain belt, but its water vapor is also highly depleted in these isotopes. Correspondingly mountain uplift can, in principle, be quantitatively reconstructed through stable isotope studies of authigenic minerals from dated syn-orogenic sediment sequences within intramontane basins, or on the leeward side of mountains. In this paper, we review stable isotope studies of (1) modern precipitation and surface waters across mountain belts and other regions of high surface elevations, and (2) authigenic minerals from syn-orogenic sediment

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sequences in regions characterized by Cenozoic tectonism to illustrate the usefulness as well as potential limitations of stable isotope paleoaltimetry.

THE ISOTOPIC COMPOSITION OF PRECIPITATION AND SURFACE WATERS

In this section we briefly review (1) global- to regional-scale atmospheric circulation patterns, (2) the main hydrological processes controlling the isotopic composition of precipitation, and (3) how the interaction of atmospheric circulation patterns with high topography affects both the spatial distribution as well as the isotopic composition of precipitation.

Atmospheric Circulation Patterns

The spatial distribution of precipitation is primarily controlled by global-scale atmospheric circulation patterns, which are driven by net insolation heating at equatorial latitudes and net radiative cooling in polar regions. Combined with the action of the Coriolis force, this leads to a pronounced latitude zonation with three major circulation cells (for example, Grotjahn, 2002): (1) Hadley cells in tropical regions from at or near the equator (where air heats, gains large quantities of evaporated water vapor, and rises at the Intertropical Convergence Zone (ITCZ)) to $\sim 30^\circ$ latitude; these are regions of predominantly equator- and westward moving surface winds (the easterly “trade winds” or “tropical easterlies”). (2) Polar cells from $\sim 60^\circ$ latitude to 90° latitude (where air cools and sinks); they are, similar to the Hadley cells, regions of predominantly equator- and westward moving surface winds (the “polar easterlies”). (3) Ferrel cells from $\sim 30^\circ$ latitude to $\sim 60^\circ$ latitude; Ferrel cells are not thermally driven but form to balance the transport by Hadley and Polar cells, and are regions of predominantly pole- and eastward moving surface winds (the “prevailing westerlies”). Near the boundaries between these cells, where the horizontal component of near-surface airflow is relatively small, winds are generally light and highly variable in direction, especially near the equator because the Coriolis force is negligible in that region. At regional to local scales, atmospheric circulation patterns are, of course, more complex. For example, the boundaries between these cells are fairly irregular and vary seasonally due to seasonal changes in insolation (for example, Dima and Wallace, 2003). Additionally, there are variations in atmospheric circulation patterns associated with variations in ocean circulation and sea surface temperatures, such as the El Niño—Southern Oscillation (ENSO) phenomenon. During ENSO events, unusually strong increases in surface air pressures over the western Pacific and decreases over the eastern Pacific lead to a pronounced weakening of the easterly trade winds, or even to westerly winds, over the tropical Pacific Ocean; this, in turn, significantly affects atmospheric circulation in adjacent regions, and even global circulation patterns (Bjerknes, 1969; Neelin and others, 1998). There also are seasonal reversals in temperature gradients between continental regions and adjacent oceans that cause strong seasonal variations of surface wind directions and precipitation rates that are referred to as monsoons (for reviews see Ramage, 1971; Hastenrath, 1991; Meehl, 1992). Monsoonal circulations influence climate both globally (Trenberth and others, 2000) and at regional scales, for example over large parts of Asia, Australia, and East Africa (Hahn and Manabe, 1975; Clemens and others, 1991; Broccoli and Manabe, 1992; Meehl and Arblaster, 1998; Webster and others, 1998; Yang and Lau, 1998; Tian and others, 2001), West Africa (Janicot and others, 1998), southwestern North America (Adams and Comrie, 1997; Barlow and others, 1998; Higgins and others, 1998), and northern South America (Zhou and Lau, 1998). Furthermore, while the predominant wind directions may be dictated by global- to regional-scale patterns, local wind directions and weather patterns are highly variable in response to irregular small-scale temporal and spatial variations in heating of the earth’s surface.

Notwithstanding these complexities, the interaction of near-surface wind patterns with topography exerts a strong systematic control on the spatial and temporal distribution as well as the stable isotope composition of precipitation, as discussed below.

Processes Controlling the Isotopic Composition of Precipitation

The oxygen and hydrogen isotope ratios of precipitation, and consequently surface waters derived from precipitation, reflect processes of the hydrologic cycle (see Gat, 1996, for a review) and are mainly controlled by: (1) kinetic and equilibrium isotope fractionation associated with the evaporation of water vapor from its source, generating atmospheric moisture depleted in ^{18}O and D (^2H) relative to that source; (2) cooling of a moist air mass by uplift, mixing with cooler air, or radiative energy loss; below the dew point temperature this results in the condensation of liquid (or solid) water, which partitions more of the heavy isotope into the condensed phase; (3) isotope exchange of liquid water drops with water vapor within clouds; (4) removal of the condensed phase from the air mass by precipitation, leaving the water remaining in the air mass depleted in the heavy isotope; and (5) below cloud-base evaporation and/or isotope exchange with water vapor. Isotope ratios are reported in δ -notation, with δD and $\delta^{18}\text{O}$ defined as $1000 \cdot (R_{\text{sample}}/R_{\text{standard}} - 1)$, where R is D/H, or $^{18}\text{O}/^{16}\text{O}$; isotope ratios are generally expressed relative to SMOW for hydrogen and oxygen in waters and non-carbonate minerals, and relative to PDB for oxygen in carbonate minerals (Coplen, 1994, 1995); in many paleoaltimetry studies, however, oxygen isotope values of carbonates are also expressed relative to SMOW.

The initial isotopic composition of water vapor in an air mass is controlled by the source water isotopic composition, and the conditions under which evaporation from the water source occurs. Changes in the source water isotopic composition will have a direct effect on the isotopic composition of water vapor in an air mass, and ultimately precipitation. Changes in the source water (typically sea surface) temperature and air temperature above the moisture source affect the isotopic composition of precipitation via the starting moisture content of an air mass, and due to the effect of temperature of the liquid-vapor fractionation factor. The net effect of these processes is a ca. $-0.55\text{‰}/^{\circ}\text{C}$ correlation of the $\delta^{18}\text{O}$ of precipitation and sea surface temperatures in middle and high latitudes (Jouzel and others, 1997), predominately due to the effect of source temperature on atmospheric water vapor content. Sequential removal of precipitation from an air mass makes the isotope ratios of precipitation decrease with successive condensation/precipitation events. This causes systematic trends of decreasing isotope ratios of precipitation with decreasing local temperature (the temperature effect; for $\delta^{18}\text{O}$ $\sim 0.58\text{‰}/^{\circ}\text{C}$ in the 0 to 20°C temperature range; Rozanski and others, 1993), and with increasing distance from the main water vapor source (the continental effect, ca. -0.2 to $-0.4\text{‰}/100$ km for $\delta^{18}\text{O}$; Dansgaard, 1964; Rozanski and others, 1993). However, "convective precipitation," that is, precipitation formed during convective storms, can lead to precipitation with extremely low isotopic ratios as storms evolve; this is largely due to the preferential depletion of ^{18}O and ^2H in a moist air mass during progressive rain-out along with less below cloud-base evaporation under more humid conditions. As a result, there generally is only a negligible temperature effect, but a pronounced negative correlation between the isotope ratio of precipitation and the rate or amount of precipitation (the amount effect) at tropical sites, where convective storms are common (Dansgaard, 1964). Moreover, the isotopic composition of monsoonal precipitation is likely to be dominated by variations of the amount effect as well as moisture transport and rain-out history, with lower isotope ratios associated with stronger monsoons (Johnson and Ingram, 2004; Vuille and Werner, 2005; Vuille and others, 2005).

The Effect of Topography on the Distribution and Isotopic Composition of Precipitation

The main effect of topography on precipitation is progressive condensation and removal of precipitation from an uplifting air mass. During the passage of a moist air mass over a mountain belt, decreasing atmospheric pressure leads to adiabatic expansion and cooling, which causes condensation and high amounts of “orographic precipitation” on the windward side of mountain ranges (for reviews, see Barros and Lettenmaier, 1994; Roe, 2005). The resulting decrease in the $\delta^{18}\text{O}$ value of precipitation with local elevation (the altitude effect) is largely controlled by temperature-dependent Rayleigh distillation and not strictly linear (Rowley and others, 2001), but empirical data show that it can, for practical purposes, be assumed to be essentially linear in most cases. Compilations of isotope data sets from precipitation and surface waters throughout the world’s mountain belts have demonstrated that oxygen isotope lapse rates (the rates of change in isotope ratio with altitude) vary locally between $-0.1\text{‰}/100\text{ m}$ and $-0.5\text{‰}/100\text{ m}$ (Chamberlain and Poage, 2000; Poage and Chamberlain, 2001), but average ca. $-0.28\text{‰}/100\text{ m}$ of net elevation change in many regions of the world, excluding latitudes $>70^\circ$ and elevations $>5,000\text{ m}$ (Poage and Chamberlain, 2001). After passing over a mountain belt to the leeward side, an air mass will have a lower moisture content and dew point temperature than on the windward side, and it will warm adiabatically upon descent; this commonly results in an “orographic rain shadow,” that is, an arid to semi-arid region located downwind of significant topography. Small amounts of precipitation may reach the ground on the leeward side of mountains as “spillover” or orographic condensation that is carried over the peaks by the wind; spillover is favored by: narrow mountain ranges, weak atmospheric stability, high wind speeds, and long precipitation fallout times (hence cold climates, because precipitation fallout times are longer for snow as compared to rain) (Colle, 2004; Smith and Barstad, 2004). The $\delta^{18}\text{O}$ and δD values of precipitation on the leeward side of mountains may, due to the ^{18}O and ^2H depletion of the water vapor in an air mass during its ascent over the windward side of the mountains, be relatively low (the “isotopic rain shadow” effect; for example, Siegenthaler and Oeschger, 1980) and independent from local elevation (Stern and Blisniuk, 2002), unless below cloud-base evaporation and/or mixing with other moisture-bearing air masses are important (Gat and Rietti-Shatti, 1999). The magnitude of this leeward ^{18}O or ^2H depletion should correspond to the elevation increase of the air mass during its passage over the mountain belt according to the local $\delta^{18}\text{O}$ lapse rate. However, if below cloud-base evaporation and/or mixing of air masses are important, these processes can lead to the presence of altitude effects on both windward and leeward sides of mountains. At dry sites with liquid precipitation, a negative correlation of $\delta^{18}\text{O}$ and δD values of precipitation with altitude on the leeward side of mountains which is caused by greater below cloud-base evaporation at sites with lower elevations (where raindrops will have fallen a greater distance than at higher elevation sites) is referred to as “pseudo-altitude effect” (Moser and Stichler, 1971). This effect is much less important under colder conditions where some of the precipitation descent takes place as solid rather than liquid; sublimation of ice causes only a negligible fractionation of isotopes (Friedman and others, 1991), whereas evaporation of a raindrop leaves the residual liquid enriched in the heavy isotopes. Thus the combined effects of enhanced spillover of solid precipitation and minimal isotope fractionation with sublimation of solid precipitation makes isotopic rain shadows more common in colder climates as opposed to warmer climates where pseudo-altitude effects are more important.

PRECIPITATION AND SURFACE WATER ISOTOPE TRANSECTS ACROSS REGIONS OF HIGH TOPOGRAPHY

Mountains influence the distribution of precipitation through upward deflection and cooling of moist air masses, generating orographic precipitation (for reviews, see Barros and Lettenmaier, 1994; Roe, 2005), and through enhancement of monsoonal convection systems (Hahn and Manabe, 1975; Ruddiman and Prell, 1997). Many

previous studies have shown that orographic precipitation typically exhibits a pattern of decreasing oxygen and hydrogen isotope ratios of precipitation with elevation on the windward side of an orogen (Dansgaard, 1964; see Poage and Chamberlain, 2001, for a global compilation). However, the patterns of the isotopic composition of precipitation across mountain ranges and plateaus, and the conditions controlling the presence or absence of isotopic rain shadows, have not previously been systematically evaluated. In the following paragraphs we describe general patterns in the isotopic composition of precipitation and surface waters across several regions of high topography to characterize the geographic and climatic conditions under which isotopic rain shadows occur, versus those that lead to more symmetric “altitude effects” on both sides of a mountain range.

Southern Patagonian Andes

The north-south trending southern Patagonian Andes, located immediately east of the Pacific ocean, and in the center of the southern hemisphere westerlies (Prohaska, 1976), have one of the most drastic orographic rain shadows on earth, with a more than ten-fold decrease in precipitation between the windward and leeward sides of the 3 to 4 km high mountains (Hoffman, 1975). The variations in the isotopic composition of present-day precipitation and surface waters sampled along a transect across these mountains between 47°20'S and 48°S are described in Stern and Blisniuk (2002). The transect (fig. 1, transect G) extended from ~73°20'W on the western side of the mountains, where maximum windward elevations are ~1,800 m, to ~71°10'W east of the up to ~3,700 m high mountains. Correspondingly, the data from this transect likely reflect an ~2 km orographic uplift of moisture-bearing air masses carried eastward over the mountains by the westerly winds. On the windward western side of the mountains, the main trend in the data is a steady eastward decrease in the $\delta^{18}\text{O}$ values of these waters. The correlation between sampling site elevation and $\delta^{18}\text{O}$ values is relatively poor, because most of the western samples were collected at low elevation within glacially carved valleys, but Rayleigh distillation modeling and the observed $\delta^{18}\text{O}$ values (vs. site elevation, catchment elevation, and maximum elevation of windward topography) imply isotopic lapse rates of about $-0.3\text{‰}/100\text{ m}$ (Stern and Blisniuk, 2002). This is in good agreement with both the $-0.28\text{‰}/100\text{ m}$ estimate derived from a global compilation, and an estimate of $-0.30\text{‰}/100\text{ m}$ for the subset of studies from Central and South America (Poage and Chamberlain, 2001). On the leeward eastern side of the mountains, in contrast, the $\delta^{18}\text{O}$ values of precipitation and non-evaporated surface waters are consistently low (4-6‰ lower than at the western end of the transect), and show no systematic trend with local elevation; a slight increase in $\delta^{18}\text{O}$ values at the easternmost site in the transect may reflect an influx of Atlantic-derived moisture in this region of persistent storm-track activity (Trenberth, 1991; Stern and Blisniuk, 2002). The spatial pattern in the $\delta^{18}\text{O}$ values of precipitation in the southern Patagonian Andes thus seems to be largely controlled by the progressive condensation and rain-out of moisture on their windward western side, with spillover of condensate to the leeward side, resulting in a notable isotopic rain shadow.

