SUPPLEMENTAL INFORMATION

Discussion of Isotopic Effects

Many factors can affect the stable isotopic composition of mineral proxies. The following is a discussion of these effects and a rationale for why we interpret the variation seen in the isotopic record of the Cenozoic North American Cordillera as elevation change and accompanying changes in seasonality of precipitation.

Latitude and Continentality Effects: Precipitation at higher latitudes becomes progressively more depleted in [18]O and D (Craig, 1961). In the eastern continental U.S., this occurs at a gradient of ~0.4 ‰/km for oxygen isotopes (Kendall and Coplen, 2001). In the western U.S., however, isotopic values appear to be dominantly controlled by elevation and air masses. On the west coast of North America, for example, $\delta^{18}O$ values in southern Washington and southern California are both -6 to -8 ‰. Furthermore, western North America has remained at approximately the same latitude in the Cenozoic (Smith et al., 1981) so it is unlikely that any of the major findings here can be attributed to changes in the latitude with time.

Isotopic composition is affected by inland distance because an air mass will become progressively depleted as it rains out. This is the same effect produced by an orographic barrier and the composition of precipitation will follow a Rayleigh fractionation curve. There has been a change in continentality as the western U.S. has undergone significant extension since the Eocene. However, even a change of 400 km (accounting for 100 % extension from northwestern Utah to the western U.S. coast) would only produce a ~0.8 ‰ increase in $\delta^{18}O_p$ prior to extension assuming 0.002 ‰/km (Criss, 1999). This isotopic increase is not of the magnitude to affect any of the conclusions presented in this work.

Temperature Effects: Temperature does affect mineral-water fractionation factors. Fractionation is usually greater at colder temperatures. This effect has already been accounted for in the conversion from mineral proxy values to paleoprecipitation values. In these calculations, we used temperature estimates from paleofloral indicators and interpolated between known values. While the temperature estimates may be rough, the temperature effect is small compared to the magnitude of the variation observed in the isotopic record. For example, in the calcite-water oxygen isotope system, a misestimate of 15 °C instead of 25 °C MAT would only result in ~2 ‰ of change in $\delta^{18}O_p$. Temperature can also affect global ice volume, which in turn affects isotopic composition of the oceans because ice sequestered at the poles is relatively depleted in [18]O and D. Therefore, modern oceans would be relatively [18]O- and D-rich compared to the ice-free Eocene. Lack of ice caps in the Eocene would produce lower $\delta^{18}O_p$ of the initial air mass and therefore the proxy isotopic composition. However, since global temperatures were higher in the Eocene, the magnitude of fractionation would decrease, leading to mineral proxies with higher $\delta^{18}O$ values. The combined competing effects of temperature and

ice volume only changed marine carbonate composition by ~0.8-1.0 ‰ in oxygen (Zachos et al., 2001)

Evaporative Effects: Evaporation can significantly affect the composition of samples by enriching surface waters in ^{18}O and D. This effect can produce isotopic shifts on the order of major elevation changes, so precautions need to be taken in order to eliminate changes in evaporation from being misinterpreted as elevation changes. The most obvious evidence for evaporative alteration would be the presence of evaporite minerals. These minerals are not present in the sections examined here, with the exception of some excluded sections from Laramide intraforeland lakes (Davis et al., 2008; Davis et al., 2009a; Davis et al., 2009b). Second, in lacustrine rocks and paleosols, other geochemical indicators such as high Mg/Ca and Sr/Ca ratios or δ^{13}C-δ^{18}O covariance in calcite may indicate evaporative environments. Carbon and oxygen isotope ratios can covary because hydrologically closed lakes have longer residence times, leading to more CO_2 outgassing. Since ^{12}C-rich CO_2 preferentially outgases, the lake will become enriched in ^{13}C. Likewise, the lake will be enriched in ^{18}O due to preferential evaporation of ^{16}O-rich H_2O. Mg and Sr will also become enriched in lakes, as they are not flushed from hydrologically closed lakes leading to high Mg/Ca and Sr/Ca ratios in calcite. This type of secondary analysis has not been done for all lacustrine and paleosol samples in this study, but a large subset of the samples used in this study used trace elements to examine evaporative effects (Horton et al., 2004; Kent-Corson et al., 2006; Davis et al., 2008; Davis et al., 2009a; Davis et al., 2009b). Sections suspected of high evaporation due to any of the above reasons were excluded from this study.

