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EMPIRICAL RELATIONSHIPS BETWEEN ELEVATION AND THE STABLE ISOTOPE COMPOSITION OF PRECIPITATION AND SURFACE WATERS: CONSIDERATIONS FOR STUDIES OF PALEOELEVATION CHANGE

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ABSTRACT. A compilation of 68 studies from throughout many of the world's mountain belts reveals an empirically consistent and linear relationship between change in elevation and change in the isotopic composition of precipitation along altitudinal transects. The isotopic composition of precipitation decreases linearly with increasing elevation in most regions of the world except in the Himalayas and at elevations >5000 m. There are no significant differences in isotopic lapse rates from most regions of the world (~0.28 permil/100 m) except at the extreme latitudes where isotopic lapse rates are higher. Given information on past changes in the isotopic composition of precipitation preserved in pedogenic or authigenic minerals, this global isotopic lapse rate can be used to place numerical constraints on the topographic development of some ancient mountain belts or plateaus.

There are many complicating factors that can confound interpretation of paleoelevation change based on stable isotopes, and many of these are unique to specific mountain belts or time periods. Relevant to all stable isotope based paleoelevation change studies is the temperature dependent isotope fractionation between a pedogenic or authigenic mineral and the water from which it forms. In cases where isotopic proxy minerals are sampled from localities where temperature will change simultaneously with elevation change, the apparent change in the isotopic composition of precipitation may be dampened by several permil. This suggests that samples taken from the rainshadow side of an emerging orographic barrier may be more likely to preserve isotopic changes resulting from mountain uplift than samples taken from atop a rising mountain range or plateau.

INTRODUCTION

One of the major problems in understanding the relationship between topography and paleoclimate is the difficulty in reconstructing the elevational history of mountain ranges and plateaus. In recent years several methods have been developed that allow for quantitative estimation of the past altitude or elevational changes of mountain ranges. These methods have focused on such diverse topics as basalt vesicularity (Sahagian and Maus, 1994), cosmogenic nuclides (Brook and others, 1995), and paleobotany (Forest and others, 1999). In addition, recent research has focused on the use of stable isotopes of authigenic or pedogenic minerals to reveal important information about both the magnitude and timing of mountain range elevation change (Chamberlain and others, 1999; Chamberlain and Poage, 2000; Garzione and others, 2000a; Garzione and others, 2000b).

It is well known that precipitation and surface waters become increasingly depleted in ^{18}O and D with increasing altitude. This altitude effect (Dansgaard, 1964; Siegenthaler and Oeschger, 1980) has been recognized in almost all the major mountain belts of the world (table 1; fig. 1). This effect results principally from Rayleigh distillation and the depletion of ^{18}O and D in precipitation and vapor as an

TABLE 1

Compilation of studies considered in this paper

#	Location	Sample Type	Latitude (approx)	Minimum Elevation (m)	Maximum Elevation (m)	Net Elev. (m)	$\Delta\delta^{18}O$	#Data Points	r^2	Lapse Rate (‰/100m)	Reference
North America											
1	Olympic Mts., WA, USA	Snow, firn	47.5N	1399	2190	791	-4.1	6	0.89	-0.51	Sharp, Epstein, and Vidziunas, 1960
2	Yellowstone Nat'l Park, USA	Surface water	45N	1700	2700	1000	-2.5	32	0.55	-0.25	Thordsen and others, 1992
3	Mt. Steele, Canada	Snow	61N	2272	4917	2645	-9.2	6	0.89	-0.35	Holdsworth and Fogarasi, 1991
4	Mt Logan, Canada	Snow	60N	1819	5690	3871	-11.1	14	0.80	-0.29	Holdsworth and Fogarasi, 1991
5	Coast Mountains, Canada	Precipitation	50N	250	3250	3000	-7.5	NA	NA	-0.25	Clark and others, 1982
6	Sierra Nevada, CA	Snow cores	37-40N	1610	3490	1880	-5.5	33	0.27	-0.29	Friedman and Smith, 1970
7	Sierra Nevada, CA	Precipitation	39N	0	2200	2200	-6.8	2	NA	-0.31	Ingraham and Taylor, 1991
8	Sierra Nevada, CA	Precipitation	36N	0	3000	3000	-6.3	2	NA	-0.21	Ingraham and Taylor, 1991
9	SE California	Precipitation	33-37N	65	2540	2475	-3.7	32	0.45	-0.17	Friedman and others, 1992
10	Eastern California	Rain	37N	526	1275	749	-3.8	6	0.78	-0.50	Smith and others, 1979
11	Eastern California	Snow	37N	1182	3843	2661	-5.6	6	0.92	-0.22	Smith and others, 1979
Central and S. America											
12	Central Chile	Precipitation	33S	41	4330	4279	-13.5	10	0.95	-0.31	Stowhas and Moyano, 1993
13	Andes, Chile	Precipitation	34S	0	4186	4186	-14.4	7	0.84	-0.34	Moser and others, 1972
14	Northern Chile	Groundwater	20S	1410	4250	2840	-7.0	10	0.86	-0.25	Margaritz and others, 1989
15	Chile	Precipitation	40S	0	4291	4291	-15.8	7	0.86	-0.37	Rozanski and Araguas, 1995
16	Argentina	Surface water	38S	90	4009	3919	-15.3	23	0.88	-0.40	Vogel, Lerman, and Mook, 1975
17	Argentina	Rivers	30S	314	3999	3685	-12.9	11	0.95	-0.35	Rozanski and Araguas, 1995
18	Amazon Basin, Andes	Surface water	5S	0	5600	5600	-11.0	2	NA	-0.20	Grootes, 1993
19	Andes	Precipitation	32S	91	3057	2966	-6.7	4	0.98	-0.23	Rozanski and Araguas, 1995
20	Bolivia	Precipitation	16.3S	165	4071	3906	-10.7	2	NA	-0.27	IAEA data
21	Ecuador	Precip./Surface	33S	2	3034	3032	-6.3	8	0.91	-0.21	Rozanski and Araguas, 1995
22	Ecuador	Precipitation	3S	6	3150	3144	-8.5	11	0.75	-0.27	IAEA data
23	Ecuador	Precip./Surface	37S	200	4181	3981	-8.7	27	0.84	-0.22	Rozanski and Araguas, 1995
24	Nicaragua	Precipitation	13N	25	1700	1675	-3.1	7	0.80	-0.18	Payne and Yurtsever, 1975
Europe											
25	Sperkhios Valley, Greece	Groundwater	38.5N	180	2050	1870	-2.7	13	0.89	-0.15	Stahl, Aust and Dounas, 1974
26	Thassos Island, Greece	Surface water	41N	200	950	750	-1.5	6	0.99	-0.20	Leontiadis and others, 1996
27	Thrace, Greece	Surface water	41N	100	1490	1390	-2.9	11	0.98	-0.21	Leontiadis and others, 1996
28	E. Macedonia, Greece	Surface water	41N	600	1300	700	-2.8	9	0.95	-0.40	Leontiadis and others, 1996
29	Paikon Mts., N. Greece	Surface water	41N	27	1200	1173	-2.4	17	0.59	-0.20	Christodoulou and others, 1993
30	Austrian Alps	Precipitation	47N	580	2260	1680	-3.4	2	NA	-0.20	Ambach and others, 1968
31	Kitzsteinhorn, Austria	Snow	47N	1795	3159	1364	6.0	3	0.99	0.44	Moser and Stichler, 1974
32	Swiss Alps	Precipitation	47N	572	2150	1578	-4.2	6	0.99	-0.28	Siegenthaler and Oeschger, 1980
33	Maritime Alps, Italy	Rain	45N	400	2000	1600	-5.0	6	0.99	-0.32	Bortolami and others, 1978
34	Italian Alps	Precipitation	46N	2	2000	1998	-5.0	17	0.77	-0.25	Zuppi and Bortolami, 1982