Central and Northern Andes

Comparing altitude transects of the isotope composition of waters from the western side of the Andes at 33°S (Moser and others, 1972) with data from the eastern flank at about the same latitude (Vogel and others, 1975), Rozanski and Araguás-Araguás (1995) found nearly identical systematic decreases in isotope composition of water with altitude on both sides of the Andes (fig. 1, transect F). However, these data are characterized by an unusual decrease of the isotopic lapse rate with increasing elevation, which is problematic because it cannot be explained by purely adiabatic processes. In this region, samples from relatively high elevations of about 1,500 to

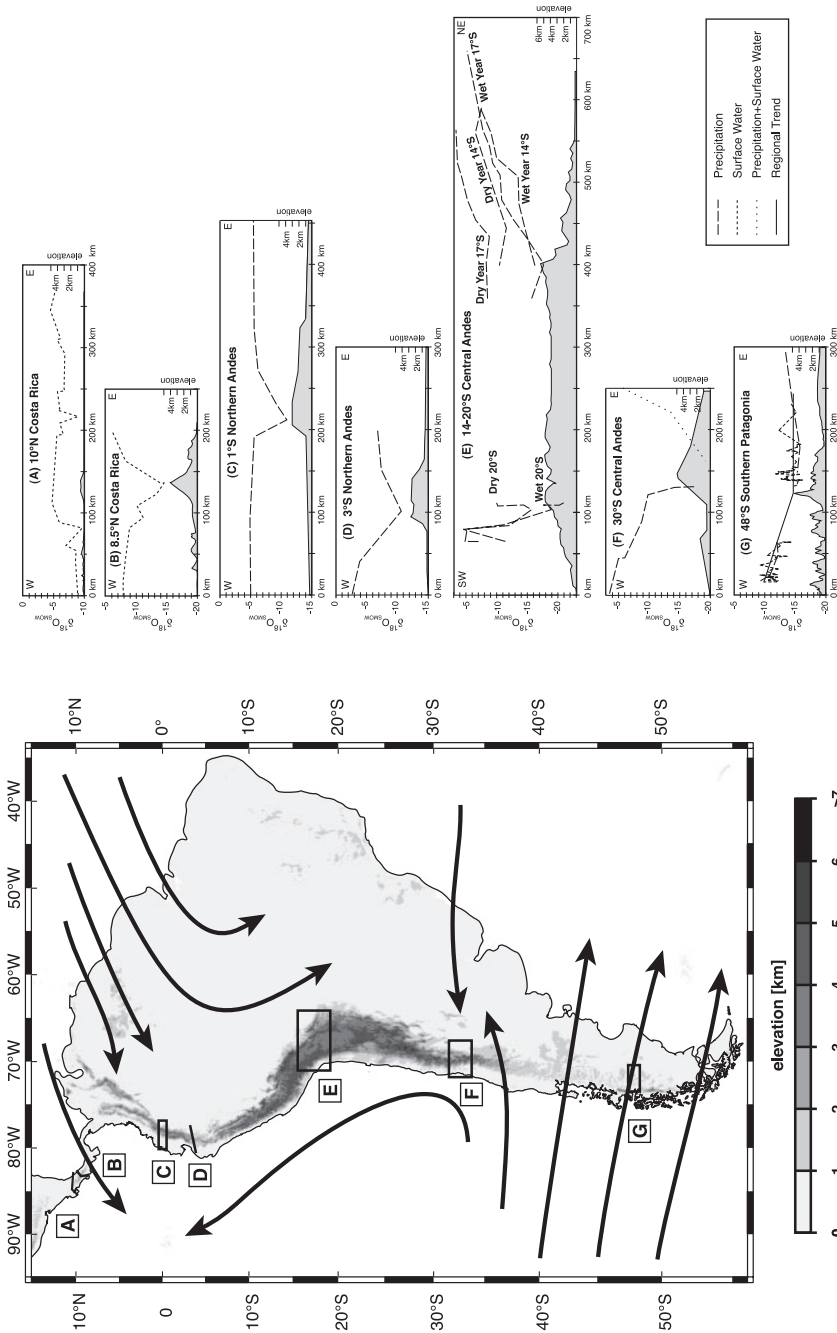


Fig. 1. Left: Topographic sketch map of the Andes and southern Central America, showing the locations of transects A–F presented to the right, and principal circulation patterns (arrows; as for all following figures, the circulation patterns are compiled from the NCEP Reanalysis Electronic Atlas, NOAA-CIRES Climate Diagnostics Center: <http://www.cdc.noaa.gov/cgi-bin/Data/tro.pl?dataset=NCEP>). Right: Oxygen isotope composition of waters across the Andes and Central American mountain ranges with schematic topographic profiles. As for figures 2–6, the lines showing the isotopic variation are averages and are only meant to show general patterns; for actual data, see the original sources. (A) and (B) after Lachnit and Patterson (2002); (C) and (D) after Garcia and others (1998); (E) after Confinantini and others (2001), Aravena and others (1999), and Fritz and others (1981); Moser and others (1972), Rozanski and Araguás-Araguás (1995), and Vogel and others (1975), including data inferred from δD data; (G) after Stern and Blisniuk (2002). Note that southern Patagonia (G) is the only site in the Andes with an isotopic rain shadow.

4,000 m yield an oxygen isotope lapse rate of about $-0.2\text{‰}/100$ m on both sides of the Andes. This is in good agreement with an oxygen isotope lapse rate estimate of $-0.21\text{‰}/100$ m for the 3,150 to 4,180 m elevation range on the western side of the Andes farther north at $\sim 24^\circ\text{S}$ (Alpers and Whittemore, 1990). In contrast, samples from about 0 to 1,500 m elevation at $\sim 33^\circ\text{S}$ yield an unusually high lapse rate of about $-0.6\text{‰}/100$ m for $\delta^{18}\text{O}$ (Rozanski and Araguás-Araguás, 1995). A possible explanation for both the anomalously high isotopic lapse rate at low elevations and the occurrence of altitude effects on each side of the Andes at $\sim 33^\circ\text{S}$ is the interference of different climatic regimes and multiple moisture sources. Atmospheric circulation in this region is seasonally variable but dominated by northeasterly surface winds containing moisture from the Amazon Basin (Garreaud, 1999, 2000), where re-precipitation of transpired water is substantial (Gat and Matsui, 1991). Thus, on the eastern side of the Andes precipitation at lower elevations seasonally may have anomalously high $\delta^{18}\text{O}$ values due to the influx of Amazonian moisture, which may have high $\delta^{18}\text{O}$ values due to lake water evaporation or transpiration, artificially maximizing the effect of altitude on the composition of mean annual precipitation. On the western side of the Andes, where surface winds are predominantly south-southwesterly due to the year-round influence of the Southeast Pacific Anticyclone (Garreaud and others, 2002), the trend in the data is presumably caused by the influx of moisture with relatively high $\delta^{18}\text{O}$ values from the Pacific Ocean at the lower-elevation western sites, as well as intense below cloud-base evaporation acting on low $\delta^{18}\text{O}$ moisture derived from the East leading to a “pseudo-altitude effect” in this hot and semi-arid region at the margin of the Atacama Desert (Aravena and others, 1999; Gat and others, 2000).

Farther north in the broad high Andean region of the Altiplano, between 20°S and 14°S , the stable isotope ratios of precipitation also decrease with altitude on both sides of the mountains (fig. 1, transect E). On the eastern side of this Andean segment surface winds are predominantly from the Northeast, but seasonally variable due to the formation of a strong high pressure system (the Bolivian High) in the upper troposphere over the Altiplano during the southern hemisphere summer (Zhou and Lau, 1998; Lenters and Cook, 1999; Vuille, 1999). On the eastern side of the Altiplano, oxygen isotope lapse rates of -0.15 to $-0.24\text{‰}/100$ m have been determined for precipitation sampled over an elevation range from ~ 200 m to $\sim 5,000$ m, with the lower rate observed during an exceptionally dry year and the higher rate observed during an exceptionally rainy year; similar differences have been observed for dry versus rainy months (Gonfiantini and others, 2001). This change in isotopic lapse rate with the precipitation rate has been interpreted to reflect the importance of the “amount effect” in this and other tropical regions (Gonfiantini and others, 2001). This is in good agreement with the results of a recent study by Vuille and Werner (2005), which demonstrated a strong negative correlation between the intensity of the South American summer monsoon and the $\delta^{18}\text{O}$ values of precipitation in central and northern South America due to the impact of the amount effect, as well as moisture transport and rain-out history. On the western side of the Andes at about 20°S , where surface winds are predominantly south-southwesterly, data from summer precipitation sampled during 1974 and 1984 yield $\delta^{18}\text{O}$ lapse rates of -0.2 to $-0.3\text{‰}/100$ m from about 2,500 to 3,500 m elevation, but extremely high $\delta^{18}\text{O}$ lapse rates of about $-1\text{‰}/100$ m at elevations of about 3,500 to 4,500 m (Fritz and others, 1981; Aravena and others, 1999), whereas the altitude effect is negligible in winter and summer rains of 1986. Aravena and others (1999) attribute the extremely high isotopic lapse rates at high elevations immediately west of the Altiplano to the local influence of moisture from the Amazon Basin to the East, which is much more depleted in ^2H and ^{18}O than Pacific-sourced moisture at lower elevations, as well as to a greater importance of convective precipitation (and thus the amount effect) at higher elevations. Thus the

observed spatial and temporal variations of the isotope lapse rates in this segment of the Andes most likely reflect the highly variable wind and precipitation patterns in this region (Zhou and Lau, 1998; Lenters and Cook, 1999; Vuille, 1999) leading to variations in the importance of the amount effect and to mixing of air masses containing isotopically distinct water vapor (Aravena and others, 1999; Gonfiantini and others, 2001).

Still farther north, near the equator, Garcia and others (1998) demonstrate a nearly symmetric pattern in the isotopic composition of precipitation versus elevation on both sides of the Andes (fig. 1, transects C, D), with Atlantic-sourced moisture dominant on the eastern side, and Pacific-sourced moisture dominant on the western side of the mountains. This region is characterized by a negligible seasonality of temperature, but precipitation rates are highly seasonal due to the intensity of the South American summer monsoon in this region (Vuille and Werner, 2005). Accordingly, the isotopic composition of precipitation is strongly controlled by the amount effect, with rainy season $\delta^{18}\text{O}$ values being 5 to 10 permil lower than those in the dry season. Despite this high seasonal variability, Garcia and others (1998) document an oxygen isotope lapse rate of $-0.2\text{‰}/100$ m at elevations $\leq 3,500$ m, in good agreement with precipitation and surface water data presented by Rozanski and Araguás-Araguás (1995). Data from higher elevations in this region yield a higher $\delta^{18}\text{O}$ lapse rate of -0.4 to $-0.5\text{‰}/100$ m; however, such an increase in isotopic lapse rate with elevation can be explained by adiabatic processes (for example, Rowley and others, 2001), and is within the range of $\delta^{18}\text{O}$ lapse rates observed in other high-elevation regions of the world (Poage and Chamberlain, 2001).

Central America

For most of the year, Central American precipitation is derived from the Caribbean Sea by the northeasterly trade winds, but some Pacific-derived precipitation is transported from the West during the summertime northward migration of the ITCZ. While most sites that exhibit isotopic rain shadows are located in temperate regions, as discussed in more detail below, Lachniet and Patterson (2002) found a notable isotopic rain shadow in northern Costa Rica at $\sim 10.5^\circ\text{N}$ with western, leeward, surface waters having $\delta^{18}\text{O}$ values 2 to 3 permil lower than those on the windward eastern side of the $\sim 1,100$ m high mountains (fig. 1, transect A). Lachniet and Patterson (2006) found a similar pattern in Panama ($\sim 8.5^\circ\text{N}$), although the geographic characteristic with the strongest control on the oxygen isotope composition of surface waters collected throughout the country was distance from the Caribbean Sea ($-0.038\text{‰}/\text{km}$ from the Caribbean coast). To the best of our knowledge this is the only documented tropical isotopic rain shadow. However, we are cautious in the importance of this finding because Lachniet and Patterson (2002) present an additional Costa Rican transect at $\sim 9.5^\circ\text{N}$ exhibiting decreasing oxygen isotope ratios of water with increasing elevation on both sides of an $\sim 3,600$ m high mountain range (fig. 1, transect B), suggesting that complex mountain geometries can create unexpected wind patterns that can distort ideal patterns of the isotope distributions in orographic precipitation.

Western North America

The west coast region of North America is characterized by the presence of two main precipitation regimes; at low latitudes in the South, it mostly receives summer precipitation related to the North American monsoon, whereas winter precipitation derived from westerly winds is dominant at middle latitudes in the North (Mock, 1996; Kendall and Coplen, 2001). The influence of the North American monsoon, which is centered over northwestern Mexico and causes northward moisture transport from both the eastern tropical Pacific and the Gulf of Mexico, is most pronounced south of $\sim 25^\circ$ latitude, but reaches at least as far north as southern California and the Great

Basin (see Adams and Comrie, 1997, for a review and additional references). These two distinct precipitation regimes appear to be reflected by characteristic differences in the stable isotope systematics of precipitation in this region. Between $\sim 33^\circ\text{N}$ and 36°N near the hot and dry Mojave Desert of southern California, Friedman and others (1992) found a strong negative correlation between the δD values of precipitation and altitude on both the windward and leeward sides of mountains. On the windward side of mountains this altitude effect can be explained by distillation of the water vapor with pseudo-adiabatic cooling as moist air is forced aloft, whereas the leeward pattern has been interpreted as a “pseudo-altitude effect” resulting from intense below cloud-base evaporation of raindrops (Friedman and others, 1992). In contrast, three detailed transects of the isotopic composition of precipitation and surface waters across the Coast Ranges and the Sierra Nevada (~ 3000 m elevation), at $\sim 37^\circ\text{N}$, 39.5°N and 42°N , demonstrate the presence of pronounced isotopic rain shadows (fig. 2, transects C, D, E). The $\delta^{18}\text{O}$ and δD values of these waters decrease by ~ 8 permil and 60 to 70 permil, respectively, from the windward side (west) to the leeward side (east) of the mountain belt (Friedman and Smith, 1970; Smith and others, 1979; Ingraham and Taylor, 1986; Ingraham and Craig, 1993). We note, however, that the waters nearest the Sierran crest in the southernmost transect (fig. 2, transect E) have $\delta^{18}\text{O}$ values ~ 3 permil lower than those more distant from the crest on both the leeward and windward side. Although the presence of an altitude effect on both sides of the mountains in this segment of the Sierras creates a deviation from an idealized isotopic rain shadow, all of the leeward waters have lower isotope ratios than the more distal windward waters, suggesting that the mountains do lower the isotopic ratio of leeward precipitation. However, in the Great Basin and Rocky Mountains region east of the Sierra Nevada, Pacific moisture from westerly-derived air masses mixes with Arctic moisture brought in by northerly winds, and moisture from the Gulf of Mexico in the South (Bryson and Hare, 1974), and this presumably decreases the importance of the rain shadow effect. North of the Sierra Nevada, precipitation across the $\sim 3,000$ m high Cascade Range ($\sim 45^\circ\text{N}$) is characterized by a 6 to 7 permil decrease in the $\delta^{18}\text{O}$ values from the windward western to the leeward eastern side of the mountains (fig. 2, transect B) (Sammel and Craig, 1981; Ingebritsen and others, 1988; Coplen and Kendall, 2000; as discussed in Kohn and others, 2002, and Smith and others, 2005). Still farther north in southwestern Canada at $\sim 50^\circ\text{N}$, Yonge and others (1989) found an analogous pattern, with a decrease of ~ 5 permil and ~ 40 to 50 permil for the $\delta^{18}\text{O}$ and δD values of precipitation, respectively, across the 2,000 m high Coast Ranges (fig. 2, transect A). However, as this transect continued eastward across the southern Canadian coastal and Rocky Mountains, at a distance of >500 km from the coast, they found “altitude effects” of decreasing isotope ratio of precipitation with increasing altitude on both sides of the mountains, which may be due to a seasonal influence of weather systems from the South (Yonge and others, 1989).

New Zealand Southern Alps

The Southern Alps of New Zealand are characterized by a climatic setting similar to that of the southern Patagonian Andes, with westerly winds providing the primary moisture source and a ten-fold decrease in rainfall from the windward to leeward side of the mountains. This site also exhibits a strongly asymmetric pattern in the $\delta^{18}\text{O}$ and δD values of surface waters across this mountain belt (fig. 3). Along a transect at $\sim 45^\circ\text{S}$, the $\delta^{18}\text{O}$ and δD values of waters on the windward western side of the $\sim 2,000$ m high mountains average -5.5 permil and -30 permil, contrasting with significantly lower values of -9.8 permil and -72 permil in the rain shadow east of the mountains, respectively (as discussed in Chamberlain and others, 1999). Waters with higher values have been observed in a few samples from farther east in the vicinity of the east coast (Stewart and others, 1983), but these probably reflect an influence of water vapor from

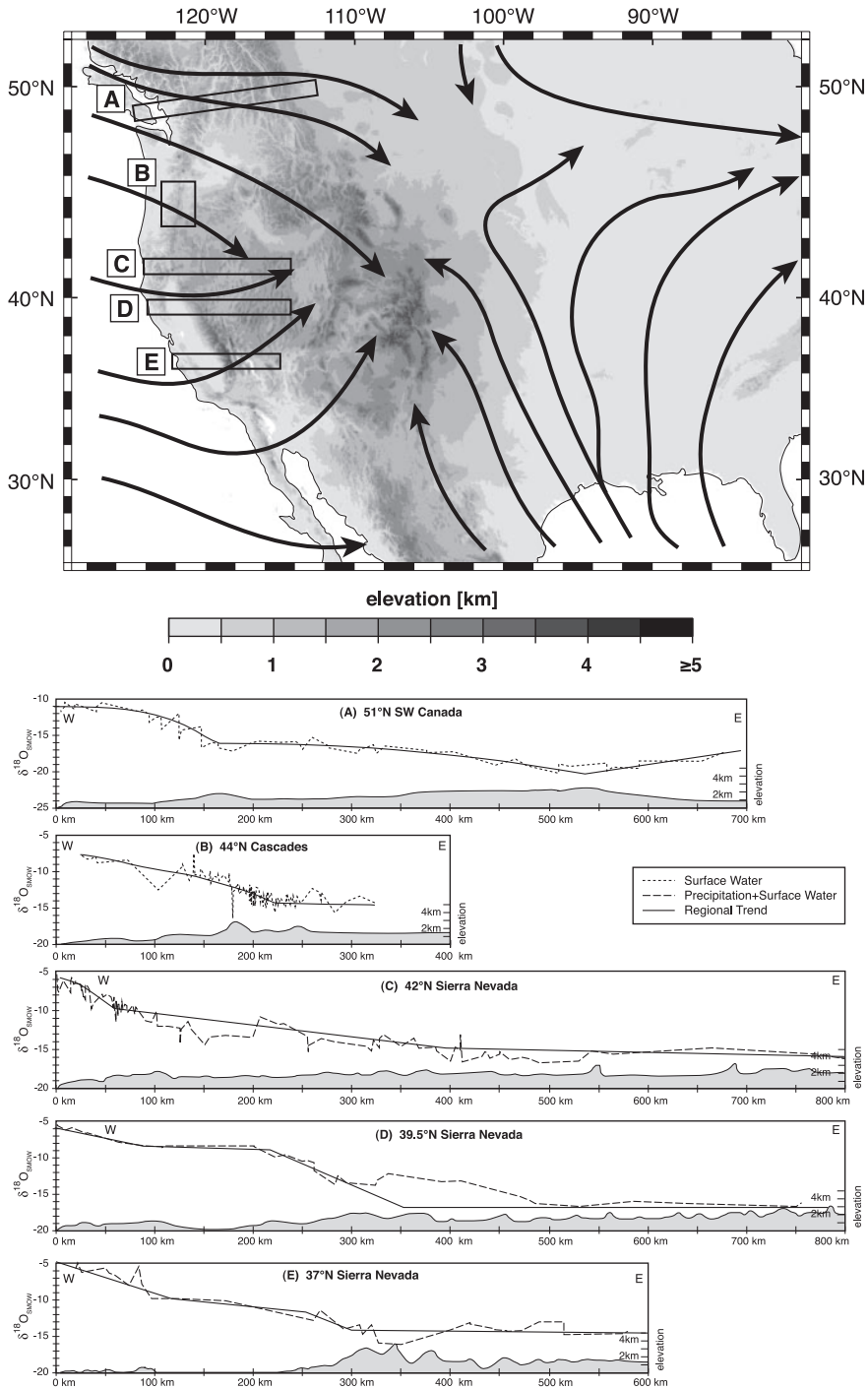


Fig. 2. Top: Topographic sketch map of central and western North America, showing the locations of transects A–E presented to the right, and principal circulation patterns (arrows). Bottom: Oxygen isotope composition of waters across western North America with schematic topographic profiles. (A) SW Canada, after Yonge and others (1989); (B) Cascades, based on data presented by Ingebritsen and others (1988) and Coplen and Kendall (2000); (C, D, E) Sierra Nevada at $\sim 42^\circ\text{N}$, 40°N , and 37°N , respectively, after Ingraham and Taylor (1991).