Diagenetic Effects: The isotopic composition of authigenic minerals is sensitive to alteration due to burial, cementation, replacement or other diagenetic effects. The most important analysis for diagenetic effects comes from field observations. The new isotopic data presented in this paper excludes any samples that contained sparry calcite, secondary calcite veins, or calcite within vugs. In addition, the published data used in this study is subject to the judgment of many authors' work cited in this paper, and none of the published data used here was suspected of diagenetic alteration.

Climate Effects: The stable isotopic composition of precipitation represents the sum of many complex climatic effects. Unraveling the relationship between elevation change and regional/global climate change is, at best, difficult. Ultimately, surface uplift produces isotopic shifts that reflect changes in local atmospheric circulation and drainage reorganization, so even the target isotopic signal for surface elevation represents the inseparable coupling of tectonics and climate. There are several reasons why we can discount changes in global climate for the isotopic patterns presented in this study. First, since the data presented here are calculated precipitation values, changes in temperature between bins or regions have already been taken into account. Moreover, the cooling that occurred at the Eocene/Oligocene boundary in North America (~8°; (Zanazzi et al., 2007) would cause a decrease of ~4 ‰ in $\delta^{18}O_p$ (Rozanski et al., 1993). However, this isotopic effect should affect all samples across the study area in this time bin rather than the diachronous isotopic shifts that we observe.

Second, global climate change could affect values through a major change in atmospheric circulation patterns. However, we do not think that a rearrangement of the air masses other than the North American Monsoon played a role in the isotopic patterns we observe. The western U.S. is affected by four major air masses. Today, maritime polar, maritime tropical and continental air masses account for 92 % of the precipitation received by the Great Basin (Friedman et al., 2002). These three air masses are all within ~1 ‰ of each other in oxygen, so changes in the relative amounts of precipitation from these air masses would not produce the large isotopic changes we observe between time bins presented in this study. The remaining 8 % of precipitation to the Great Basin comes from the Arctic, which is ~6-7 ‰ lower than the other three primary air masses. However, it is unlikely that the north to south changes we observe in $\delta^{18}O$ values of precipitation can be attributed to southward migration of the Arctic air mass. The Arctic air mass would have to be responsible for increasing influence from north to south during the Eocene, a relatively warm period when it would have been less active at these latitudes. In addition, changes in atmospheric circulation would not likely manifest as the rapid negative shifts seen in the Sage Creek and Elko Basins as this would involve a complete change to Arctic-sourced moisture in a relatively warm period of time. Finally, the isotopic records show that precipitation had relatively low $\delta^{18}O$ in the Eocene and Oligocene before increasing in the late Miocene to Holocene (Horton et al., 2004; Horton and Chamberlain, 2006; Kent-Corson et al., 2006) during a period of global cooling. Again, increased influence of the Arctic air mass would be expected in cooler times, but the isotopic records from the Miocene to Holocene trend toward increasing $\delta^{18}O$ values as cooling progressed.

As we discussed in the paper, monsoonal effects likely played a role in changing the isotopic composition of precipitation across the study area. Currently, [18]O- and D-enriched precipitation from the Gulfs of Mexico and California arrives in the western US in summer monsoonal storms. Less than 5 % of the precipitation reaching central Utah and none in eastern Nevada originate from the Gulf of Mexico (Friedman et al., 2002). The lack of monsoonal storms in this region is due to the orographic barrier created by the Rocky Mountains and the high elevations of the Great Basin. Monsoonal effects are more apparent east of the Rocky Mountains as well as in the American Southwest. In the Eocene, monsoonal storms may have pushed farther north than in the modern. We argue this for several reasons: 1) Warmer temperatures of the Eocene would have increased the moisture carrying capacity of the atmosphere. 2) The Mississippi Embayment extended into Colorado in the Eocene and, thus, the moisture source for the summer monsoon was closer to the Cordillera than in the modern. 3) Since we argue that high topography in much of the Great Basin had not yet formed, there would be fewer orographic barriers preventing monsoonal precipitation from continuing north, although it appears that many of the Rocky Mountain Laramide uplifts were high at this time (Fan and Dettman, 2009). This may be part of the reason for [18]O and D-rich precipitation in Nevada, Utah, Colorado, and Wyoming shown in this study before 49 Ma. 4) Results from climate models suggest an intensified monsoon in the late Cretaceous (Fricke et al., 2010) and Eocene (Sewall and Sloan, 2006).