| #	Location	Type	Lat								Reference
35	Italian Alps	Surface water	45N	400	2475	2075	-4.9	2	NA	-0.24	Dray and others, 1998
36	Mont Blanc	Snow	46N	2194	3914	1720	-7.7	7	0.96	-0.45	Moser and Stichler, 1974
37	Hasli Valley, Switzerland	Surface water	47N	625	2058	1433	-3.0	6	0.76	-0.21	Kullin and Schmassmann, 1990
38	W. Vosges, Switzerland	Surface water	47N	243	848	515	-1.7	12	0.52	-0.33	Kullin and Schmassmann, 1990
39	Central Jura Mts., Europe	Surface water	47N	495	1147	652	-1.2	8	0.99	-0.18	Kullin and Schmassmann, 1990
40	NE Jura Mts., Europe	Surface water	47N	269	1552	1283	-2.4	21	0.73	-0.19	Kullin and Schmassmann, 1990
41	Southern Jura Mts., Europe	Surface water	47N	440	985	545	-1.3	17	0.70	-0.25	Kullin and Schmassmann, 1990
42	Bernese Alps, Europe	Surface water	47N	1380	2161	781	-1.9	9	0.96	-0.24	Kullin and Schmassmann, 1990
43	Central Europe	Groundwater	52N	0	750	750	-0.8	7	0.70	-0.10	Richter, 1986
44	Germany	Groundwater	48N	200	900	700	-2.0	10	0.78	-0.28	Eichler, 1966
45	Czechoslovakia	Precipitation	50N	1030	1410	380	-1.4	2	NA	-0.36	Dincer and others, 1970
46	Canary Islands	Groundwater	28N	178	1900	1722	-2.0	8	0.87	-0.12	Gonfiantini and others, 1975
	Himalayas										
47	Kumuan Himalaya, India	Surface water	29N	915	2150	1235	-1.7	32	0.22	-0.14	Baratrya and others, 1995
48	Central Himalaya	Surface water	30N	1115	4290	3175	-4.5	14	0.27	-0.14	Bahadur, 1976
49	Xizang Plateau	Surface water	30N	220	5200	4980	-13.0	58	0.64	-0.25	Jinsheng and others, 1984
	Others										
50	Southern Alps, New Zealand	Surface water	43-44S	0	2000	2000	-3.7	8	0.78	-0.19	Chamberlain and others, 1999
51	Mt. Cameroon, Cameroon	Precipitation	4N	0	4005	4005	-6.8	32	0.97	-0.17	Fontes and Olivry, 1976
52	Hiei Mts., Japan	Surface water	35N	70	775	705	-0.9	7	0.73	-0.13	Kobayashi and others, 1997
53	Toyama Prefecture, Japan	Rivers	36N	260	1570	1310	-3.1	12	0.95	-0.24	Mizutani and Satake, 1997
54	Saiq Plateau, northern Oman	Precipitation		400	2000	1600	-3.2	NA	NA	-0.2	Stanger, 1986