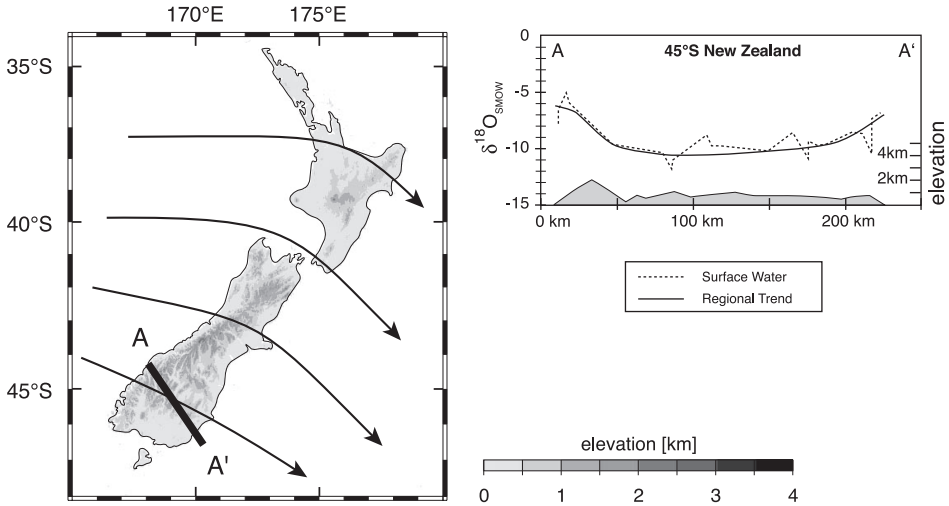


Fig. 3. Left: Topographic sketch map of New Zealand, showing the location of transect A-A' presented to the right, and principal circulation patterns (arrows). Right: Oxygen isotope composition of surface waters along transect A-A' across the Southern Alps of New Zealand with schematic topographic profile, after Chamberlain and others (1999).

the Pacific Ocean to the East. The observed decrease of ~ 4 to 5 permil and ~ 40 permil for the $\delta^{18}\text{O}$ and δD values of surface waters, respectively, has been suggested to be largely due to progressive rain-out during the passage of moist air across the mountains (Chamberlain and others, 1999), which is consistent with an estimate of $-0.28\text{‰}/100$ m for the global average isotopic lapse rate for $\delta^{18}\text{O}$ (Poage and Chamberlain, 2001).

European Alps

The European Alps are characterized by prevailing westerly winds, modified by the influence of the North Atlantic Oscillation (NOA). The NOA is a north-south variability in the strength and position of storm tracks across the North Atlantic and into Europe that is mainly controlled by the position and relative strength of the Azores High and the Icelandic Low (see Hurrell and others, 2003, for a review). During the northern hemisphere winter, when the influence of the high-latitude Icelandic Low predominates, precipitation in the Alps is mainly brought by westerly winds, whereas in the summer, when the Azores High is stronger and farther north than in the winter, northwesterly winds are dominant (Sturman and Wanner, 2001). Schürch and others (2003) and Schotterer and others (1997) found an oxygen isotope altitude effect of $-0.2\text{‰}/100$ m for precipitation at $\sim 47^\circ\text{N}$, 8.5°E , on the northern side of the Central Alps, while on the southern side of the Alps at that longitude ($\sim 45.5^\circ\text{N}$, 8.5°E) Longinelli and Selmo (2003) found an oxygen isotope lapse rate of $-0.18\text{‰}/100$ m. These observed isotopic lapse rates at $\sim 8.5^\circ\text{E}$ are typical Alpine patterns, with those in other transects on the southern side of the Alps ranging from -0.1 to $-0.2\text{‰}/100$ m (Longinelli and Selmo, 2003). Likewise, Yehdegho and Reichl (2002) found similar oxygen isotope altitude effects on both the northern and southern side of the Semmering Massif in the eastern Alps (47°N , 15°E), $-0.27\text{‰}/100$ m and $-0.21\text{‰}/100$ m respectively, based on springs and the altitude of their recharge zones. These patterns of decreasing isotope ratios of precipitation on both the southern and northern flanks of the Alps, along with the absence of an Alpine rain shadow (fig. 4), suggest that the northern and southern sides of the central and eastern Alps derive

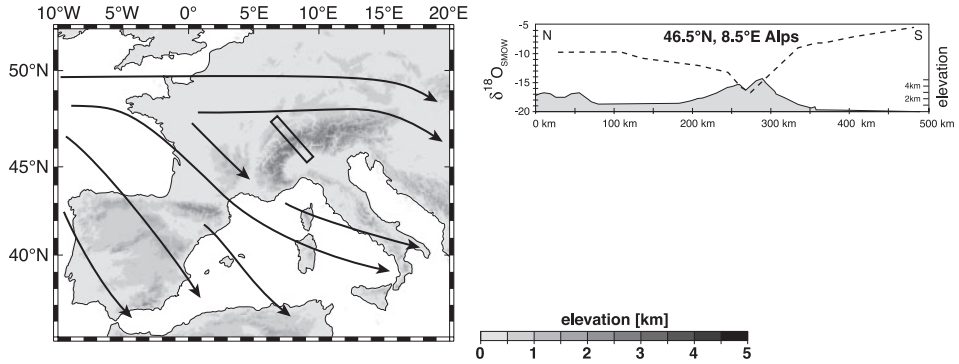


Fig. 4. Left: Topographic sketch map of the Alps and adjacent regions, showing the location of the transect presented to the right (frame), and principal circulation patterns (arrows). Right: Oxygen isotope composition of precipitation across the Swiss/Italian Alps with schematic topographic profile, after Schürch and others (2003) and Longinelli and Selmo (2003).

their moisture predominantly from different sources and orographic uplift is the dominant process controlling condensation and precipitation on both sides of these mountains. Despite the overall pattern of the absence of an Alpine isotopic rain shadow, Dray and others (1998) do document a seasonal isotopic rain shadow in Italy as winter winds carry low $\delta^{18}\text{O}$ moisture across the western Alps, whereas summer precipitation, derived from convective storms, lacks an isotopic rain shadow.

Himalaya/Tibet Orogen

Along the Himalayas precipitation is dominated by rainfall during the summer monsoon, when north- to northwest-directed air masses carrying moisture derived from the equatorial Indian Ocean and the Bay of Bengal (fig. 5) undergo cooling, convection, condensation and progressive rain-out during their passage over the mountains, resulting in profound heavy isotope depletions in precipitation. Several studies document the altitudinal variation in the oxygen isotope composition of

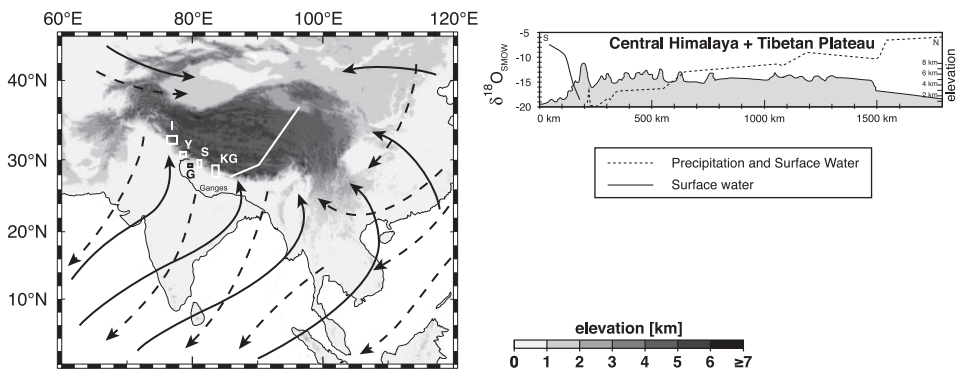


Fig. 5. Left: Topographic sketch map of the Himalaya/Tibet orogen and adjacent regions, showing principal circulation patterns during summer (solid arrows) and winter (stippled arrows). White frames indicate locations of studies discussed in text: G, Gaula (Bartarya and others, 1995); I, Indus (Pande and others, 2000); KG, Kali Gandaki (Garziona and others, 2000b); S, Seti (Garziona and others, 2000b); Y, Yamuna (Dalai and others, 2000). Right: Oxygen isotope composition of summer precipitation and surface waters across the Himalaya/Tibet orogen with schematic topographic profile, after Garziona and others (2000b) and Tian and others (2001).

tributaries to rivers draining the Himalayas towards the South (fig. 5). These studies document altitude effects, from west to east, of: $-0.09\text{‰}/100\text{ m}$ in the Indus River, 77 to 78°E (Pande and others, 2000); $-0.11\text{‰}/100\text{ m}$ in the Yamuna River, 78°E (Dalai and others, 2002); $-0.19\text{‰}/100\text{ m}$ in the Ganga (Ganges) headwaters, 79°E (Ramesh and Sarin, 1992); $-0.14\text{‰}/100\text{ m}$ in the Gaula River, 80°E (Bartarya and others, 1995); $-0.18\text{‰}/100\text{ m}$ in the Seti River, 81°E (Garziona and others, 2000a); and $-0.29\text{‰}/100\text{ m}$ in the Kali Gandaki River, 84°E (Garziona and others, 2000a). Thus the magnitude of the isotopic lapse rates appears to increase toward the East. However, with the exception of the work by Garziona and others (2000a), these studies mix data from trunk streams and tributaries that have quite different catchment elevations, and thus may poorly represent the actual isotopic lapse rates of precipitation. For example, Ramesh and Sarin (1992) estimate that the isotopic lapse rate they obtained from Ganges river waters is only about half of that for precipitation, due to mixing of waters from different altitudes. Some data sets imply a general trend towards higher isotopic lapse rates with increasing elevations, consistent with a Rayleigh distillation model assuming equilibrium fractionation during the uplift of moist air masses from the SE over the high topography of the Himalayas (Garziona and others, 2000a; Rowley and others, 2001). However, it has been argued that the altitude effects observed in the Himalayas may result predominantly from the monsoonal-driven convection controlling the “amount effect” rather than purely uplift-driven cooling (Dalai and others, 2002; Vuille and others, 2005); for example, at most Himalayan sites the average $\delta^{18}\text{O}$ values of summer precipitation are lower than those of winter precipitation despite the higher summer air temperatures (Araguás-Araguás and others, 1998). Furthermore, the interpretation of stable isotope data from the Himalayas is complicated by the mixing of moist air masses sourced from the Bay of Bengal and the Indian Ocean (Lau and others, 2000; Hou and others, 2003).

Transects of the isotopic composition of precipitation across the Himalayas and into Tibet are difficult to assess in terms of an isotopic rain shadow because this is a topographically and climatologically complex region with rather limited observations of the isotopic compositions of water. The high Himalayan Mountains serve as profound topographic barrier, and the main source of moisture to the Tibetan plateau is the Pacific monsoon rather than the Indian Ocean monsoon (Araguás-Araguás and others, 1998), although precipitation in southeastern Tibet is mostly derived from moisture sourced in the Bay of Bengal (Tian and others, 2001). Kang and others (2002) document the oxygen isotope lapse rate on the northern side of the central Himalayas during several storms and found little to no altitude effect for 3 of 4 storms and a low gradient of only $-0.1\text{‰}/100\text{ m}$ during the fourth, suggesting that there may be an isotopic rain shadow in lee of the Himalayas. Farther north, however, there is a general North-to-South decrease in the oxygen isotope ratios of precipitation and surface waters across the Tibetan plateau, despite rather monotonous high topography (fig. 5) (Tian and others, 2001). This is more consistent with the evolution of air masses derived from the North and East moving across the Tibetan plateau. Further evidence of distinct moisture sources for the Himalayan front versus regions to the North comes from the western portion of the orogen. The $\delta^{18}\text{O}$ values of river waters on the southern flank of the western Himalayas range from ca. -11 permil to -17 permil between ca. $\sim 1,000$ and $\sim 4,600\text{ m}$ elevation (Pande and others, 2000). Precipitation on the northern margin of the narrow western Tibetan plateau, in contrast, has much higher $\delta^{18}\text{O}$ values (-5.5‰ at Hetian, elevation $\sim 1,400\text{ m}$; Araguás-Araguás and others, 1998), which is also inconsistent with moisture being derived from the South (fig. 5).

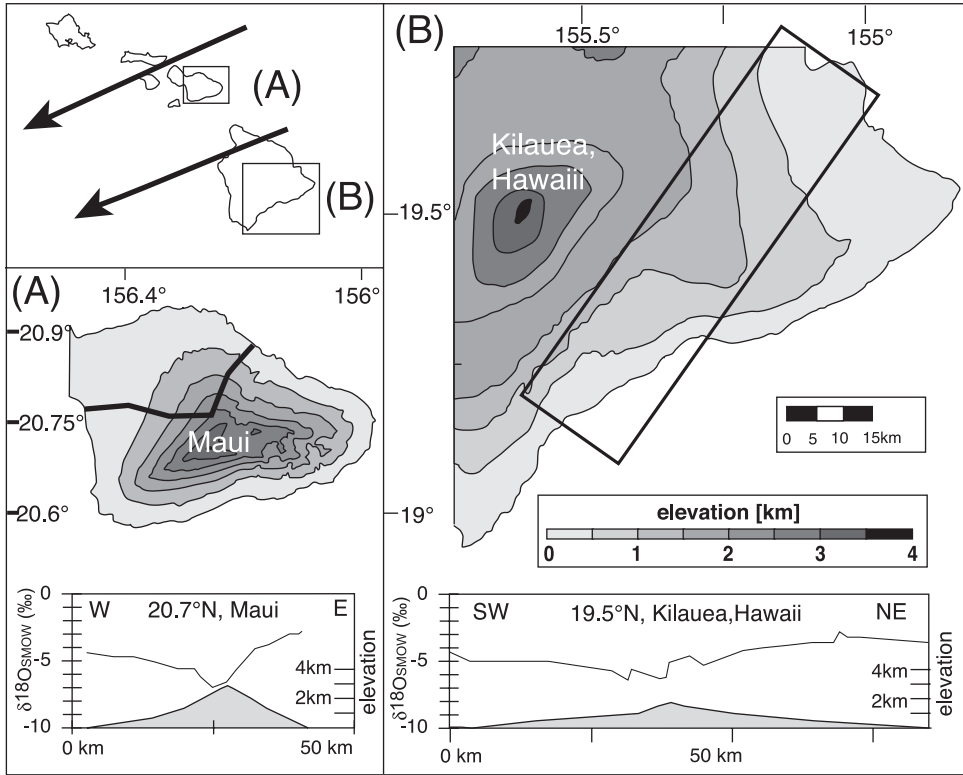


Fig. 6. Inset on upper left: Outline of the Hawaiian Islands with the location of maps for transects A and B, showing surface wind directions (arrows). (A) Top: Topographic sketch map of the eastern half of Maui. Bottom: Topography and oxygen isotope composition of precipitation along the heavy black line shown on the map above (after Scholl and others, 2002). (B) Top: Topographic sketch map of the eastern half of the Kilauea region of the island of Hawaii. Bottom: Topography and oxygen isotope composition of precipitation along the heavy black line shown on the map above (after Scholl and others, 1995, 1996).