Extended Methods

Conversion to $\delta^{18}O$ of Precipitation: We began with raw isotopic values for various mineral proxies. These values were converted with the appropriate fractionation equation (see below). For calcite and smectite values, we assigned the temperatures needed for the fractionation equations based on paleofloral estimates, and when necessary, interpolated temporally using the Cenozoic temperature history for mid-latitude (45°) coastal North America (Wolfe, 1994). Some studies have already calculated $\delta^{18}O_p$ using similar methodology. In the cases of the muscovite and aragonite values included in this compilation, we used the authors' calculated water value. The fractionation factor for glass is not temperature dependent, so in this case, we simply converted to water values without temperature estimates. In the case of glass, muscovite and kaolinite, the raw data were hydrogen isotopic values. In these cases, after conversion to the isotopic composition of precipitation, δD values were converted to $\delta^{18}O$ using the meteoric water line.

Fractionation equations used:
Calcite—Kim and O'Neil, 1997
Smectite—Capuano, 1992, and Sheppard and Gilg, 1996
Kaolinite--Gilg and Sheppard, 1996
Glass—Friedman et al., 1993

Other components of the conversion:
Cenozoic mid-latitude temperature history—Wolfe, 1994
Meteoric water line—Craig, 1961

Calculation of Paleoelevations: Paleoelevations were calculated using a single moisture source Rayleigh distillation model (Rowley et al., 2001). We used Eocene relative humidity and sea surface temperatures from Huber and Caballero (2003). The error cited with these estimates includes the uncertainty of the isotopic lapse rate that is calculated from randomly sampling 1000 pairs of relative humidity and mean annual temperature from the output of the Eocene climate model and calculating the change in isotopic composition of precipitation during orographic ascent (Rowley, 2007). For our calculations we used a starting $\delta^{18}O_p$ determined from the $\delta^{18}O$ of Eocene kaolinites ($\delta^{18}O_p$ = -9.9 ‰) located at the Eocene Pacific shoreline (Mulch et al., 2006). Our modeled elevations assume the primary vapor source for these samples was from the Pacific, as it is today.

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Supplemental Table DR1. Isotopic Values used in Compilation