	High Altitude Studies										
55	Cordillera Blanca, Peru	Snow	9S	4939	5259	320	-2.5	8	0.70	-0.77	Niewodniczanski and others, 1981
56	Cordillera Blanca, Peru	Snow	9S	4745	6173	1428	-5.0	15	0.39	-0.35	Niewodniczanski and others, 1981
57	Cordillera Blanca, Peru	Snow	16S	4796	6349	1553	-6.4	14	0.51	-0.41	Niewodniczanski and others, 1981
58	Hindu Kush, Afghanistan	Snow	36N	5098	7406	2308	-12.2	24	0.45	-0.53	Niewodniczanski and others, 1981
59	Hindu Kush, Afghanistan	Snow	36N	5199	6838	1639	-5.6	20	0.42	-0.34	Niewodniczanski and others, 1981
60	Central Hindu Kush, Afghan.	Snow	35N	4987	5865	878	8.2	14	0.54	0.93	Niewodniczanski and others, 1981
61	Himalaya, India	Snow	30N	4427	6077	1650	-0.7	15	0.17	-0.04	Niewodniczanski and others, 1981
	Extreme Latitude Studies										
62	NW Territories, Canada	Snow	75N	200	1870	1670	-10.5	33	0.84	-0.63	Koerner and Russell, 1979
63	Antarctica	Snow	70-75S	40	3010	2970	-29.6	48	0.95	-1.00	Lorius and Merlivat, 1975
64	Victoria Land, Antarctica	Firn	71S	2	2573	2571	-29.4	27	0.87	-1.14	Dansgaard and others, 1973
65	Byrd Land, Antarctica	Firn	85S	150	2195	2045	-6.0	7	0.73	-0.29	Dansgaard and others, 1973
66	East Antarctica	Firn	70S	2744	3717	973	-17.8	7	0.82	-1.83	Dansgaard and others, 1973
67	Mid Greenland, W. Slope	Firn	78N	1675	3168	1493	-9.3	5	0.99	-0.62	Dansgaard and others, 1973
68	NW Greenland	Firn	70N	1855	2700	845	-6.2	3	0.99	-0.72	Dansgaard and others, 1973
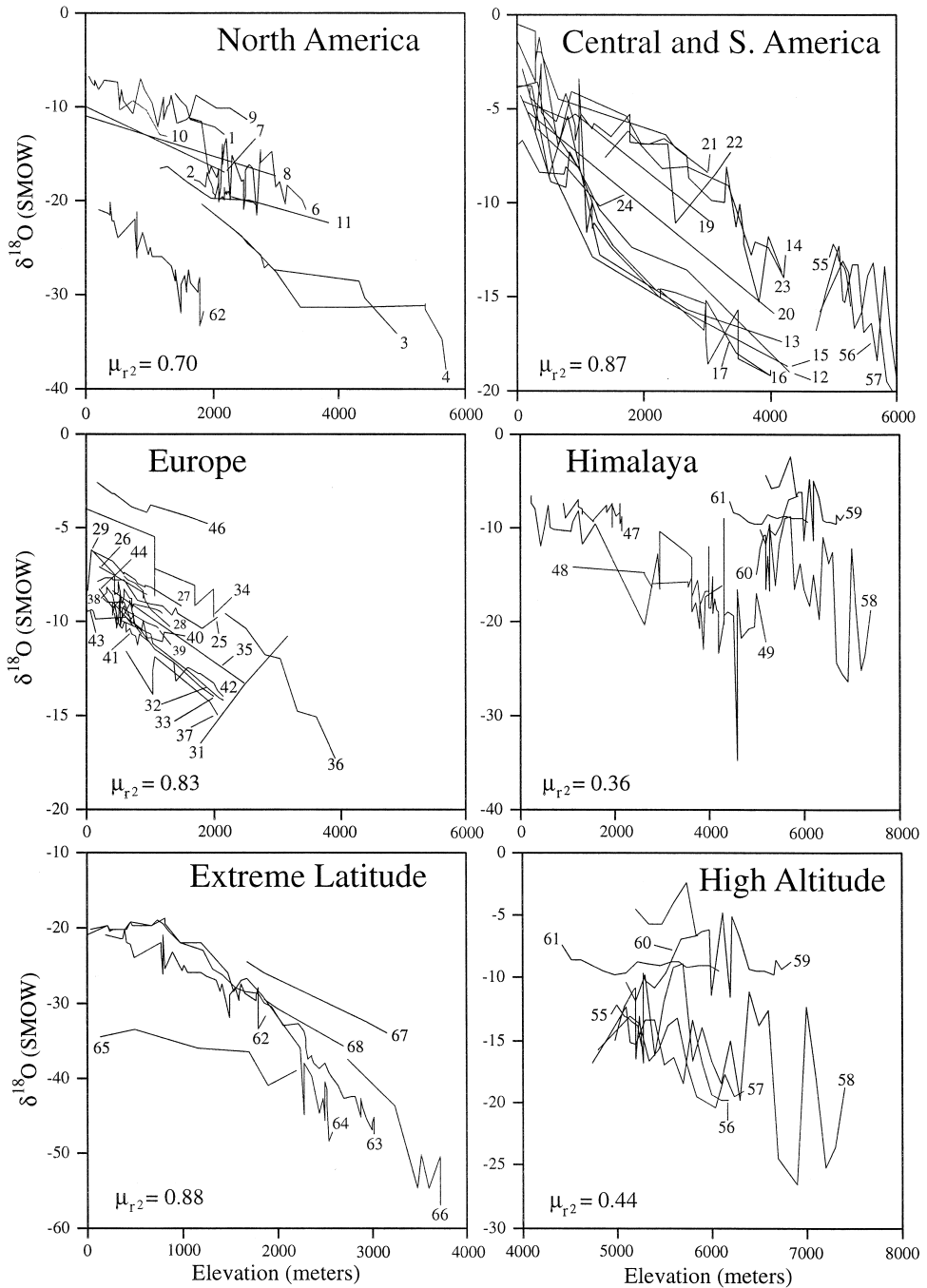
Fig. 1. Raw data for all studies used in this paper. Studies from North America Europe, Central and South America, and the Himalayas are grouped together. In addition, studies from extreme northern and southern latitudes (>70°) and studies from exclusively high altitudes (>~5000 m) are grouped together. Note that there is some overlap between groups as several studies fall into more than one category. The average r^2 value for each category is shown as μ_{r2} and provides a sense of the linearity of individual studies in each group.