Hawaiian Islands

The tropical Pacific islands of Hawaii and Maui are dominated by trade winds from the East (fig. 6, inset), and their eastern slopes receive up to 10 times more precipitation than the western slopes. Precipitation on both the leeward and windward slopes of each island has similar oxygen isotope lapse rates ($-0.12\text{‰}/100\text{ m}$ versus $-0.11\text{‰}/100\text{ m}$ for Maui after Scholl and others, 2002, and $-0.15\text{‰}/100\text{ m}$ versus $-0.16\text{‰}/100\text{ m}$ on the slopes of Kilauea volcano on the Island of Hawaii after Scholl and others, 1995, 1996, 2002). Despite the similar altitude effects, precipitation in the rain shadow on both islands has $\delta^{18}\text{O}$ values about 1.8 permil (Maui; fig. 6, transect A; Scholl and others, 2002) or 1.6 permil (Kilauea; fig. 6, transect B; Scholl and others, 1995, 1996) lower than windward sites at the same elevation, which Scholl and others (1995, 1996, 2002) attribute to precipitation in the rain shadow to be predominantly from convective storms, whereas the windward sites receive both orographic and storm-derived precipitation. The storm-derived precipitation usually has lower $\delta^{18}\text{O}$ values because of higher cloud height and long duration of storms, often beginning before their arrival at the islands. Thus leeward precipitation in the Hawaiian Islands has lower $\delta^{18}\text{O}$ values than windward precipitation, implying the presence of a minor isotopic rain shadow, yet is still characterized by a “normal” altitude effect.

DISCUSSION OF PRECIPITATION AND SURFACE WATER TRANSECTS

The studies summarized above illustrate that the effects of topography and climate on processes controlling the stable isotope distribution in precipitation (fig. 7) can be highly variable. The transects from Patagonia, New Zealand, and western North America (excluding the southernmost sites near the Mojave desert) show pronounced isotopic rain shadows (fig. 1, transect F; fig. 2; fig. 3). These seven transects share several geographic characteristics: (1) They are located in middle latitudes, where westerly winds are dominant year-round. (2) They cross mountain ranges that are in close proximity to windward oceans. (3) The mountains crossed by these transects are oriented nearly perpendicular to the predominant wind direction. (4) The mountains are high, but not broad like the Tibetan plateau or the Altiplano. (5) They are located in temperate climates. These characteristics are optimal for the creation of an isotopic rain shadow for the following reasons: First, the consistent wind directions and the proximity of the mountains to the main moisture source (the ocean upwind) limit the mixing of moisture sources with distinct isotopic compositions (fig. 7A). Second, the orientation and shape of the mountain ranges causes the winds to drive air masses over, rather than around the mountains. Mountain ranges oriented parallel to the prevailing winds or large plateaus may cause winds to be diverted (Barros and Lettenmaier, 1994; Roe, 2005). Finally, the relatively cool temperatures make the amount effect and below cloud-base evaporation relatively unimportant; they also favor high altitude precipitation occurring as snow which, along with high wind speeds of the “roaring forties,” enhances the incidence of spillover precipitation (Colle, 2004; Smith and Barstad, 2004; Smith and others, 2005).

The studies from the European Alps illustrate the influence of an elongate mountain belt's orientation relative to a predominant wind direction. The Alps are in a similar latitude and climatic regime to the transects with isotopic rain shadows, with abundant precipitation in the form of snow. However, in the central and eastern Alps the approximate east-west orientation of the mountain belt, nearly parallel to the predominant wind direction, apparently limits spillover and inhibits the expression of an isotopic rain shadow (fig. 7B). In contrast, an isotopic rain shadow has been observed in winter precipitation southeast of the more north-south trending western Alps, reinforcing the importance of wind orientation relative to a mountain range for the expression of an isotopic rain shadow.

The other transects characterized by altitude effects on both sides of mountains are also informative in identifying the importance of geographic sites characteristics for isotopic rain shadow formation. The transect from southwestern Canada, for example, reinforces the importance of relatively steady wind directions and the proximity of a mountain range to a single dominant moisture source in creating an isotopic rain shadow. In the coastal range portion of the transect, precipitation derived from Pacific-sourced moisture carried over the mountains by predominantly westerly winds exhibits an isotopic rain shadow (fig. 7A). In contrast, precipitation over the inland Canadian Rocky Mountains ~ 500 km farther east, where mixture with other moisture-bearing air masses is likely, shows a nearly symmetric pattern in the oxygen isotope composition of waters on eastern and western flanks of the mountains (fig. 2, transect A; fig. 7C).

The transects from tropical and subtropical regions support the importance of temperature in controlling the presence or absence of isotopic rain shadows. Although some of these studies document weak isotopic rain shadows (Hawaiian Islands, two of the three transects from Central America), most studies observe altitude effects on both sides of mountains (Hawaiian Islands, Central and Northern Andes, Central America, Mojave Desert). Apparently, the importance of convective storms, frequent in tropical regions, leads to a predominance of the amount effect on the isotope ratio

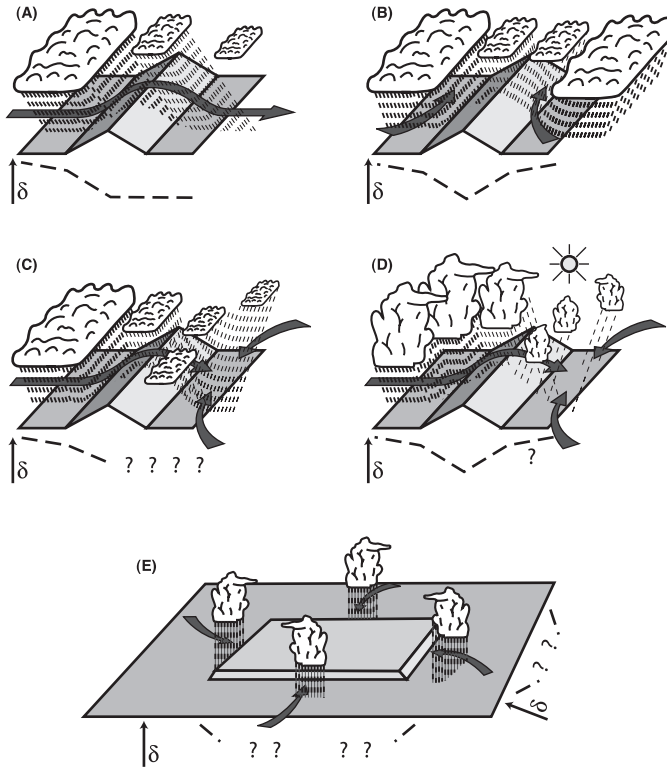


Fig. 7. Schematic drawings illustrating the variable effects of topography and climate on processes controlling the stable isotope composition and distribution of precipitation. Arrows indicate predominant wind directions. Dashed lines beneath each drawing indicate hypothetical general patterns in the $\delta^{18}\text{O}$ and δD values of precipitation. The regions where the geographic patterns of the isotopic composition of precipitation has uncertainties are indicated with question marks. (A) Mountain range situated in a temperate region with a single predominant wind direction, and oriented roughly perpendicular to the prevailing winds. In situations like this, there typically is a systematic decrease in the $\delta^{18}\text{O}$ and δD values of precipitation with altitude on the windward side of the mountains (the "altitude effect"). On the leeward side, in contrast, where precipitation is dominated by spillover from the windward side of the mountains, the $\delta^{18}\text{O}$ and δD values of precipitation are relatively low (reflecting the elevation of the orographic barrier) and independent from local elevation. This condition is referred to as an "isotopic rain shadow." (B) Mountain range oriented nearly parallel to predominant wind direction. In situations like this, there is an altitude effect on both sides of the mountains. (C) Mountain range oriented nearly perpendicular to a predominant wind direction, but with mixing of air masses on the leeward side. In this situation, there is an altitude effect on the windward side, and there may be an altitude effect on the leeward side of the mountains. However, depending on the isotopic composition of the different air masses, and on the amount and geometry of mixing between them, the isotopic composition of precipitation may be predominantly controlled by these mixing processes, not altitude. (D) Mountain range as described in (C), but with a warm and dry climate leading to significant below cloud-base evaporation on the leeward side. In this situation, there is an altitude effect on the windward side, and there may be an altitude effect on the leeward side of the mountains. As in (C), the isotopic composition of precipitation may be predominantly controlled by mixing of air masses, but raindrops at higher elevations, where fallout distances to the ground are shorter, experience less evaporation, resulting in a "pseudo-altitude effect" on the leeward side. (E) Large and highly elevated plateau. Particularly at low latitudes, plateaus promote monsoonal conditions, with most precipitation occurring during summer when sensible heating of the plateau surface leads to low surface air pressures over the plateau. This, in turn, results in surface winds towards the plateau, high precipitation rates along the plateau margins, and mixing of air masses over the plateau. Accordingly, systematic altitude effects are more likely along plateau margins as compared to the region over the plateau, where mixing of air masses is to be expected. However, in low latitudes the isotopic composition of precipitation along plateau margins is likely to be dominated by the high variations in precipitation rates (through the "amount effect") and in the importance of convection associated with both short- and long-term monsoonal climate fluctuations; in such settings, the interpretation of stable isotope data in terms of paleotopography requires a particularly detailed assessment of paleoclimate.

of precipitation, obscuring the rain shadow effect. Furthermore, the dominance of liquid precipitation is likely to enhance pseudo-altitude effects resulting from below cloud-base evaporation, and limit spillover (fig. 7D).

Finally, the effects of broad plateaus on the isotope ratios of precipitation (fig. 7E) are illustrated by the data from the Himalaya/Tibet region and the Central Andes. The Tibetan plateau and the Altiplano promote monsoon climates, with precipitation occurring predominantly during the summer when plateaus act as elevated sources of sensible heat. This sensible heating of the plateau surface, together with latent heat released during condensation and precipitation focused on the plateau margins, localizes strong surface pressure lows, which in turn leads to the flow of moist air towards the low-pressure region, and ultimately strong summer rains. Correspondingly, the isotopic composition of precipitation along plateau margins may undergo strong seasonal variations, as well as interannual variations due to fluctuations in monsoon intensity, but the plateau margins generally display an isotopic altitude effect (fig. 7E). The isotopic composition of precipitation over plateaus is generally characterized by strong heavy isotope depletions, but poorly documented in detail. The isotope ratios of precipitation over the Tibetan plateau may largely be controlled by mixing of air masses containing isotopically distinct moisture derived from different source regions (fig. 7E). However, the limited sample density prevents the assessment of the presence or absence of an isotopic rain shadow across the main Himalayan Range. There is, in contrast, some evidence of transport of low $\delta^{18}\text{O}$ value moisture westward across the Altiplano during wet years, perhaps due to its smaller size and lower elevation as compared to Tibet.

IMPLICATIONS FOR STABLE ISOTOPE-BASED PALEOALTIMETRY STUDIES

The studies summarized above provide some insight and constraints on the interpretation of stable isotope-based paleoaltimetry studies. At temperate midlatitude sites, mountain ranges oriented roughly perpendicular to year-round westerly winds commonly have isotopic rain shadows, with precipitation on the leeward side of the mountains exhibiting systematically lower isotopic ratios than the windward precipitation, and lacking an altitude effect. In such settings, the isotopic composition of surficially formed minerals from the windward side of mountains should be interpreted as a proxy for the paleoelevation of the sampling site, whereas data from leeward sites should be interpreted as a proxy for the paleoelevation of the windward mountains, largely independent of the site paleoelevation. To infer whether or not a site may have had a pronounced isotopic rain shadow, paleogeographic parameters such as paleolatitude, mountain geometry and paleowind directions must be estimated.

In most tropical and subtropical regions, in contrast, isotopic altitude effects are observed on both sides of mountains. Even on the Hawaiian Islands, where there is an overall ~ 2 permil decrease in the oxygen isotope ratio of precipitation from the windward to the leeward side of the mountains at any given altitude, there is a similar isotopic lapse rate on both sides (Scholl and others, 2002). Thus we cautiously generalize that in many low latitude locations the oxygen or hydrogen isotope ratio of authigenic minerals may be used as a direct proxy for paleoelevation at the location of mineral formation, irrespective of the aspect of the slopes. However, at some locations the isotope composition of tropical precipitation is strongly controlled by climatic processes such as the ENSO (for example, Gouffier and others, 2001), the migration of the ITCZ (for example, Rozanski and Araguás-Araguás, 1995), and monsoonal circulation patterns (Aravena and others, 1999; Johnson and Ingram, 2004; Vuille and Werner, 2005; Vuille and others, 2005). Where these phenomena are important, a change in one these processes may affect the isotopic composition of precipitation

more than even a radical change in elevation, and estimating changes in these circulation patterns in the past is difficult.

Tectonic settings, which dictate the location and geometry of mountain ranges, can also influence the interpretation of stable isotope results. Mountains at subduction zones are located near oceans, thus where paleosubduction zones were on the windward coast of a continental plate and oriented roughly perpendicular to the prevailing winds, they are likely to have had an isotopic rain shadow. Mountains in continental collision zones are more likely to have been located within continental interiors, and such regions are more susceptible to mixing of multiple moisture sources.

STABLE ISOTOPE-BASED PALEOELEVATION ESTIMATES

Background

The systematic variations in the isotopic composition of precipitation with topography described above are potentially useful for reconstructions of paleotopography because the isotopic composition of paleoprecipitation can be recorded by authigenic minerals that contain oxygen and/or hydrogen. Hydrogen isotope values of authigenic minerals are, however, more susceptible to evaporation and secondary exchange; accordingly, most studies of minerals containing both oxygen and hydrogen infer spatial and/or temporal trends from the $\delta^{18}\text{O}$ data, and limit the use of the δD data to evaluating the importance of evaporative effects. Examples of mineral proxies that have been used to infer the isotopic composition of paleoprecipitation for paleoelevation estimates include: pedogenic (soil-formed) minerals in paleosols (fossil soils) such as pedogenic carbonate (Garziona and others, 2000b; Blisniuk and others, 2005) or pedogenic clay (Lawrence and Rashkes Meaux, 1993; Chamberlain and others, 1999; Poage and Chamberlain, 2002; Takeuchi and Larson, 2005; Sjostrom and others, 2006), biogenic materials such as fossil shells (Dettman and Lohmann, 2000; Garziona and others, 2000b; Dettman and others, 2001; Morrill and Koch, 2002) or teeth (Koch and others, 1995; Dettman and others, 2001; Kohn and others, 2002; Fricke, 2003), lake carbonates (Drummond and others, 1993; Norris and others, 1996, 2000; Garziona and others, 2000b; Dettman and others, 2003; Horton and others, 2004), cherts (Horton and others, 2004; Abruzzese and others, 2005), fluid inclusions (Winograd and others, 1985), and carbonate cements (Poage and Chamberlain, 2002; Graham and others, 2005). Recent studies have even used δD values of hydrothermal minerals to estimate the hydrogen isotope composition of paleoprecipitation (Mulch and others, 2004; Sharp and others, 2005); here, however, we only review studies using minerals formed at or near the earth's surface.

The $\delta^{18}\text{O}$ (or δD) value of such minerals is largely controlled by the isotopic composition of the water present during mineral formation and mineral/water fractionation factors, which vary with temperature (fig. 8A); in the case of fossils, vital effects can present an additional influence on the isotope ratio of biogenic minerals (for example, McConnaughey, 1989; Koch and others, 1995; Fricke and others, 1998). The temperature dependence of mineral/water fractionation factors is not strictly linear, but does not vary greatly over a reasonable range of earth surface temperatures (fig. 8B); for example, among all the minerals discussed in this paper (calcite, aragonite, kaolinite, smectite, chert, and apatite) the oxygen fractionation factors vary between -0.19 to $-0.31\text{‰}/^{\circ}\text{C}$ in the temperature range of 0 to 30°C . Accordingly, if the temperature at which authigenic minerals formed is known (that is, typically the mean annual temperature as a proxy for surface water or soil temperature, or the body temperature in the case of fossils), the $\delta^{18}\text{O}$ value of the water present during mineral formation can be estimated. In the case of biogenic apatite from mammal teeth, the constant body temperature of mammals ($\sim 37^{\circ}\text{C}$) makes changes in the mineral/water

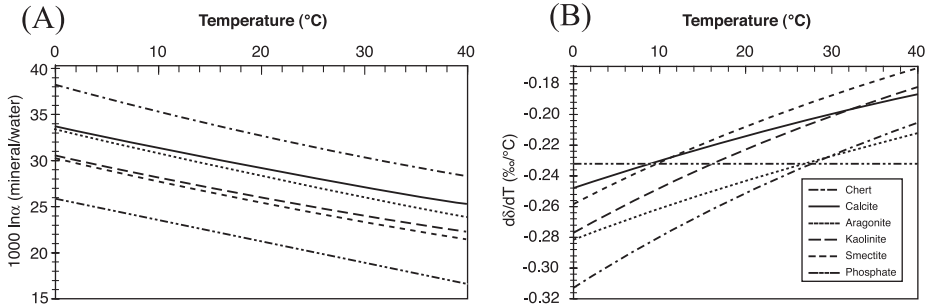


Fig. 8. (A) Oxygen isotope fractionation factors between minerals and water over a range of reasonable earth surface temperatures, expressed as $1000 \ln \alpha$. Chert from Knauth and Epstein (1976); calcite from Kim and O'Neil (1997); aragonite from Zhou and Zheng (2003); kaolinite and smectite from Sheppard and Gilg (1996); phosphate is for fish bone from Longinelli and Nuti (1973) and is expressed as Δ phosphate-water rather than $1000 \ln \alpha$. (B) The change in the temperature dependence of oxygen isotope fractionation versus temperature for the same temperatures range, and based on the same studies, as in (A).

fractionation factor unimportant. However, to use the $\delta^{18}\text{O}$ value of the water present during mineral formation as a proxy for paleoelevation requires additional inferences on local geographic and climatic conditions. Pedogenic minerals may have formed in isotopic equilibrium with soil water enriched in ^{18}O and ^2H relative to precipitation as the result of evaporation (for example, Hsieh and others, 1998). For chemically precipitated or biogenic minerals in lacustrine and fluvial deposits it is important to assess what the catchment area of the corresponding surface waters was; river and lake water may represent precipitation from high above the site elevation (Friedman and others, 1964; Kendall and Coplen, 2001). Furthermore, surface waters (especially lakes) may experience significant amounts of evaporation. In the case of studies on plate tectonic time scales, it can also be important to consider the effect of changes in the paleolatitude of the studied site on the oxygen isotope composition of local precipitation. Another factor affecting the interpretation of stable isotope values of authigenic minerals is the mineral formation history. Mineral growth can be seasonal; for example, pedogenic calcite may predominantly form during dry seasons, when soil waters have experienced more evaporation (Stern and others, 1997; Tabor and Montañez, 2002, 2005). Also, differences in the total duration of mineral formation in a sample make different minerals record very different time periods; for example, the relatively short life span of animals in the case of biogenic minerals, or even seasonal variations in the case of mollusk shells, versus the much longer time spans required for the accumulation of pedogenic carbonate (typically $>10^2$ yrs), or even longer time spans required for the accumulation of pedogenic clay (typically $>10^3$ yrs).

Another critical consideration in the interpretation of mineral stable isotope compositions is whether authigenic minerals have experienced alteration (for example, Lohmann, 1988). Different minerals have variable susceptibility to diagenetic alteration of their primary isotopic compositions; chert and phosphate are particularly robust due to the low rate of silica dissolution (White and Brantley, 1995) and the low rate oxygen isotope exchange of the phosphate ion with water by inorganic processes (Zazzo and others, 2004) under non-acidic conditions (O'Neil and others, 2003). Likewise, clay minerals are resistant to secondary oxygen isotope exchange under low-temperature conditions (Yeh and Epstein, 1978; Savin and Hsieh, 1998). Shells composed of aragonite or high magnesium calcite are highly susceptible to changes of their primary oxygen isotope composition during the alteration to secondary calcite; this transformation is common under near-surface conditions due to the chemical

instability of aragonite. However, if fossil aragonite shells are preserved, one can be assured that they have experienced little change in their isotopic composition. Calcite is relatively stable under near-surface conditions, but may recrystallize during burial and heating, and/or in the presence of secondary fluids (for example, Banner and Kaufman, 1994). To identify alteration of a shell (or other materials) composed of primary calcite, petrographic techniques may be used; preservation of primary textures, along with the absence of sparry cement, and the absence of luminescent calcite under optical cathodoluminescence all support little alteration (for example, Banner and Kaufman, 1994). Similarly, the δD value of fluid inclusions may not represent that of the primary fluids, and great care must be taken in the extraction of these fluids for isotopic analysis (see for example, Dennis and others, 2001).

Considering the various influences on the stable isotope composition of authigenic minerals summarized in the previous paragraphs, there is no simple rule for the interpretation of stable isotope data with respect to paleotopography. To illustrate how this affects the interpretation of stable isotope data sets, we discuss results of previous paleoaltimetry studies from different geographic and climatic settings below.

Southern Patagonian Andes

The Lower to Middle Miocene elevation history of the southern Patagonian Andes has been inferred from the $\delta^{18}O$ values of pedogenic carbonate in terrestrial foreland basin deposits in the present-day rain shadow east of the Andes (fig. 9A; Blisniuk and others, 2005). The samples used in this study were dense micritic carbonate nodules ($n = 268$) from 102 well-developed paleosol horizons within a sequence dominated by alternating sand-, silt- and claystone as well as minor conglomerate beds (the Santa Cruz Formation), which is ca. 500 m thick and 23 to 14 Myr old in the study area (Blisniuk and others, 2005). The paleosol horizons are relatively rare in the ca. 23 to 18 Ma lower part of the section, and much more common in the younger part of the section, particularly after ~ 17 Ma.

The main trend in the $\delta^{18}O$ data is an ~ 2 permil decrease of the mean $\delta^{18}O$ values from the lower to the upper part of the studied section; most of this decrease occurs in the ~ 17 to 14 Ma upper ~ 300 m of the section, and it is particularly pronounced at ~ 200 m above the base of the studied section, or ~ 17 Ma (Blisniuk and others, 2005). This shift been interpreted as the result of significant surface uplift and rain shadow formation in this Andean segment. Influences such as mixing of isotopically distinct moisture sources, changes in atmospheric circulation patterns, and changes in the latitude of the studied site are unimportant at this location (Ziegler and others, 1981; Hartnady and le Roex, 1985). Similarly, regional climatic changes that may influence the oxygen isotope values of pedogenic carbonate were relatively minor during the time interval of interest. The marine oxygen isotope record and the results of ocean circulation models suggest that the South Pacific region has undergone regional cooling by $\sim 3^\circ C$ from ~ 24 Ma to ~ 20 Ma; this temperature decrease was presumably related to Drake Passage opening at ~ 24 Ma, and had similar effects on sea surface temperatures and temperatures on land (Flower and Kennett, 1993, 1995; Nong and others, 2000; Toggweiler and Bjornsson, 2000; Zachos and others, 2001; Pfuhl and McCave, 2005). This cooling episode was followed by a similar temperature increase from ~ 20 Ma to ~ 17 Ma, and the 17 to 14 Ma “middle Miocene climatic optimum” during which temperatures fluctuation were minor (Flower and Kennett, 1993, 1995; Zachos and others, 2001).

The net effect of these regional climatic changes on the oxygen isotope ratio of pedogenic carbonate in the studied section can be crudely estimated as an ~ 1 permil increase from 23 to 20 Ma, and a similar decrease from ca. 20 to 17 Ma (Blisniuk and others, 2005); this is based on estimates for the effects of ice volume changes on South Pacific ocean water (an $\sim 0.3\text{‰}$ increase from 23-20 Ma, and an $\sim 0.3\text{‰}$ increase from

20-17 Ma), decreasing sea surface temperatures ($-0.55\text{‰}/^{\circ}\text{C}$), as well as decreasing temperatures on land on the oxygen isotope values of precipitation ($+0.58\text{‰}/^{\circ}\text{C}$) and on the calcite/water fractionation factor ($-0.23\text{‰}/^{\circ}\text{C}$). Accordingly, regional climatic changes are not a likely cause for the observed decrease in the $\delta^{18}\text{O}$ value of pedogenic carbonate east of the southern Patagonian Andes after ~ 17 Ma. However, major regional cooling since ~ 14 Ma (for example, Miller and others, 1987; Flower and Kennett, 1993; Zachos and others, 2001; Shevenell and others, 2004; Holbourn and others, 2005) may have affected samples in the uppermost part of the studied section; this would presumably lead to an increase in the $\delta^{18}\text{O}$ value of pedogenic carbonate, and partially mask the decrease resulting from surface uplift. Moreover, a transition to a more arid climate in the eastern foreland at ~ 17 Ma, inferred from the $\delta^{13}\text{C}$ values of the studied carbonate nodules (Blisniuk and others, 2005) and consistent with less precisely dated changes in vertebrate faunas (Pascual and Ortiz Jaureguizar, 1990; Flynn and Swisher, 1995), is likely to have promoted evaporative ^{18}O enrichment of soil waters, and to have affected some pedogenic carbonate samples. This hypothesis is supported by a marked increase in the variability of the $\delta^{18}\text{O}$ values from pedogenic carbonate samples at ~ 17 Ma, which is largely directed towards lower $\delta^{18}\text{O}$ values (Blisniuk and others, 2005). Correspondingly, the ~ 2 permil decrease of $\delta^{18}\text{O}$ values in the mean soil water composition recorded by these data probably underestimates the shift in the $\delta^{18}\text{O}$ values of precipitation. It has therefore been suggested that the 3 to 4 permil decrease of the lowest $\delta^{18}\text{O}$ values (presumably from the samples least affected by evaporation) is a better measure for the magnitude of the shift in the isotopic composition of precipitation (Blisniuk and others, 2005). Using the global isotopic lapse rate estimate of $-0.28\text{‰}/100$ m (Poage and Chamberlain, 2001), a 3 to 4 permil decrease in the $\delta^{18}\text{O}$ value of precipitation would imply that 1.3 ± 0.2 km of surface uplift occurred in the southern Patagonian Andes between ~ 17 and 14 Ma. We note, however, that uncertainties related to the effects of climate change, as well as uncertainties related to the isotopic lapse rate will increase the total uncertainty of this estimate. It is probably more realistic to infer an isotopic shift of 3.5 ± 1.5 permil and a $\delta^{18}\text{O}$ lapse rate of $-0.3 \pm 0.1\text{‰}/100$ m, resulting in an estimate of 1.2 ± 0.5 km for the amount of this surface uplift.

Western North America

In this section we review stable isotope studies that have implications for the elevational history of the western United States. In particular, these studies provide constraints on the topographic evolution of three physiogeographic provinces: 1) the Cascade Range, 2) the Sierra Nevada and Great Basin region, and 3) the eastern Rocky Mountains.

In the present-day rain shadow of the Cascade Range at $\sim 43^{\circ}\text{N}$, Kohn and others (2002) found an ~ 5 permil decrease in the mean $\delta^{18}\text{O}$ value of fossil equid tooth enamel since ~ 27 Ma (fig. 9B). Estimating that global cooling may have caused an ~ 1 permil decrease, and that increased evaporation may have led to an empirically based increase of at least 1.5 to 3 permil in tooth $\delta^{18}\text{O}$, they infer a topographically driven 6 to 8 permil decrease in the $\delta^{18}\text{O}$ value of precipitation since the Late Oligocene, which is similar to the present-day ~ 7 permil difference in the $\delta^{18}\text{O}$ values of precipitation across the Cascade Range (Sammel and Craig, 1981; Ingebritsen and others, 1988; Coplen and Kendall, 2000). From this, they infer a relatively steady elevation increase of the Cascade Range since the Late Oligocene. While geologic models of the topographic evolution of the Cascade Range are controversial (see discussion in Kohn and others, 2002), this interpretation is in good agreement with paleontologic and paleopedologic evidence for a progressive increase in aridity east of the Cascade since the Late Oligocene (Ashwill, 1983; Retallack and others, 2000; Retallack, 2004). It is also strongly supported by more recent work of Takeuchi and Larson (2005), who

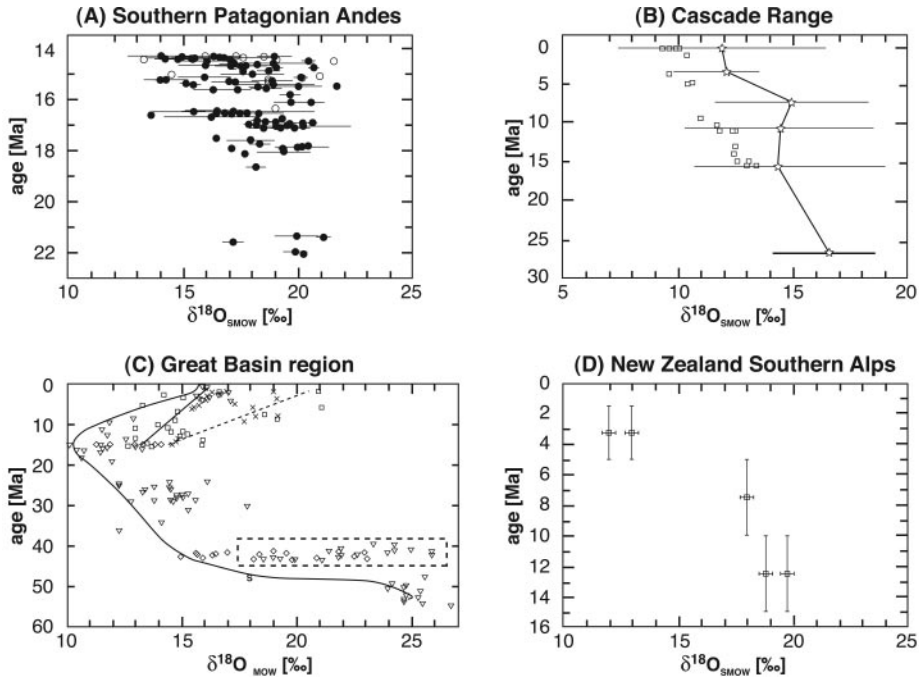


Fig. 9. Examples for data sets that have been used to infer paleotopography. (A) The $\delta^{18}\text{O}$ values of pedogenic carbonate nodules in the rain shadow east of the southern Patagonian Andes. Filled circles represent average $\delta^{18}\text{O}$ value from several (2–5, typically 3) nodules, horizontal lines are error bars representing one standard deviation; for filled circles without error bars, one standard deviation is smaller than width of symbol. Open circles represent $\delta^{18}\text{O}$ value of single nodule. After Blisniuk and others (2005). (B) The $\delta^{18}\text{O}$ values of fossil equid teeth enamel (star symbols, after Kohn and others, 2002) and pedogenic smectite (squares, after Takeuchi and Larson, 2005) in the rain shadow east of the Cascade Range. The star symbols indicate the mean isotopic composition of 2–13 individual teeth per sampled horizon, the horizontal lines represent the range of $\delta^{18}\text{O}$ values for the studied tooth samples. (C) The $\delta^{18}\text{O}$ values of authigenic smectite (squares, after Poage and Chamberlain, 2002; and Horton and others, 2004), lacustrine carbonate (triangles, after Horton and others, 2004) and chert (diamonds, after Abruzzese and others, 2005), and carbonate cement (rotated crosses, after Poage and Chamberlain, 2002) in the Great Basin region. Curved line shows the isotopic trend inferred from all samples thought to represent unevaporated surface waters (after Abruzzese and others, 2005). Straight lines in the upper part of the section show separate trends of samples from the northern (solid line) and southern (stippled line) parts of the western Great Basin region (after Poage and Chamberlain, 2002), interpreted to imply a southward increasing elevation loss of the Sierra Nevada during Mio-/Pliocene time (Poage and Chamberlain, 2002). (D) The $\delta^{18}\text{O}$ values of authigenic kaolinite (squares) in the rain shadow east of the Southern Alps of New Zealand. The vertical (horizontal) error bars represent approximate age (analytical) uncertainties. After Chamberlain and others (1999).

studied the $\delta^{18}\text{O}$ values of authigenic smectites from Miocene to Quaternary deposits east of the Cascades at $\sim 47^\circ\text{N}$. Based on a 3 to 4 permil decrease in the $\delta^{18}\text{O}$ values of these smectites from 15.6 Ma to the Quaternary, and estimating the likely net effect of changes in source water composition due to ice volume changes and decreasing local temperatures to be a minor increase, they infer a 3.5 to 4.5 permil decrease in $\delta^{18}\text{O}$ values of meteoric waters during this time. As the most likely cause for this change, they suggest a 1.2 to 1.7 km elevation increase of the Cascade Range, based on the $-0.28\text{‰}/100\text{ m}$ global isotopic lapse rate estimate of Poage and Chamberlain (2001).