air mass rises orographically and rains out moisture, though there are numerous other factors that can modify an air mass's isotopic distillation profile. As a result, in regions of a single dominant moisture source, precipitation and surface waters at high elevations or on the lee side of mountain ranges are often strongly depleted in ^{18}O and D as compared to waters on the windward side. The altitude effect is most often expressed as an isotopic lapse rate and given as a permil change in δ^{18}O or δD of precipitation per 100 m of elevation change.

This relationship can be and has been exploited to infer elevation changes of mountain ranges (Kolodny and Luz, 1991; Lawrence and Rashkes Meaux, 1993; Chamberlain and others, 1999; Dettman and others, 2000). In an earlier paper (Chamberlain and Poage, 2000), we showed that the change in δ^{18}O of authigenic minerals in chronosequences exposed within rainshadows of mountain ranges can in principle be used to constrain both the timing and the historical net elevation change of mountain ranges. Compiling δ^{18}O data from precipitation and surface waters throughout the globe, we showed that there is a near linear empirical relationship between the change in oxygen isotopes ($\Delta\delta^{18}$O) and change in elevation (figure 1 of Chamberlain and Poage, 2000). Using this empirical isotopic lapse rate and measuring the isotopic composition of near-surface formed minerals as a proxy for changes in the ancient isotopic composition of precipitation, we were able to place constraints on the topographic histories of the Southern Alps of New Zealand and the Sierra Nevada of California.

In this paper, we expand the empirical relationship between elevation and δ^{18}O of precipitation and surface water to include 68 data sets collected throughout the world and compare isotopic lapse rates from different geographic regions of the world, as well as at high altitudes and extreme latitudes. In addition, we consider the role of temperature changes in the interpretation of stable isotope data from different mineral proxies formed in either the rainshadow of an emerging mountain belt or atop an uplifting surface such as the Tibetan Plateau. Herein we show that: (1) isotopic lapse rates are fairly consistent across much of the world and are better constrained at elevations less than ~5000 m; (2) isotopic lapse rates are generally higher at latitudes greater than 70°; (3) with appropriate assumptions, changes in the stable isotopes of authigenic or pedogenic minerals from the geologic record can be used to semiquantitatively determine changes in mountain range elevation but are likely less useful for constraining absolute paleoelevations; and (4) samples for paleoelevation studies taken from atop an uplifting surface may be more difficult to interpret because of mean annual surface temperature changes occurring simultaneously with surface uplift.

EMPIRICAL RELATIONSHIP BETWEEN STABLE ISOTOPES IN WATERS AND ELEVATION

Ideally, it would be possible to combine all variables affecting the isotopic composition of precipitation into a numerical model characterizing modern δ^{18}O$_{precipitation}$ as a function of absolute elevation or elevation change. Given sufficient information about the ancient input parameters to such a model, past absolute elevations or changes in elevation of mountain ranges or plateaus might then be inferred from isotopic information preserved in the geologic record. However, our numerical understanding of all the processes and factors influencing δ^{18}O$_{precipitation}$ in the modern setting is limited, and many of these parameters were essentially unconstrained in the past. For example, the degree to which moisture is recycled by evapotranspiration is difficult to quantify in both the modern or ancient setting but may be very important in understanding changes in the isotopic composition of precipitation derived from an air mass moving across an orographic barrier (Ingraham and Craig, 1993).

An alternative approach in using stable isotopes to constrain changes in paleoelevation is to establish a modern global empirical relationship between $\delta^{18}O$ of precipitation (or waters derived directly from precipitation) and both absolute elevation and elevation differences and assume that the total variation in this relationship has remained constant through geologic time. With respect to interpreting paleoelevation changes, this assumption limits the need to characterize changes in ancient model parameters and allows for interpretation of isotopic information from the geologic record within the framework of modern $\delta^{18}O_{precipitation}$ versus elevation relationships. That said, however, it is highly unlikely that the physical processes governing isotope fractionation in a rising air mass in the past differs from that of today. What may and has changed in the past are global mean temperature and precipitation sources, which must be considered in any isotope paleoelevation study.

Numerous studies have characterized the relationship between elevation and the stable isotope composition of precipitation or meterically derived waters (see table 1). Most of these studies are from single mountain ranges and provide little information about global variations in this $\delta^{18}O_{precipitation}$ versus elevation relationship. To explore the global relationship between elevation and the $\delta^{18}O$ or δD of meterically derived waters, we compiled data from 68 published studies covering virtually all the major mountain ranges of the world (table 1; fig. 1). The intent was to amass as much information as possible about elevation-isotope relationships in the modern setting in order to aid interpretation of isotope data from the geologic record. We focused our compilation on studies of meteoric waters along altitudinal transects though we included studies sampling surface waters or shallow groundwater which are derived from precipitation. We avoided using data from the International Atomic Energy Agency-Global Network of Isotopes in Precipitation (IAEA-GNIP). Because IAEA-GNIP samples are taken over a wide geographic region, comparison of stable isotope versus elevation relationships between sites is complicated by numerous factors such as the continentality effect (Dansgaard, 1964), variation of absolute isotopic values with latitude, and differences in the source of precipitation.

The $\delta^{18}O$ values of precipitation/surface waters and corresponding altitude for the 68 transects are shown in figure 1. Included in table 1 are the types of samples (rain, snow, firn, surface water, groundwater), approximate latitude, elevation range of the transect, the change in $\delta^{18}O$ ($\Delta\delta^{18}O$) along the regression line between the minimum and maximum elevation, and the calculated isotopic lapse rate from the $\Delta\delta^{18}O$ and net elevation. In studies where only δD data were presented, we converted δD to $\delta^{18}O$ using the expression for the global meteoric water line: $\delta D = 8\delta^{18}O + 10$ (Craig, 1961). To assess geographic similarities and differences in the relationship between $\delta^{18}O$ values of surface waters/precipitation, we grouped these studies by geographic region (North America, Central and South America, Europe, and the Himalaya), extreme latitude ($>70°$ either north or south), and high altitude (>5000 m). Studies shown in figure 1 are cross-referenced to table 1 by the assigned number.