Farther south, a series of stable isotope studies have been conducted on the leeward eastern side of the Sierra Nevada extending eastward into the Basin and Range province (fig. 9C), attempting to constrain the Cenozoic paleoelevational history of this region. In the first of these studies, Winograd and others (1985) estimated that

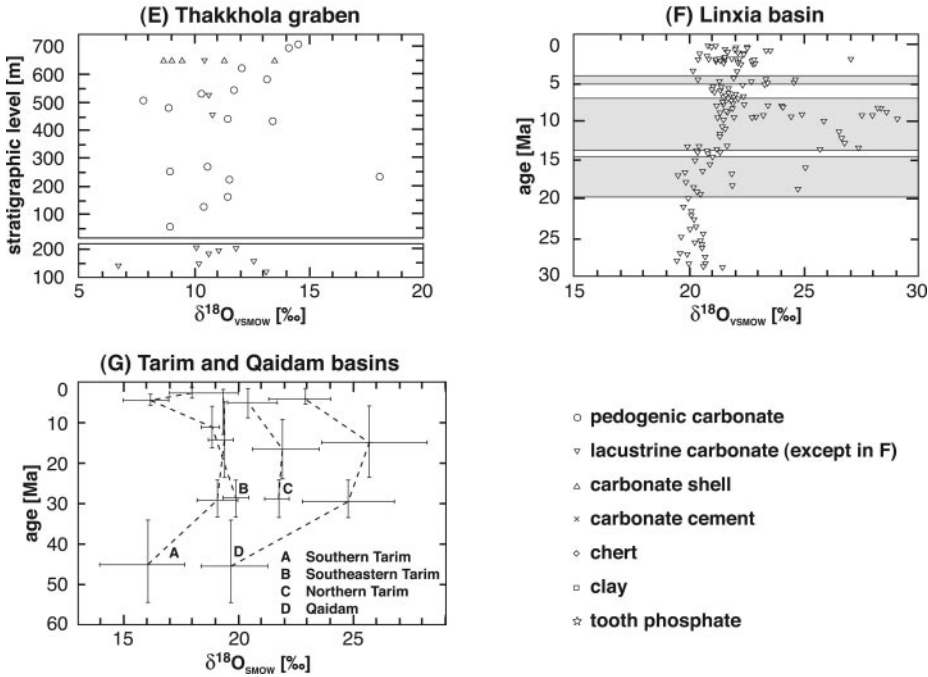


Fig. 9 (continued). (E) The $\delta^{18}\text{O}$ values of pedogenic carbonate (circles), lacustrine carbonate (triangles), and aragonitic shells (inverted triangles) in the Thakkhola graben, Nepal Himalaya. Lower frame shows results from the ~11 Ma to ~9.6 Ma Tetang Formation, upper frame shows results from the ≤ 8 Ma Thakkhola Formation. After Garzzone and others (2000b). (F) The $\delta^{18}\text{O}$ values of lacustrine carbonate (triangles in the shaded portions of the graph, which represent lake deposits) and carbonate cement as well as pedogenic carbonate (triangles outside the shaded portions of the graph) in the Linxia basin northeast of the Tibetan plateau. After Dettman and others (2003). (G) The $\delta^{18}\text{O}$ values of various types of nonmarine carbonate (pedogenic carbonate, lacustrine carbonate, carbonate cement) in the Tarim and Qaidam basin regions north and northeast of the Tibetan plateau. The vertical error bars represent the approximate age uncertainties, the horizontal bars show the total range of analytical values from up to 12 samples per age bracket. After Graham and others (2005).

~600 m uplift of the central Sierra Nevada occurred over the last 2 Ma, based on an ~40 permil decrease in the D/H ratio of fluid inclusions in dated calcite-bearing veins which were thought to represent a viable proxy for paleogroundwater. However, later work in this region has questioned this interpretation. In the western Basin and Range region, Poage and Chamberlain (2002) used changes in the $\delta^{18}\text{O}$ values of smectites formed by weathering of volcanic ashes as well as carbonate cements in Mio-Pliocene sediments as proxies for changes in the $\delta^{18}\text{O}$ values of paleoprecipitation (fig. 9C). In 14 to 6 Ma sediments at ~35°N, they observed an ~5 to 6 permil increase in the $\delta^{18}\text{O}$ values of smectites, and a similar trend for carbonate cements. In a 16 to 2 Ma composite section combining data from 6 locations between ~37°N and 39°N, the $\delta^{18}\text{O}$ values of smectites increase by ~2 to 3 permil and carbonate cements, limited to the younger part of the section, show a similar trend. Estimating the likely effects of changes in the $\delta^{18}\text{O}$ values of ocean water and local temperature since the Miocene to cause an ~0.6 permil increase in the $\delta^{18}\text{O}$ values of authigenic smectites, they suggested that most of the observed change in the $\delta^{18}\text{O}$ values of the analyzed smectites is related to the topographic history of the Sierra Nevada. Using an isotopic lapse rate of $-0.28\text{‰}/100\text{ m}$ (Poage and Chamberlain, 2001) to estimate the magnitude of this elevation change, they interpreted these data to reflect an elevation

decrease of $\sim 2,000$ m in the S, and ~ 700 m in the N. While at odds with some geologic studies indicating significant surface uplift in the Sierra Nevada during the past ~ 10 Ma (Loomis and Burbank, 1988; Unruh, 1991; Clark and others, 2004), this interpretation is consistent with others suggesting a Late Cenozoic elevation decrease which was presumably more pronounced in the south (Small and Anderson, 1995; Wernicke and others, 1996; House and others, 1998, 2001). Farther east in the Basin and Range Province, authigenic smectite from ~ 16 Ma to ~ 1 Ma ashes collected from 6 sections in the northern Basin and Range (Horton and others, 2004) and authigenic smectite and carbonate from Miocene to Quaternary deposits at 6 sites in the central Basin and Range Province and adjacent areas to the South (Poulson and John, 2003; Horton and Chamberlain, 2006) generally show a similar ~ 5 permil increase since the middle Miocene. The only exception to this regional pattern is a 25 to 13 Ma decrease documented in the Lake Mead area, which has been interpreted to reflect an ~ 1.5 km increase in local topographic relief due to magmatic activity in the eastern central Basin and Range Province during this time (Poulson and John, 2003; Horton and Chamberlain, 2006). The nearly ubiquitous Miocene to Quaternary $\delta^{18}\text{O}$ increase in the southwestern United States implies that this change is regional in nature. One plausible cause for this regional increase in the $\delta^{18}\text{O}$ values of precipitation from the middle Miocene into the Quaternary would be a decrease in the surface elevation of the Sierra Nevada (the interpretation of Poage and Chamberlain, 2002); as discussed in the first part of this paper, the $\delta^{18}\text{O}$ values of precipitation on the leeward side of mountains are independent from local elevation unless below cloud-base evaporation and/or mixing with other moisture-bearing air masses are important. However, precipitation in the Great Basin is derived by mixing of different air masses, with the $\delta^{18}\text{O}$ values of winter precipitation derived from westerly sources being ~ 5 permil lower than the $\delta^{18}\text{O}$ values of summer precipitation derived from more southerly sources. Therefore the 4 to 5 permil increase in the $\delta^{18}\text{O}$ values of paleoprecipitation in this region is more likely the result of an elevation decrease of 1 to 1.5 km in the Basin and Range region in response to Neogene extension and crustal thinning (see Horton and Chamberlain, 2006, for discussion and references), or a change from a predominantly westerly to a predominantly southerly moisture source, or a combination of these two processes (Horton and others, 2004; Horton and Chamberlain, 2006).

The earlier topographic history of the Great Basin has been inferred from a section in the northern Great Basin, where the $\delta^{18}\text{O}$ values of micritic calcite and chert from lacustrine deposits, as well as smectite from weathered ashes, show a systematic decrease by almost 10 permil from ~ 55 Ma to ~ 40 Ma, and an ~ 2 to 3 permil decrease from ~ 40 Ma to ~ 15 Ma (fig. 9C; Horton and others, 2004; Abruzzese and others, 2005). Because the ~ 2 to 3 permil decrease of $\delta^{18}\text{O}$ values from ~ 40 Ma to ~ 15 Ma can be accounted for by changes in temperature alone, it has been assumed that no significant changes in paleotopography occurred during this time (Horton and others, 2004; Abruzzese and others, 2005). The change in the $\delta^{18}\text{O}$ values of authigenic minerals from ~ 55 Ma to ~ 40 Ma, in contrast, has been interpreted as the result of regional topographic uplift. Horton and others (2004) estimated that the likely effects of diagenesis as well as changes in ice volume and moisture source areas during this time are minor. From the Mg/Ca and Sr/Ca values of the studied carbonates, they inferred that most samples were not significantly affected by evaporation, with the exception of a narrow stratigraphic interval at ~ 41 Ma, where high Mg/Ca and Sr/Ca values suggest significant evaporation, and where a relatively large scatter in the $\delta^{18}\text{O}$ values of both carbonate and chert samples has been observed (Horton and others, 2004; Abruzzese and others, 2005); from this interval, they considered only the lowest $\delta^{18}\text{O}$ values in their interpretation, assuming that these reflect the isotopic composi-

tion of the least evaporated waters. However, from independent records of paleoclimate (Wolfe, 1994) they estimated that decreasing mean annual temperatures may account for ~ 4 permil of the observed ~ 10 permil decrease during this time. Accordingly, they inferred a topographically driven ~ 6 permil decrease in the $\delta^{18}\text{O}$ value of paleoprecipitation in response to an ~ 2 km increase in surface elevation. They propose this surface elevation change to be local to the Basin and Range rather than due to an enhancement of the isotopic rain shadow of the Sierra Nevada, based on independent evidence suggesting that the Sierra Nevada was already up to 4 km high by the Late Cretaceous, and on evidence for Eocene crustal shortening and thickening in the Basin and Range Province (see Horton and others, 2004, for a more detailed discussion and references).

Farther east in the Rocky Mountains at $\sim 42^\circ\text{N}$, Koch and others (1995) measured the $\delta^{18}\text{O}$ values of pedogenic carbonate nodules, biogenic carbonate in bivalve shells, and tooth enamel apatite from Paleocene-Eocene deposits of the Bighorn Basin, Wyoming. From all three mineral proxies, they estimated that the $\delta^{18}\text{O}$ values of meteoric waters in this region ranged from -8 permil to -12 permil during this time, and interpreted this variability as the result of a temporary influx of surface waters that had low $\delta^{18}\text{O}$ values due to rain-out. Farther south in Wyoming, in the Green River Basin at $\sim 41^\circ\text{N}$, Norris and others (1996, 2000) estimated the $\delta^{18}\text{O}$ values of early to middle Eocene lake waters from the isotopic composition of lacustrine carbonates in the Green River Formation, southern Wyoming. The $\delta^{18}\text{O}_{\text{PDB}}$ value of most of these carbonates was -2.5 permil to -5.0 permil, but values as low as -15.8 permil were observed in several stratigraphic intervals. Carbonates with a $\delta^{18}\text{O}_{\text{PDB}}$ value of -15.8 permil would be in equilibrium with water with a $\delta^{18}\text{O}$ value of -12.9 to -19.8 permil for calcite precipitation temperatures between 0 and 30°C , respectively. Considering that the $\delta^{18}\text{O}$ values of water in modern lakes in the western United States are 4.7 to 11.9 permil higher than those of their tributaries due to evaporation, Norris and others (1996) suggested that the Eocene lake waters in which the carbonates with the lowest $\delta^{18}\text{O}$ values were formed had $\delta^{18}\text{O}$ values of less than -18 permil, as the result of relatively short-lived snow melt events from mountains with minimum altitudes of $>3,000$ m. This is in good agreement with the results of a study by Dettman and Lohmann (2000) who estimated that the $\delta^{18}\text{O}$ values of river waters in several Late Cretaceous and Early Tertiary basins in the Rocky Mountains between 40°N and 50°N ranged from -5 permil to -23 permil, based on the $\delta^{18}\text{O}$ values of aragonite mollusk shells. While acknowledging that part of this variability is likely due to climate variations during the time interval covered by their samples, as well as to the different paleolatitudes of their sampling sites, they also attribute a significant portion of it to the variable influence of basin and highland precipitation. From the 10 to 12 permil maximum range of $\delta^{18}\text{O}$ values within individual basins, and assuming a rather high isotopic lapse rate of $-0.4\text{‰}/100$ m, they estimated that local relief was on the order of $2,500$ to $3,000$ m, similar to present-day topographic conditions.

However, the interpretation of $\delta^{18}\text{O}$ values from authigenic calcite in this region as reliable evidence for high surface elevation or local relief has been questioned by Morrill and Koch (2002) and Fricke (2003). Morrill and Koch (2002) found that freshwater clam shells from middle Eocene deposits of the Green River Formation that were diagenetically altered to calcite had $\delta^{18}\text{O}$ values 3 to 11 permil lower than unaltered aragonite shells, and estimated the $\delta^{18}\text{O}$ values of the paleolake waters as ranging from -7 permil to -12 permil. Similarly, Fricke (2003) estimated the $\delta^{18}\text{O}$ values of early Eocene river water in the Bighorn Basin to range from -6.9 permil to -9.4 permil, based on $\delta^{18}\text{O}$ values of biogenic phosphate in tooth enamel from a semi-aquatic terrestrial mammal that are thought to be a reliable proxy for the isotopic composition of river waters. These studies suggest that instead of high local relief, the

rare low $\delta^{18}\text{O}$ values reported by Norris and others (1996, 2000) and Dettman and Lohmann (2000) might reflect diagenetic alteration, increased seasonality due to changes in orbital parameters, or temporary introduction of moisture from the Pacific that is isotopically depleted after moving over the Sierra Nevada.

Drummond and others (1993) constrained the paleoaltimetry of northwestern Wyoming during the Late Miocene by studying the oxygen isotope composition of lacustrine limestones. They found the $\delta^{18}\text{O}_{\text{PDB}}$ values of these limestones to be exceptionally low, ranging between -21.5 and -35.2 permil, from which they inferred that meltwaters reflecting winter paleoprecipitation had $\delta^{18}\text{O}$ values of -24.6 to -26.7 permil. Modern river waters in Wyoming typically have $\delta^{18}\text{O}$ values as low as -17 or -18 permil (Coplen and Kendall, 2000), so using a $-0.42\text{‰}/100$ m isotopic lapse rate they interpreted their results as indicating that the catchment region (the Gros Ventre Mountains) may have been at elevations up to ~ 2 km higher than today. The absolute altitude changes suggested by Drummond and others (1993) are likely unrealistic in light of more recent constraints on isotopic lapse rates and the mixing of different moisture-bearing air masses in this region (for example, Amundson and others, 1996), but these results do suggest the presence of high mountains in the Late Miocene of northwestern Wyoming.

Still farther east, Sjostrom and others (2006) studied authigenic smectite from weathered volcanic airfall deposits collected along a transect from the eastern Rocky Mountains into the western Great Plains. They observed a persistent ≥ 6 permil decrease in the $\delta^{18}\text{O}$ values of roughly contemporaneous smectites from the western Great Plains to sites near the crest of the Rocky Mountains for both Oligocene and Miocene samples. This geographic pattern is similar to the distribution of isotopes in modern precipitation along this transect. Eocene smectites were sampled along a much shorter transect from central to western Wyoming and span a relatively narrow elevation range, so inferring spatial trends from their $\delta^{18}\text{O}$ values is more ambiguous. Nonetheless, the data clearly reflect relatively low absolute $\delta^{18}\text{O}$ values of Eocene precipitation in central and western Wyoming, consistent with the results of earlier studies in this region (for example, Drummond and others, 1993; Norris and others, 1996, 2000; Dettman and Lohmann, 2000). Realizing that there is a variety of potential causes for low $\delta^{18}\text{O}$ values of paleoprecipitation in this region (including cool local temperatures, a relatively large amount of low- $\delta^{18}\text{O}$ winter precipitation, high local relief, or the presence of an isotopic rain shadow), Sjostrom and others (2006) abstained from estimating absolute paleoelevations, but concluded that the results of their study imply high surface elevation of the Rocky Mountains since the Eocene.

The results of all the studies carried out in the western United States demonstrate that stable isotope data from authigenic minerals provide valuable information on paleotopography of this region. Integrating these results, it also becomes apparent that the interpretation of such data is most straightforward in regions characterized by simple conditions with respect to topography as well as atmospheric circulation and wind patterns. The most robust inferences on paleotopography can be made on the basis of the data from the leeward eastern side of the N–S trending Cascade Range. The Cascade Range is a relatively narrow mountain range oriented nearly perpendicular to predominantly westerly winds, and the effect of topography on the stable isotope composition of precipitation is much stronger than the likely effects of changes in climate or atmospheric circulation patterns in this region (Kohn and others, 2002; Takeuchi and Larson, 2005). Farther south and east in the Great Basin and Rocky Mountains, the interpretation of stable isotope data with respect to paleotopography is more problematic. As discussed in many of the studies summarized above, this is largely due to more complex atmospheric circulation and wind patterns arising from the interaction of near-surface winds with the complex topography of that vast region,

particularly in the context of the evolution of the North American monsoon. For example, higher surface elevation in the Great Basin region prior to the middle Miocene would, presumably, lead to relatively low surface pressure over this region in the summer, enhancing precipitation from air masses derived from the North, South, and East, while winter precipitation would be mostly westerly derived. A more monsoonal climate would also locally increase the importance of convection and, at least in the southern part of this region, lead to potentially significant variations in the importance of the amount effect and evaporation. However, the evolution of the North American monsoon is still poorly understood; accordingly, inferences on paleotopography based on data from the Great Basin and Rocky Mountains region have higher uncertainties than similar data from the Cascades region which is little affected by monsoonal climate variations.

New Zealand Southern Alps

In the present-day rain shadow on the eastern side of the Southern Alps of New Zealand, the $\delta^{18}\text{O}$ values of authigenic kaolinites in middle Miocene (~ 15 Ma) to Quaternary sediments show a 4 to 6 permil decrease in the late Miocene to early Pliocene (fig. 9D). This isotopic shift has been suggested to largely reflect surface uplift of the Southern Alps (Chamberlain and others, 1999). During the past ~ 15 Myr, influences such as mixing of isotopically distinct moisture sources, changes in atmospheric circulation patterns, and changes in latitude have not been important at the studied site (Ziegler and others, 1981; Stock and Molnar, 1982; Chamberlain and others, 1999). There has, however, been a significant influence of climate change during this time; since the 17 to 14 Ma mid-Miocene climatic optimum, the southwest Pacific area has undergone regional cooling of 5 to 7°C, with similar effects on sea surface temperatures and temperatures on land (Flower and Kennett, 1993, 1995; Shevenell and others, 2004). To evaluate the influence of this climate change on their results, Chamberlain and others (1999) considered the effect of temperature changes on land on precipitation ($+0.58\text{‰}/^\circ\text{C}$) and on the kaolinite/water fractionation factor (-0.28 to $-0.20\text{‰}/^\circ\text{C}$ for the 0°C to 30°C temperature range), as well as the observed ~ 1 permil increase in the oxygen isotope ratio of Southwest Pacific ocean water. From this they inferred that as much as 2 permil of the observed decrease may have been the effect of climate change. We note, however, that this ignores the significant influence of decreasing sea surface temperatures on the $\delta^{18}\text{O}$ ratio of precipitation ($-0.55\text{‰}/^\circ\text{C}$) due to the moisture content effect; including this effect implies that the net effect of regional cooling on the kaolinite $\delta^{18}\text{O}$ values would have been an ~ 2 permil increase, instead. Furthermore, this ignores the possibility that increased evaporation due to rain shadow formation east of the mountains may have led to increased $\delta^{18}\text{O}$ values of soil water. Thus the observed 4 to 6 permil decrease in the $\delta^{18}\text{O}$ values of the authigenic kaolinite samples may, in fact, underestimate the isotopic effect of rain shadow formation. Because of the predominance of coarse clastic deposits in Neogene sediments southeast of the Southern Alps, age control on the studied sediments is limited, and units containing authigenic kaolinite are scarce; as a result, the age and magnitude of the observed isotopic shift are relatively poorly constrained. However, it seems plausible that the isotopic effect of rain shadow formation was a 6 ± 2 permil decrease in the $\delta^{18}\text{O}$ values of precipitation. Assuming that the isotopic lapse rate at this location is $0.3 \pm 0.1\text{‰}/100$ m, the magnitude of late Miocene to early Pliocene surface uplift in the studied section of the Southern Alps can be estimated as $\sim 2 \pm 1$ km, consistent with the results of thermochronologic and sedimentologic studies, as well as geodynamic models (see discussion in Chamberlain and others, 1999).

Himalaya/Tibet Orogen

In this section we review stable isotope studies from (1) the foreland basin south of the Himalayas, (2) the Himalayas, (3) the Tibetan plateau, and (4) Cenozoic basins located north of the plateau. Not all of these studies present quantitative paleoelevation estimates; those that do not, however, do provide constraints on the evolution of the monsoon climate as indirect evidence for the presence of a large and highly elevated plateau.

The earliest stable isotope studies used to infer climate changes related to the evolution of the Tibetan plateau are from Neogene Siwalik Group molasse deposits in the Himalayan foreland of NW-Pakistan. The $\delta^{18}\text{O}$ values of pedogenic carbonate nodules and smectite show a 3 to 4 permil increase at ~ 8 Ma (Quade and others, 1989; Quade and Cerling, 1995; Stern and others, 1997). Similarly, the $\delta^{18}\text{O}$ values of mammal teeth and ratite eggshells show an increase around this time, although the scatter in these data is considerably higher; presumably, the shorter time scale at which such biominerals form causes their oxygen isotope compositions to reflect short-term variations in the oxygen isotope composition of precipitation and surface waters (Quade and others, 1992; Stern and others, 1994). This increase in the $\delta^{18}\text{O}$ values of authigenic minerals at ~ 8 Ma has been interpreted as the result of rapid uplift of the Himalayas and Tibetan plateau, leading to increased aridity and greater soil water evaporation due to monsoon strengthening (Quade and others, 1989, 1995), or to a change in the main moisture source area for the studied region by the rising orogen cutting off low- $\delta^{18}\text{O}$ precipitation from the North (Quade and others, 1995; Stern and others, 1997). The temporally invariant, yet higher than equilibrium, offset in the $\delta^{18}\text{O}$ values of pedogenic calcite and smectite suggests that aridity was highly seasonal (monsoonal) before 8 Ma (Stern and others, 1997), more consistent with the hypothesis of a change in moisture source due to orogenic uplift.

Farther east in the Himalayan foreland of Nepal, Dettman and others (2001) estimated the $\delta^{18}\text{O}$ values of paleoprecipitation from authigenic minerals contained in 13 Ma to recent fluvial and flood-plain deposits of Siwalik Group sediments. The mineral proxies they used include fossil freshwater bivalve shells consisting of pristine aragonitic material, as well as enamel from mammal teeth. The shell samples, taken from eleven shell beds dated as 10.7 Ma to 3.1 Ma, yielded $\delta^{18}\text{O}_{\text{PDB}}$ values ranging from -14.4 permil to $+6.3$ permil, and microsampling of individual shells revealed within-shell variabilities of up to ~ 16 permil. Assuming a broad range of reasonable paleotemperature estimates from floral assemblages, Dettman and others (2001) estimated that the shells formed in water with $\delta^{18}\text{O}$ values from -12 permil to $+7$ permil. This high variability has been interpreted as evidence for strong seasonal $\delta^{18}\text{O}$ variations in surface waters due to a pronounced monsoon climate, implying the presence of a high and broad Tibetan plateau by 10.7 Ma. The lowest values presumably reflect the $\delta^{18}\text{O}$ values of precipitation during wet seasons, and the highest values those of evaporated waters during dry seasons. Considering only the lowest $\delta^{18}\text{O}$ values, which are presumably from the samples least affected by evaporation, suggests that wet season precipitation had average $\delta^{18}\text{O}$ values of -9.6 permil before, and -6.5 permil after ~ 7.5 Ma. The tooth samples, taken from 5 stratigraphic levels with ages ranging from 13 Ma to Holocene, are characterized by a less marked variability of only 3 to 5 permil within each age group, but show an $\sim +5.8$ permil shift between ~ 9.5 Ma and ~ 3 Ma. Because the $\delta^{18}\text{O}$ ratio of teeth is strongly influenced by animal physiology and diet (for example, Kohn, 1996; Fricke and others, 1998), this is considered a less direct measure of the $\delta^{18}\text{O}$ of surface waters, but consistent with the data obtained from the freshwater shell samples. As the cause of the ~ 2 permil increase in the shells' $\delta^{18}\text{O}$ values at 7.5 Ma, Dettman and others (2001) suggested a change to a more arid

climate, leading to increased evaporation, as well as a less intense monsoon reducing the importance of the amount effect.

Within the Himalayan mountain ranges of west-central Nepal, Garzione and others (2000a, 2000b) used the isotopic composition of pedogenic, lacustrine, and biogenic carbonate (fig. 9E) in sediments of the Thakkhola graben as a proxy for paleoprecipitation and thus paleoelevation. For lacustrine micrites in the ~11 Ma to ~9.6 Ma Tetang Formation, they determined $\delta^{18}\text{O}_{\text{PDB}}$ values of -17.2 permil to -23.4 permil, with a mean of -19.4 permil. From pedogenic carbonate, as well as lacustrine micrites and fossil mollusk shells in the ≤ 8 Ma Thakkhola Formation, they obtained $\delta^{18}\text{O}_{\text{PDB}}$ values that fall, with one exception, in a very similar range of -15.9 permil to -22.4 permil, with a mean of -18.8 permil. Using known oxygen isotope fractionation factors between calcite and water, and ignoring the relatively small uncertainty introduced by changing mean annual temperatures in this region, they estimated the mean $\delta^{18}\text{O}$ values of meteoric waters as -21.7 permil during deposition of the Tetang Formation, and as -21.2 permil during deposition of the Thakkhola Formation. These values are similar to the values of modern precipitation in this region as inferred from surface waters in small local catchments (Garzione and others, 2000a, 2000b). Furthermore, applying a $-0.29\text{‰}/100$ m isotopic lapse rate to starting moisture with the -5.8‰ $\delta^{18}\text{O}$ value of modern precipitation at New Delhi (elevation 212 m) yields a paleoelevation of 5450 m for the Thakkhola samples, similar to the modern mean elevation in the catchment area of the sampled region. This is also consistent with the results of work by Rowley and others (2001), who modeled the changes in the isotopic composition of Himalayan precipitation with elevation as a Rayleigh distillation process. Accordingly, the results of these studies imply that the Himalayan Thakkhola graben region had achieved an elevation near the present-day value by ~11 Ma (Garzione and others, 2000a, 2000b; Rowley and others, 2001). We note, however, that according to our present understanding of the tectonic and topographic evolution of the Himalaya/Tibet orogen (for example, Hodges, 2000; Yin and Harrison, 2000) constraints on the paleotopography of the Himalayas have no clear implications for the elevation history of the Tibetan plateau.

Reliable stable isotope based paleoelevation estimates for the Tibetan plateau itself are, unfortunately, presently unavailable due to the lack of adequate data sets. However, limited data from the Oiyug Basin of southern Tibet suggest that the application of stable isotope paleoaltimetry in this region may, in principle, be promising. Currie and others (2005), studying 7 pedogenic calcite nodules from a 2-m-thick zone of alluvial paleosols in middle Miocene (>15.1 Ma) deposits (Spicer and others, 2003), determined $\delta^{18}\text{O}_{\text{PDB}}$ values in the range of -18.0 to -20.2 permil, with an average value of -19.6 permil. Assuming that these nodules formed at an estimated mean annual temperature of $6.8 \pm 3.4^\circ\text{C}$, inferred from paleobotanical work (Spicer and others, 2003), they estimated that these nodules formed in the presence of meteoric water with a $\delta^{18}\text{O}$ value of -21.7 ± 1.7 permil. This is similar to the $\delta^{18}\text{O}$ values of modern surface waters in the Oiyug Basin, which based on 6 streamwater samples and one sample from a hot spring, range from -18.6 to -21.1 permil (Currie and others, 2005). Furthermore, assuming that precipitation was derived exclusively from the South and crossing the Himalayas, they inferred a paleoelevation of $5,200 + 1370/-600$ m from these data, based on comparing the isotopic composition of their inferred Oiyug Basin paleoprecipitation with an estimated $\delta^{18}\text{O}$ value of -7.2 permil for roughly synchronous paleoprecipitation in the Himalayan foreland of NW Pakistan (from Cerling and Quade, 1993). They interpreted these data to imply high elevation of southern Tibet since at least 15 Myr ago, consistent with the results of tectonic and paleobotanical studies (Blisniuk and others, 2001; Spicer and others, 2003). It is worth noting, however, that today the high Himalaya mountains effectively block the north-

ward transport of moisture, and that the main source of moisture for the Tibetan plateau is the Pacific monsoon (Araguás-Araguás and others, 1998); accordingly, stable isotope data from the Tibetan plateau may largely reflect the evolution of air masses sourced from the Pacific Ocean and crossing the eastern plateau margin rather than the Himalayas.

In the Linxia Basin, located in the northeastern foreland of the Tibetan plateau (fig. 9F), Dettman and others (2003) studied the stable isotope composition of authigenic minerals in a nearly continuous sequence of fluvial and lacustrine deposits covering an age range from 29 Ma into the Holocene (Fang and others, 2003). The mineral proxies used in this study include carbonate cement from fluvial sandstone, micritic carbonate layers in the lacustrine sediments, and pedogenic carbonate from post-2.6 Ma loess deposits. The 29 to 20 Ma lower part of the studied section comprises fluvial sandstones with only a minimal content of detrital carbonate grains and abundant carbonate cement. The $\delta^{18}\text{O}_{\text{PDB}}$ values of these cements range from approximately -12 permil to -9 permil, and average ca. -10.5 permil. The <20 Ma upper part of the section is dominated by lacustrine deposits, including micritic carbonate layers which are primary precipitates from lake water on the basis of textural and mineralogical evidence. The $\delta^{18}\text{O}_{\text{PDB}}$ values of these lacustrine carbonates are characterized by a high-frequency 2 to 6 permil variation that was presumably caused by variable lake water evaporation, as inferred from a pronounced covariance of $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values. From 20 to 13 Ma, the $\delta^{18}\text{O}_{\text{PDB}}$ values range from about -11 permil to -5 permil; however, the lowest values, presumably from carbonate formed in the presence of the least evaporated lake waters, average approximately -10.5 permil, similar to the carbonate cements in the older 29 to 20 Ma fluvial interval. After ~ 12 Ma, the $\delta^{18}\text{O}_{\text{PDB}}$ values range from about -10 permil to -3 permil, with similar averages of approximately -9 permil for the values from fluvial carbonates and the lowest values from lacustrine carbonates. Because there is no evidence for a significant regional climate change at this time, the ~ 1.5 permil shift to higher $\delta^{18}\text{O}$ values at approximately 13 to 12 Ma has been interpreted to reflect a shift in the $\delta^{18}\text{O}$ value of precipitation; assuming a typical $25 \pm 4^\circ\text{C}$ temperature range for micrite precipitation, this implies a change in the $\delta^{18}\text{O}$ value of precipitation from -8.5 ± 1 permil before ~ 13 Ma to -7.0 ± 1 permil (similar to the -6 to -7‰ $\delta^{18}\text{O}$ value of modern precipitation) after ~ 12 Ma. As the most likely cause for this ~ 1.5 permil increase, Dettman and others (2003) suggest a 13 to 12 Ma increase in aridity resulting from uplift of the Tibetan plateau, or some portion of it, to elevations high enough to limit the supply of moisture from the Indian or the Pacific Ocean.

On a more regional scale, Graham and others (2005) presented data from four Cenozoic sections north of the Tibetan plateau (fig. 9G). The mineral proxies used in this study include calcite from paleosol carbonate, lacustrine limestone and mudstone, and carbonate cement from fluvial sandstone. The data resulting from this work document a 4 to 6 permil shift to higher $\delta^{18}\text{O}$ values of the analyzed calcite samples from Eocene to Oligocene time. This shift has been interpreted to reflect initial growth of the Himalaya/Tibet orogen and a resulting change to a warmer and more arid climate with increased summer precipitation, consistent with the results of general circulation models (Graham and others, 2005, and references therein). From Oligocene to Miocene time, there is no discernible change in the $\delta^{18}\text{O}$ values of authigenic minerals in the studied sections. From Miocene to Pliocene time, in contrast, there is 2 to 3 permil shift toward lower $\delta^{18}\text{O}$ values. As plausible causes for this decrease, Graham and others (2005) suggested (1) the retreat of the Neotethys from the Tarim Basin leading to the loss of a nearby moisture source, (2) the rise of the Tibetan plateau to a critical threshold elevation where it started to strongly affect atmospheric circulation, and (3) surface uplift of the Tian Shan creating an isotopic rain shadow to

the South and East of this mountain belt. This result contrasts strongly with that of Dettman and others (2003), who documented an increase in the $\delta^{18}\text{O}$ values of authigenic minerals in the Linxia Basin at approximately 13 to 12 Ma. However, this difference may be related to changes in the region studied by Graham and others (2005) involving the influence of moisture sourced north of Tibet, whereas the change in the Linxia Basin documented by Dettman and others (2003) may be more closely related to the influence of moisture sourced east and south of Tibet.