As expected, in all but two studies (#31 and #60) there is a systematic decrease in the $\delta^{18}O$ of meterically derived waters with increasing elevation. Of particular significance in the application of stable isotopes to paleoelevation studies is the fact that there is wide variability in the $\delta^{18}O$ of precipitation or surface waters at a particular elevation. Similarly, there is a wide range of elevations at which precipitation of a particular $\delta^{18}O$ can fall. Some of this variation is likely due to seasonal variations of sampling for particular studies and the fact that individual studies do not necessarily represent annually averaged isotopic compositions for precipitation. Nevertheless, it does suggest that our limited ability to empirically constrain the modern isotopic composition of precipitation at a specific *absolute* elevation itself limits the practicality of using stable isotopes as an indicator of *absolute* paleoelevation.

In most individual studies this observed δ^{18}O versus elevation relationship is close to linear over the sampled elevation range. The r^2 value for each study is reported in table 1, and to provide a sense of how consistently linear the δ^{18}O versus elevation relationship is in individual studies, the mean r^2 value for each grouping is shown as μ_{r2} on figure 1. Only in rare individual cases does a polynomial fit produce a significantly better r^2 than simple linear regression. Note in the cases of North America and Central and South America, studies that also fell into either the High Altitude or Extreme Latitude grouping were not used in the calculation of μ_{r2}. In all groupings except the Himalayan and High Altitude groups (between which there is significant overlap), the average r^2 value is 0.70 or higher. The fact that this δ^{18}O versus elevation relationship appears linear over much of the world indicates that the combination of processes dominantly responsible for isotopic distillation during rainout behave linearly over much of the world, at least within the precision of the studies we considered.

In both the Himalaya and High Altitude (>5000 m) groupings there is significantly more scatter about individual regression lines indicating an increased complexity of modern precipitation patterns in both the High Himalaya and Peruvian Andes. We are unsure what causes the wide variability in δ^{18}O values at high elevations, however it has been suggested that this variation may be due to post-depositional changes in snow (Niewodniczanski and others, 1981) or even the addition of a secondary source of moisture from the upper troposphere (Holdsworth and Fogarasi, 1991).

Because the modern δ^{18}O versus elevation relationship in most studies is close to linear, isotopic lapse rates can be calculated and compared for different parts of the world. Two methods for calculating average lapse rates yield results within 0.04permil/ 100 m of each other and involve either (1) taking the arithmetic mean of the lapse rates from all studies in a given region or, (2) plotting $\Delta\delta^{18}$O versus elevation difference for each study and using the slope of the regression line forced through the origin as a regional isotopic lapse rate as shown in figure 2. The slopes of the regression lines are −0.27, −0.30, and −0.21permil/100 m for the North America, Central and South America, and Europe groupings respectively. The averaged lapse rates are statistically indistinguishable for the North America, Central and South America, and Europe groupings: −0.31, −0.27permil/100 m, and −0.24permil/100 m, respectively. Note that studies shown in figure 1 that fall into the High Altitude or Extreme Latitude group were not used in these calculations.

A similar plot for the Extreme Latitude shows significantly greater lapse rates of −1.14 and −0.64permil/100 m for extreme north and south latitudes respectively. Studies from the Himalaya have a similar average lapse rate of −0.24permil/100 m, though the generally poor r^2 (μ_{r2}=0.36) values of individual studies limit the usefulness and validity of this statistic. The High Altitude grouping (including Himalayan studies) has a higher mean isotopic lapse rate of −0.41permil/100 m, though again, low r^2 values of individual studies limit the usefulness of this number. Statistically, there is no significant difference between the averaged lapse rates of any pair of groupings except the Extreme Latitude group which, at 95 percent confidence, is significantly higher than all other groupings.