In summary, the presently available stable isotope data sets used to infer paleotopography in the region of the Himalaya/Tibet orogen show that the influence of climate on isotope data in monsoon systems is very complex. The main problems for reliable estimates of paleotopography in this region are the mixing of isotopically distinct air masses from different source regions resulting from the complicated near-surface wind patterns associated with the monsoon, and the difficulty in quantitatively assessing the effect of monsoon strength variations (through the amount effect) on the isotopic composition of precipitation. As discussed earlier in this paper, oxygen isotope data from modern precipitation and surface water samples in this region suggest that the former (mixing of air masses) is particularly troubling for sites located on the Tibetan plateau and in adjacent regions farther north, while the latter (monsoonal climate variability) is most important at the relatively low latitudes of the Himalayas and the Himalayan foreland. Furthermore, monsoon strength variations may have a profound influence on the importance of evaporation of the waters present during the formation of authigenic minerals. Given the relatively small number of data sets available for this vast and topographically complex region, as well as the limitations of our present understanding of the evolution of the monsoon climate, quantitative inferences on paleotopography in this region are associated with relatively large uncertainties.

UNCERTAINTIES OF PALEOELEVATION ESTIMATES

The studies summarized above demonstrate that oxygen isotope studies of authigenic minerals permit useful inferences on paleotopography, as long as potential complexities arising from global to regional-scale changes of temperature and/or climate patterns are carefully considered. Garzzone and Libarkin (2002) presented the first systematic error estimate for stable isotope paleoaltimetry, yielding total errors of ± 800 m at sea level and ± 1100 m at 3 km elevation for carbonate. Their estimate included: 1) an analytical uncertainty of ± 0.1 permil for $\delta^{18}\text{O}$ of carbonates; 2) the uncertainty in the carbonate/water fractionation factor; 3) uncertainties in the isotopic lapse rate of precipitation; 4) a ± 100 m error on estimating the average elevation of the paleodrainage basin; and 5) a $\pm 5^\circ\text{C}$ error in estimating local temperature. In the following paragraphs we review and expand upon these components of uncertainty.

(1) The most straightforward, and arguably the smallest error arises from the analytical uncertainties. Typical values for this (2σ) are ± 0.1 to 0.2 permil for $\delta^{18}\text{O}$ of carbonates, ± 0.2 to 0.4 permil for $\delta^{18}\text{O}$ of silicates, ± 2 to 3 permil for δD of silicates, and ± 0.3 to 0.5 permil for $\delta^{18}\text{O}$ of phosphates; these analytical uncertainties can be reduced by replicate measurements, but usually the variability in the isotopic composition of terrestrial minerals far exceeds the analytical uncertainties.

(2) The errors in mineral-water fractionation are also relatively small, resulting in an uncertainty of approximately ± 0.2 permil for $\delta^{18}\text{O}$ values of calcite (estimated from Kim and O'Neil, 1997), and similar to somewhat higher uncertainties for the other minerals discussed in this paper. However, any error in the isotopic fractionation is only relevant when estimating absolute paleoelevations; in the case of paleoaltimetry studies interpreting *changes* in mineral isotopic compositions with time, this error estimate is irrelevant because it applies equally to all samples.

(3) As discussed earlier in this paper, modern isotopic lapse rates vary over a fairly wide range from approximately -0.1 to $-0.5\text{‰}/100$ m in most low and middle latitude locations. However, isotopic lapse rates (and uncertainties of isotopic lapse rate estimates) can be much higher at extreme latitudes ($>70^\circ\text{S}$ or N), and generally increase with elevation (Poage and Chamberlain, 2001; Rowley and others, 2001). Isotopic lapse rates may be relatively low at low latitude sites because convection is generally more important at lower latitudes, and increased convection favors lower lapse rates (Ciais and Jouzel, 1994). However, in regions characterized by monsoonal climates, the predominance of the amount effect on the isotopic composition of precipitation may lead to increasing isotopic lapse rates as the result of higher precipitation rates (Gonfiantini and others, 2001). In many stable isotope based paleoaltimetry studies, modern isotopic lapse rates are used as proxies for isotopic lapse rates in the past; however, due to the potentially significant influence of climate variability, estimates of past isotopic lapse rates should be assigned a higher uncertainty than that of the modern isotopic lapse rate estimate at any given site. For sites located in middle latitudes and in topographically as well as climatically simple settings, it seems reasonable to assume isotopic lapse rates of approximately $0.3 \pm 0.1\text{‰}/100$ m; for sites at relatively low or high latitudes, lower or higher values, respectively, may be more appropriate, and for sites in monsoonal settings the uncertainty of isotopic lapse rate estimates may be higher. We note that it is also important to consider if a studied site may have been in an isotopic rain shadow. In this case, there is no systematic relationship between elevation and oxygen isotope values of precipitation on the leeward side of the mountains, and changes in the stable isotope values of authigenic minerals from that region will reflect elevation changes in the orographic barrier, not local elevation changes.

(4) In the case of authigenic minerals formed in the presence of river or lake waters, derived from a range of elevations, paleoelevation estimates critically depend on estimates for the average elevation of the paleocatchment area of the sampled site. Using data from the Thakkhola Graben region of Nepal as an example, Garzione and Libarkin (2002) suggested an uncertainty of ± 100 m for this estimate. However, in regions where topography has been strongly modified by erosion since the time of authigenic mineral formation, this uncertainty may be higher. Moreover, using the average elevation of the paleocatchment area to estimate the isotopic composition of paleoprecipitation ignores the possible influence of variations in precipitation rates with elevation. In the case of paleoaltimetry studies using pedogenic minerals, which generally reflect the isotopic composition of local precipitation, this source of uncertainty is negligible.

(5) The isotopic composition of authigenic minerals depends on local temperature due to the (opposite) effects of temperature on the isotopic composition of precipitation (for oxygen, about $0.58\text{‰}/^\circ\text{C}$ at mid-latitudes, see background) and on mineral/water fractionation factors (for oxygen, -0.19 to $-0.31\text{‰}/^\circ\text{C}$ in the temperature range of 0 to 30°C ; see background). Depending on the mineral proxy used, the net effect of local temperature changes on the $\delta^{18}\text{O}$ value of authigenic minerals can be estimated as approximately 0.27 to $0.39\text{‰}/^\circ\text{C}$, and the uncertainty of paleotemperature estimates will have a corresponding effect on paleoelevation estimates. Although the Cenozoic history of continental climates is not as well constrained as the marine climate record, it seems reasonable to assume an uncertainty of $\pm 5^\circ\text{C}$ for paleotemperature estimates (Garzione and Libarkin, 2002); if independent constraints on paleotemperatures are available, such as paleotemperature estimates based on studies of leaf physiognomy (for example, Wolfe, 1990, 1993, 1995; Spicer and others, 2003), uncertainties may be as low as ± 1 to 2°C .

(6) The isotopic composition of authigenic minerals is not only affected by local temperature changes, but also by global-scale temperature changes leading to ice volume changes; the accumulation of large amounts of snow and ice, which are generally characterized by very low $\delta^{18}\text{O}$ and δD values, leads to increasing $\delta^{18}\text{O}$ and δD values of ocean waters and vice versa. This effect resulted, for example, in an ~ 1.2 permil variability of the mean $\delta^{18}\text{O}$ value of ocean water during Quaternary glacial-interglacial cycles, and in an ~ 2 permil variability since the Oligocene. Because ocean water exerts a primary control on the isotopic composition of atmospheric moisture, this variability is reflected by the isotopic composition of precipitation and surface waters. However, the history of changes in the isotopic composition of ocean water is relatively well documented (for example, Zachos and others, 2001, and references therein), so that the resulting uncertainties are small (typically $<0.5\%$) after corresponding corrections have been made.

(7) A more problematic aspect of paleoelevation estimates is the strong effect of changes in sea surface temperatures, which control the starting moisture content of an air mass, on the isotopic composition of precipitation. With the exception of our work in the Patagonian Andes (Blisniuk and others, 2005) this effect, estimated as ca. -0.55% / $^{\circ}\text{C}$ for sea surface temperatures between 10 and 30°C (Aristarain and others, 1986; Jouzel and others, 1997), has been ignored in all the studies summarized above, but can be relatively important. For example, on the basis of oxygen isotope data from planktonic foraminifera it has been estimated that tropical sea surface temperatures have increased by almost 10°C since the early Oligocene (Wright, 2001), and a rapid 6 to 7°C decrease in sea surface temperatures during the middle Miocene has been documented at a paleolatitude of $\sim 55^{\circ}\text{S}$ in the South Pacific (Shevenell and others, 2004). Accordingly, we emphasize that the consideration of possible sea surface temperature variations during the time of interest will help to reduce the uncertainty of future stable isotope paleoelevation estimates.

(8) An additional uncertainty arises from quantifying the amount of evaporation which may have occurred to the water present during mineral formation. Evaporation can dramatically increase the $\delta^{18}\text{O}$ value of waters and thus the minerals forming in the presence of those waters. Lake waters are particularly susceptible to alteration by evaporation, and concern over evaporation should be heightened for rock sequences that contain evaporate minerals. For minerals that contain strongly-bound hydrogen and oxygen (chert, clays), one may use the combination of both of these isotopic systems to constrain the extent of evaporation. There is a linear relationship between $\delta^{18}\text{O}$ and δD values in precipitation ($\delta\text{D} = 8 * \delta^{18}\text{O} + 10$, the global meteoric water line; Craig, 1961), and thus minerals forming in the presence of non-evaporated waters will have a predictable δD value for any $\delta^{18}\text{O}$ value (a line parallel to the global meteoric water line with an offset specific for the phase and temperature; for example, Lawrence and Taylor, 1971). Deviations from this relationship indicate that this mineral did not form from pristine meteoric water. Samples that have undergone evaporation typically are more enriched in ^{18}O as compared to D, and plot below the meteoric water line. For carbonates, the presence or absence of significant evaporation can be inferred from the Mg/Ca and Sr/Ca ratios, with higher ratios indicating higher paleosalinity and thus more evaporitic conditions (Chivas and others, 1993), and covariant trends in $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ are also indicative of evaporation (Talbot and Kelts, 1990). Evaporation can also be constrained for by comparing data from contemporaneous mineral pairs such as pedogenic clays and calcite that form under differing moisture regimes (see for example, Stern and others, 1997; Tabor and Montañez, 2002, 2005). Additionally, independent evidence for the presence or absence of evaporation can be derived from floral and/or faunal records. These methods to infer the presence or absence of significant evaporation do not allow a quantitative assess-

ment of evaporative effects on the isotopic composition of authigenic minerals. However, in several of the studies summarized in this paper it has been observed that authigenic minerals from stratigraphic intervals in which Mg/Ca and Sr/Ca ratios of carbonates are high (Horton and others, 2004; Horton and Chamberlain, 2006), where $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values show pronounced covariance (Dettman and others, 2003), or where there is independent paleofaunal or -floral evidence for increased aridity (Blisniuk and others, 2005), are typically characterized by highly variable $\delta^{18}\text{O}$ values. Because evaporation invariably leads to increasing $\delta^{18}\text{O}$ values, it is generally assumed that the lowest $\delta^{18}\text{O}$ values from such intervals are the best proxy for the least evaporated waters, and only these values are considered for interpretations with respect to paleotopography.

(9) The effects of climate change on the isotopic composition of precipitation that result from changing atmospheric circulation and wind patterns are also difficult to assess quantitatively. However, in climatologically complex settings such as the central and northern Andes, the Great Basin and Rocky Mountains region, or in the Himalaya/Tibet orogen, it is important to consider the possibility of temporal changes in the isotopic composition of precipitation resulting from variable mixing of water vapor derived from isotopically distinct moisture sources. For example, Amundson and others (1996) inferred that Pleistocene precipitation in the Wind River Basin had $\delta^{18}\text{O}$ values > 4 permil higher than Holocene precipitation, opposing the pattern predicted from the known temperature increase from Pleistocene to Holocene times, but consistent with evidence for an increased importance of summer precipitation sourced from the Gulf of Mexico. Accordingly, they suggested a combination of weakened westerly winds and stronger southerly flow during the Pleistocene, presumably reflecting an increased northward extent of the North American monsoon.

(10) In low latitudes, particularly in areas with monsoonal climates and regions affected by the ENSO phenomenon, it is also important to consider the effect of climate change on precipitation rates and the amount effect. This effect is difficult to quantify, but clearly very important; for example, in New Delhi (elevation 212 m), summer precipitation (84% of the annual precipitation from June to September) is ~ 4.3 permil lower in $\delta^{18}\text{O}$ than winter precipitation (16% of the annual precipitation from October to May) despite an $\sim 8.7^\circ\text{C}$ higher summer temperature (Araguás-Araguás and others, 1998). The importance of this effect increases with increasing distance from the main moisture source and with increasing elevation; at Lhasa (elevation 3649 m), summer precipitation (90% of the annual precipitation from June to September) is ~ 6.3 permil lower in $\delta^{18}\text{O}$ than winter precipitation (10% of the annual precipitation from October to May) despite an $\sim 10.4^\circ\text{C}$ higher summer temperature (Araguás-Araguás and others, 1998). Considering that the temperature effect alone would lead to values that are $0.58\text{‰}/^\circ\text{C}$ higher in the summer (Rozanski and others, 1993), these data imply that precipitation rates exert the main control on the isotopic composition of precipitation in regions characterized by a monsoonal climate. The importance of variations in precipitation rates may, however, be estimated from constraints on seasonal to interannual fluctuations in the $\delta^{18}\text{O}$ values of precipitation derived from authigenic mineral proxies such as fossil shells or teeth.

DISCUSSION AND CONCLUSIONS

Because only analytical uncertainties apply equally to all stable isotope-based estimates of paleotopography, the largest components of uncertainty for each individual paleoelevation estimate vary from study to study. However, some general aspects and potential guidelines for future work emerging from the studies summarized in this paper are as follows:

(1) The effect of topography on the isotopic composition of precipitation is most straightforward in temperate mid-latitude regions and in topographically as well as

climatically simple settings. In temperate regions where mountains are oriented roughly perpendicular to the dominant wind direction, and water vapor is derived from a single dominant moisture source (for example, the Patagonian Andes, Sierra Nevada, Cascade Range, and Southern Alps of New Zealand), there is a systematic decrease in the $\delta^{18}\text{O}$ values of precipitation with elevation on the windward side of mountains, while “isotopic rain shadows” are observed on their leeward side. In other geographic and climatic regimes, altitude effects are observed on both sides of mountain ranges, but the relationship between altitude and the isotopic composition of precipitation on the leeward side of mountains is often complex, due to highly variable pseudo-altitude effects caused by below cloud-base evaporation and/or mixing of water vapor from isotopically distinct moisture sources. At low latitudes, and particularly at sites characterized by monsoonal climates, the predominance of the amount effect on the isotopic composition of precipitation can lead to strong variations in the altitude effect with interannual changes in precipitation rates.

(2) Correspondingly, stable isotope paleoaltimetry has relatively low uncertainties for regions where simple and stable climatic conditions existed and which were located near a single consistent moisture source, and where independent evidence for past climatic conditions is available. In climatically complex regions, particularly at low-latitude sites characterized by monsoonal climates, the uncertainties of stable isotope-based paleoelevation estimates can be relatively high due to potential long-term variations in surface wind patterns and in the importance of the amount effect on the isotopic composition of precipitation. However, in such settings stable isotope data can still provide useful information on paleotopography, provided there is a detailed understanding of paleoclimate.

(3) In any paleoclimatic setting, it is of fundamental importance to evaluate the presence and effect of local- to global-scale climatic change during the time span relevant for the isotopic change considered. Because climate changes are likely to have a stronger effect on the absolute $\delta^{18}\text{O}$ (or δD) values of precipitation than on isotopic lapse rates, interpreting changes in minerals’ isotope compositions with time as proxy for elevation *changes* is less prone to error than estimating absolute paleoelevations, as already emphasized by Poage and Chamberlain (2001).

(4) The difficulty to distinguish between the effects of mountain uplift and climate change on surface processes (Molnar and England, 1990), combined with the paucity of information on the paleoelevation of most mountain belts, significantly hampers a detailed understanding of the interactions between tectonic, climatic and earth surface processes. Oxygen and/or hydrogen isotope studies of authigenic minerals can, in appropriate settings, provide reliable constraints on paleotopography, and thus may contribute significantly to a better quantitative understanding of the links between these processes.

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