Considering only studies from North America, Central and South America, and Europe together with other low altitude studies ("Others" in table 1) that did not fit into any specific geographic grouping, an estimate of an isotopic lapse rate that has more global implications can be made. These are plotted as the Compilation dataset in figure 2 and show a near linear relationship between the net elevation of each study and the change in the isotopic composition of waters sampled in that study. The slope of the regression line (r^2=0.80) through these points suggests a "global isotopic lapse
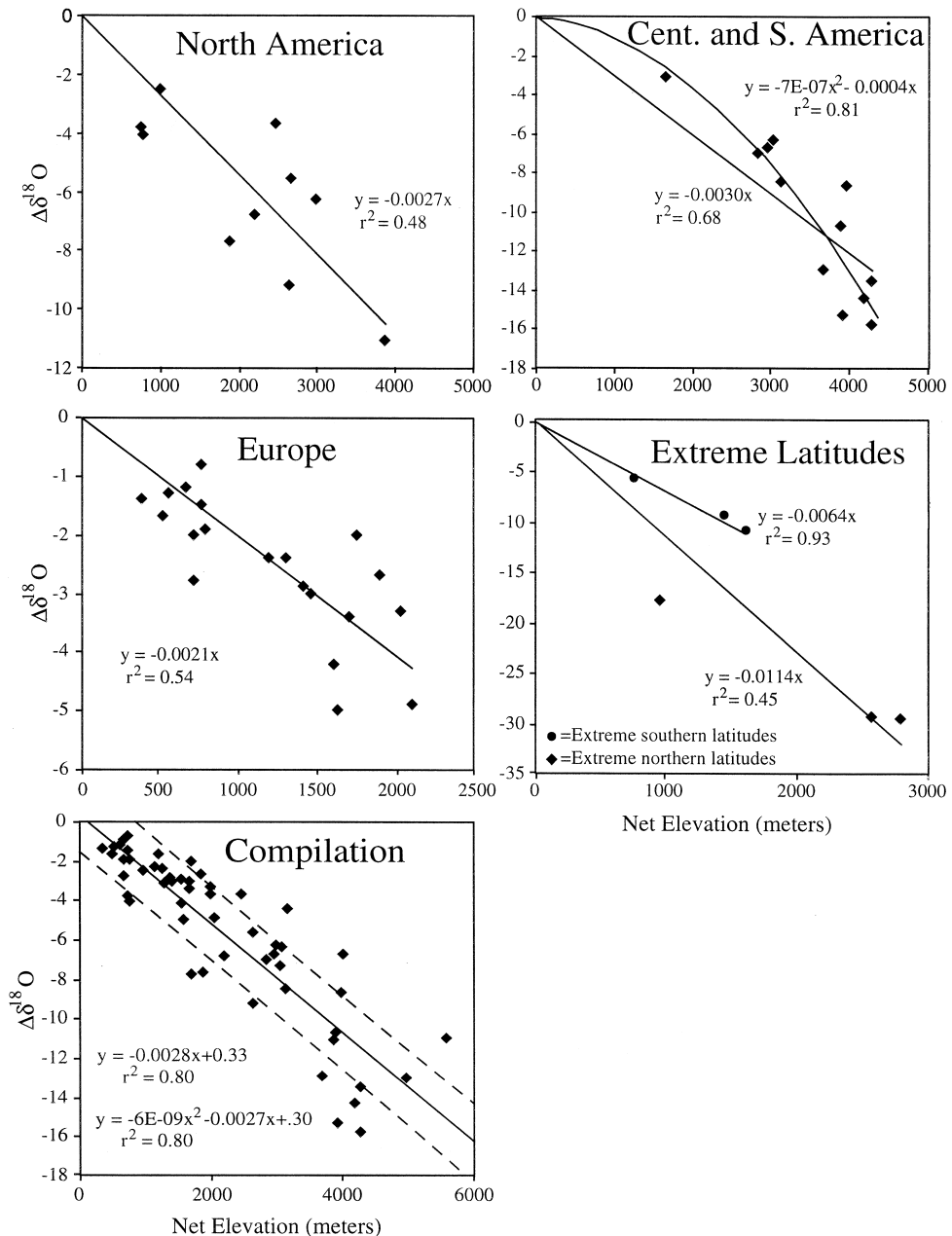
Fig. 2. Plots of the change in $\delta^{18}O$ of precipitation versus net elevation change. The North America, Central and South America, and Europe categories are the same as in figure 1. High altitude studies (#'s 55,56,57) are excluded from the Central and South America plot due to low r^2 values. We include no plot for the Himalayas due to poor r^2 values. The Extreme Latitude grouping of figure 1 is broken into northern and southern groupings. Also included is a Compilation plot showing all the data from North America, Central and South America, Europe, as well as studies 50 to 54. The best fit regression line shown on each graph is forced through the origin except for the Compilation plot. Also shown on the Central and South America category is the best fit polynomial curve. The regression on the Compilation plot is not forced through the origin for purposes of calculating the predictive error; however, this does not change the slope of the regression line significantly. The calculated 1 standard error envelope is shown on the Compilation plot. The best fit polynomial curve for the Compilation plot is indistinguishable from the best linear fit.

rate" of -0.28 permil/100 m which is -0.07 permil different from our previous estimate (-0.21 permil/100 m; Chamberlain and Poage, 2000). Note that the best second-order polynomial fit shown in figure 2 has an approximately equal r^2 value, and thus for practical purposes, the global empirical relationship between $\Delta\delta^{18}O_{precipitation}$ and net elevation change below $\sim$5000 m can be modeled as essentially linear.

This relationship provides a framework for interpreting changes in the isotopic composition of proxy minerals from the geologic record in terms of net elevation change of a mountain range and consequent increased isotopic distillation of air masses forced over that range. Assuming an accurate assessment of historical changes in $\delta^{18}O_{precipitation}$ associated with changing rainout conditions, using this model to estimate mountain or plateau surface uplift requires an estimation of the predictive precision of this linear regression. The Compilation plot of figure 2 shows the 1 standard error envelope calculated by conventional error approximation methods for an individual measurement of $\Delta\delta^{18}O_{precipitation}$ (Miller and Miller, 1993, p. 113). These standard error curves suggests that despite a good statistical correlation between $\Delta\delta^{18}O_{precipitation}$ and elevation in the modern environment, the statistical usefulness of this correlation for paleoelevation change studies is restricted to mountain belts with historically large elevation changes.

We emphasize that this model is strictly empirical and as such assimilates most of the modern global variation in processes influencing the relationship between the isotopic composition of precipitation and elevation change. That (1) the relationship between the $\delta^{18}O$ of meteorically derived waters and elevation is approximately linear in most studies, and (2) most isotopic lapse rates from elevations below 5000 m fall within a narrow range warrants some comparison to simple Rayleigh distillation models (Dansgaard, 1964; Pierrehumbert, 1999).

At present there are few numerical models that describe the relationship between altitude and the isotopic composition of precipitation. Dansgaard (1964) provides the simplest isotopic rainout model of an air mass experiencing moist adiabatic cooling. This model is based on temperature controlled Rayleigh distillation and is approximately analogous to an air mass rising over an orographic barrier. Though modeled isotopic distillation curves ($\Delta\delta^{18}O_{precipitation}$ per °C cooling) are not mathematically linear, over temperature changes of <20°C, they are very close to linear. For example, air masses with initial temperatures of 0° and 20°C will average a change in the isotopic composition of precipitation of 0.64 and 0.33 permil/°C, respectively, over 20°C of cooling (Dansgaard, 1964). However, translating this into a change in $\delta^{18}O_{precipitation}$ with change in elevation is complicated by (1) the strong dependence of isotopic distillation rate on initial air mass temperature, (2) the fact that moist adiabatic lapse rates governing the temperature change of rising saturated air mass are dependent on both initial temperature and absolute elevation (Ahrens, 1994), and (3) the fact that moist adiabatic lapse rates are not mathematically linear. Using approximate moist adiabatic lapse rates of -7.5 and -5.5°C/1000 m for air masses with initial temperatures of 0 and 20°C respectively (Ahrens, 1994) will result in calculated isotopic lapse rates of -0.48 and -0.18 permil/100 m for elevations below 4000 m. These isotopic lapse rates are similar to those determined in a Rayleigh model specific for the Andes (Pierrehumbert, 1999). Calculated average isotopic lapse rates up to $\sim$4000 m varied from -0.45 permil/100 m for an air mass with an initial temperature of 7°C to -0.13 permil/100 m for an initial temperature of 25°C (Pierrehumbert, 1999). Above this altitude the isotopic lapse rates increased dramatically.

These numerical results generally bracket the majority of observed isotopic lapse rates from the individual studies as well as the compiled global isotopic lapse rate. It is unknown why isotopic lapse rates should fall into such a narrow range over most of the world's mountain belts but likely reflects the dominance of temperature/elevation

controlled isotopic distillation in controlling $\delta^{18}O_{precipitation}$ changes. Alternatively, it may reflect a combination of other factors that systematically modify the effects of temperature-controlled Rayleigh distillation. We also suggest that the observed linear relationship between the isotopic composition of meteorically derived waters and elevation in most studies may in part be explained by the approximate though not mathematical linearity in both moist adiabatic lapse rates, and the $\delta^{18}O_{precipitation}$ versus temperature relationship in an ideal air mass undergoing moist adiabatic cooling.

<div align="center">SAMPLING LOCATION AND MINERALOGIC CONSIDERATIONS</div>

Above, we assess the nature of the relationship between absolute elevation, net elevation differences, and changes in the isotopic composition of precipitation. From this we suggest that changes in the isotopic composition of precipitation through time as recorded in the geologic record are more suited to assessing changes in elevation rather than absolute elevation of ancient mountain belts. Developing an understanding of a mountain range's ancient history, however, requires generation and preservation of a measureable proxy from which to assess changes in ancient precipitation. Numerous studies have used changes in the isotopic composition of minerals formed in the soil or near surface environment and preserved in the sedimentary record as a proxy for historic changes in the isotopic composition of precipitation, soil water, or groundwater. Commonly used minerals formed in the soil or near surface environment include calcite (Cerling and Quade, 1993), smectite (Stern and others, 1997), goethite (Yapp, 1993; Poage and others, 2000), and kaolinite (Lawrence and Rashkes Meaux, 1993; Chamberlain and others, 1999). The use of mineral proxies is not without complications, however, as many other factors in addition to changes in $\delta^{18}O_{precipitation}$ can influence the isotopic composition of these minerals. Many of these confounding factors are difficult to isolate; for example, changes in seasonal moisture sources may easily be mistaken for a change in the isotopic composition of a single moisture source. In addition, understanding the timing of mineral formation with respect to the timing of mountain building is of utmost importance in these studies. If authigenic minerals form on the order of 10^5 yrs than they should record information on large-scale tectonic processes that take place on the order of 10^6 yrs. Although no one has quantitatively determined the rate of authigenic mineral formation in the shallow subsurface, numerous empirical studies show that authigenic and pedogenic minerals form on time scales sufficiently short to record tectonic processes (Cerling and Quade, 1993; Chamberlain and others, 1999).

The isotopic composition of pedogenic or authigenic minerals depends upon the isotopic composition of the water from which it forms as well as the temperature at which it forms. In this sense it is very important to consider that surface temperature change may be coupled with elevation change and that this surface temperature change may modify the isotopic record produced by changing $\delta^{18}O_{precipitation}$ (Savin and Hsieh, 1998). As mineral-water fractionation factors depend upon absolute temperature and do not vary linearly with temperature, assessing the isotopic effect of changing mineral formation temperatures, in principle, requires *a priori* knowledge of absolute temperatures. This problem can be circumvented by assuming that mineral-water fractionation factors vary linearly across a reasonable range of earth surface temperatures, allowing for estimation of changes in mineral-water fractionation based on temperature changes rather than absolute temperatures. Linearization of mineral-water fractionation for the common pedogenic minerals calcite, smectite, kaolinite, and goethite over the temperature range from 0° to 30°C yields slopes of approx −0.233, −0.217, −0.163, and −0.137 permil/°C respectively (figure 3A). This simplification limits the need to have independent knowledge of absolute mineral formation
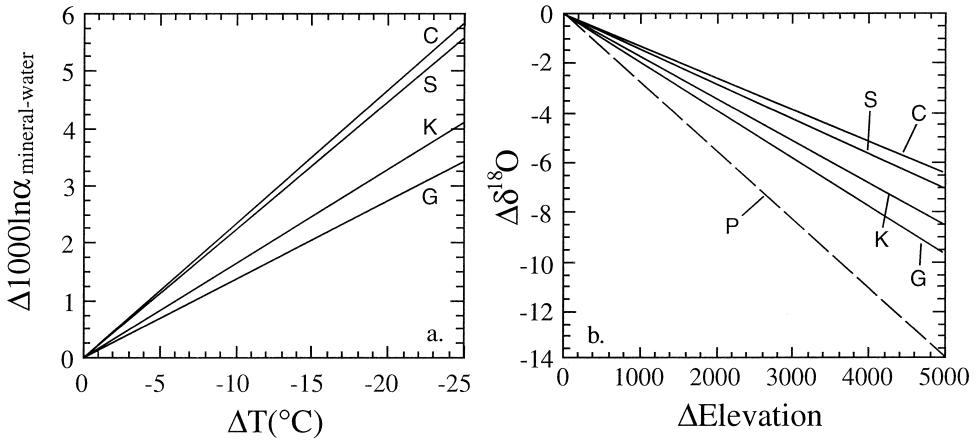
Fig. 3(A) shows approximate changes in the $1000\ln\alpha_{mineral-water}$ with change in temperature for calcite (Friedman and O'Neil, 1977), smectite (Savin and Lee, 1988), kaolinite (Savin and Lee, 1988), and goethite (Yapp, 1990). Fractionation factors for individual minerals were linearized over a reasonable range of earth surface temperatures to limit the need to know absolute temperature differences. Lines in B show the calculated change in $\delta^{18}O$ of a particular mineral with change in elevation for samples taken on an uplifting surface and taking into consideration 6.5°C change in mineral formation temperature per kilometer of surface uplift. For comparison, the dotted line shows the expected change in $\delta^{18}O$ of precipitation associated with elevation change. C=calcite, S=smectite, K=kaolinite, G=goethite, P=precipitation.

temperatures and introduces minimal error of $<\sim0.2$ permil in assessing the effects of temperature changes on the isotopic composition of near surface formed minerals.

Mineral-water fractionations for the various proxy minerals respond to temperature changes differently, suggesting that some minerals may be more useful than others in resolving the isotopic effects of elevation changes. For example, goethite and kaolinite show a lesser degree of sensitivity to temperature changes than calcite and smectite (fig. 3B). Therefore, in cases where changes in mineral formation temperature must be considered goethite and kaolinite will more accurately record isotopic changes in precipitation.

Consideration of coupled elevation-temperature changes is particularly relevant in the case where samples are taken from the top of an uplifted mountain range or plateau where the temperature at which proxy minerals formed is expected to have changed with absolute elevation changes through the period of surficial uplift. Assuming that a change in mean annual temperature associated with elevation change translates into a similar magnitude change in temperature of mineral formation, the isotopic change recorded by a particular proxy mineral will be influenced by: (1) the magnitude of elevation change of the mountain range or plateau, (2) the temperature lapse rate with respect to elevation, and (3) changes in mineral-water fractionation factor brought on by the temperature change.

Figure 3B shows a plot of the predicted $\Delta\delta^{18}O_{mineral}$ versus elevation change for hypothetical samples taken on top of an uplifting surface factoring in an estimated temperature lapse rate of -6.5°C/1000 m (Ahrens, 1994). We recognize that temperature lapse rates are likely to vary globally, but for purposes of illustration we have chosen to use the standard tropospheric lapse rate of -6.5°C/1000 m. In this calculation we also assume that (1) surface temperature change translates directly into an equal change in the temperature of mineral formation, and (2) no other variable other than mean annual temperature and changes in $\delta^{18}O_{precipitation}$ due to isotopic distillation affect the $\delta^{18}O$ of the proxy mineral. For reference, the dashed line in figure 3B shows the estimated isotopic depletion of precipitation resulting from

altitude induced rainout. Figure 3B shows a significant dampening of the expected change in the isotopic composition of the common proxy minerals as compared to the expected change in the isotopic composition of precipitation. In this case kaolinite and goethite preserve more of the expected change in $\delta^{18}O_{precipitation}$ as compared to smectite and calcite. In all cases, neglecting the temperature change when interpreting changes in $\delta^{18}O$ of a proxy mineral will result in underestimation of surface uplift.

The isotopic effects of altitude induced temperature changes can be substantial, suggesting that the isotopic composition of samples taken in the rainshadow of a developing mountain range may be a more sensitive indicator of paleoelevation change than the isotopic composition of minerals taken from atop the uplifting surface. Samples from the rainshadow side of a mountain range may not be directly affected by temperature changes associated with elevation change; however, they are subject to other influences that can confound interpretation. As a rainshadow develops, increased evaporation may shift the isotopic composition of soil or groundwater to higher $\delta^{18}O$ values which may also result in underestimation of elevation change. Similarly, the emerging rainshadow may result in a mean annual surface temperature change which in turn may further complicate interpretation. These effects are difficult to assess quantitatively and must be considered on a case by case basis.

There are clearly many other factors that influence the $\delta^{18}O$ of pedogenic or authigenic minerals regardless of sample location. In this sense, calculations presented here cannot accurately represent the complexities of the natural system. Many of these factors are specific to either particular mountain belts or particular periods in geologic history and must be considered as individual cases. For example, changes in precipitation source or shifts in regional and local climate resulting from mountain or plateau uplift, global temperature changes, or changes in the isotopic composition of the world ocean can all influence the $\delta^{18}O$ of precipitation and surface waters and thus the $\delta^{18}O$ values of authigenic minerals. We emphasize here the role of changes in mineral formation temperature associated with samples taken from a rising surface because of its likelihood to significantly affect the apparent change in the isotopic composition of precipitation and complicate interpretation of paleoelevation changes. This suggests that the geographic position of the sampling site with respect to a rising mountain belt is an important consideration and can exert some additional interpretive difficulties.

<div align="center">CONCLUSIONS</div>

Analysis of the relationship between $\delta^{18}O_{precipitation}$ and elevation from 68 altitudinal transects reveals consistently linear decreases in $\delta^{18}O_{precipitation}$ with increasing elevation throughout much of the world. The exceptions to this are studies at elevations greater than 5000 m and studies from the Himalayas, both of which show significant departure from linearity. Despite the consistent linearity of this relationship within a given study, there is little correlation between $\delta^{18}O_{precipitation}$ and absolute elevation, limiting the usefulness of stable isotopes of pedogenic or authigenic minerals in studies of absolute paleoelevation. The slopes of the regression lines for each study define a local isotopic lapse rate. These isotopic lapse rates fall within a narrow range for many regions of the world, with the exception of the extreme latitudes where the magnitude of $\delta^{18}O_{precipitation}$ change with elevation change is significantly higher. Collectively, these local isotopic lapse rates define a global isotopic lapse rate that can be used to semiquantitatively estimate changes in paleoelevation of a mountain range or plateau given reasonable estimation of changes in $\delta^{18}O_{precipitation}$ from a suitable isotopic proxy mineral. Quantifying historical elevation changes of mountain ranges based on changes in $\delta^{18}O_{precipitation}$ also depends upon changes in the temperature at which the isotopic proxy mineral forms. As such, sample location and the mineral specific dependence of isotope fractionation on temperature must be

considered when interpreting paleoelevation changes based on stable isotopes of near-surface formed minerals.

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