Bin 1: Pre-49 Ma

Region	Latitude	Longitude	Age Ma	Mean δ¹⁸O VSMOW	Material	Mean T °C	δ¹⁸Op VSMOW	1σ δ¹⁸Op	n	Study
Axehandle Basin (UT)	39.41	-111.70	55	22.0	Calcite	22†	-6.7	1.2	262	Bowen and Bowen, 2008
Axehandle Basin (UT)	39.40	-111.68	71.3-50	21.0	Calcite	16.9†	-8.3	2.0	95	This study
Bannock Basin (ID)	42.09	-112.18	50-49	16.4	Calcite	12.5*	-14.3		2	This study
Bighorn Basin (WY)	45.00	-109.00	57.7-53.2	22.1	Calcite	16*	-7.8	0.5	90	Koch et al., 1995
Bighorn Basin (WY)	44.00	-108.80	58.8-53.6		Aragonite	§	-10.3	0.8	272	Dettman and Lohmann, 2000
Bighorn Basin (WY)	44.84	-108.84	55.8	21.6	Calcite	16.5*	-7.9	0.6	338	Bowen et al., 2001
Bighorn Basin (WY)	42.22	-108.53	50.8-51.9	21.6	Calcite	17.5*	-7.1	3.0	118	Clyde et al., 2001
Crazy Mtn Basin (MT)	46.40	-110.20	64.4-59.4		Aragonite	§	-13.8	1.2	45	Dettman and Lohmann, 2000
Elko Basin (NV)	40.56	-116.01	54.5-49.0	25.0	Calcite	21.5*	-4.0	1.1	11	Horton et al., 2004
Flagtaff Basin (UT)	39.00	-110.50	55	21.7	Calcite	22†	-7.0	0.9	48	Bowen et al., 2008
Great Plains/Rocky Mts	43.83	-108.55	50	15.8	Smectite	10.9*	-11.8		2	Sjostrom et al., 2006
Greater Green River Basin (WY	40.76	-108.89	51.8-52.0		Aragonite	§	-7.4	1.8	59	Dettman and Lohmann, 2000
Greater Green River Basin (WY	41.50	-109.00	53		Aragonite	§	-8.0	0.6	31	Morrill and Koch, 2002
Greater Green River Basin (WY	41.50	-109.00	54		Aragonite	§	-10.6	0.7	59	Morrill and Koch, 2002
Greater Green River Basin (WY	41.50	-109.00	51		Aragonite	§	-9.4	0.9	48	Morrill and Koch, 2002
Greater Green River Basin (WY	41.75	-108.80	49	27.0	Calcite	17.2†	-2.7	1.3	79	Carroll et al., 2008
Greater Green River Basin (WY	41.50	-109.50	52	28.3	Calcite	17.2†	-1.4	0.7	151	Norris et al., 1996
Okanagan-Kettle MCC (WA)	48.50	-118.25	49.1		Muscovite	§	-13.3	-0.9	8	Mulch et al., 2007
Piceance Creek Basin (CO)	39.41	-108.18	52	20.3	Calcite	22†	-8.4	3.9	42	Davis et al., 2009
Powder River Basin (WY)	45.50	-106.00	53.7-57.3		Aragonite	§	-13.7	4.2	110	Dettman and Lohmann, 2000
Princeton Basin (BC)	49.44	-120.50	49	11.0	Calcite	8.3†	-19.7	1.7	21	This study
Sage Creek Basin (MT/ID)	44.74	-112.61	65-49	22.2	Calcite	11-16*	-8.4	1.7	14	Kent-Corson et al., 2006
Shuswap MCC (BC)	51.00	-118.25	49.0-47.9		Muscovite	§	-15.8	-0.1	16	Mulch et al., 2007
Uinta Basin (UT)	39.81	-110.94	55	18.9	Calcite	22†	-9.8	1.9	93	Davis et al., 2008
Wind River Range (WY)	43.24	-107.50	52.8-52.0	16.3	Calcite	16.5*	-13.6	4.8	16	This study
Wind River Range (WY)	43.33	-107.63	53.2-52.7	17.2	Calcite	17.5*	-12.6	5.8	21	This study
Wind River Range (WY)	43.49	-109.45	55.5-53.2	20.6	Calcite	19.8*	-8.5	3.9	22	This study
Wind River Range (WY)	43.02	-107.08	60-55	19.4	Calcite	21.5*	-9.3	2.0	10	This study
Wind River Range (WY)	43.22	-107.30	52-50.9	15.9	Calcite	16.5*	-13.9	3.8	15	This study

* Chase et al., 1998

Supplemental Table DR1. Isotopic Values used in Compilation

† Wolfe et al., 1997

§ Used author's calculated water value

Bin 2: 49 to 39 Ma

Region	Latitude	Longitude	Age Ma	Mean $\delta^{18}O$ VSMOW	Material	Mean T °C	$\delta^{18}Op$ VSMOW	1σ $\delta^{18}Op$	n	Study
Bannock Basin (ID)	42.09	-112.18	48-41	18.5	Calcite	12.3*	-12.3	1.4	6	This study
Elko Basin (NV)	40.56	-116.02	47.5-39.4	22.4	Calcite	15.8*	-6.7	2.9	19	Horton et al., 2004
Elko Basin (NV)	40.58	-115.99	42.6-40.6	26.6	Calcite	12.4*	-4.1	4.7	21	This study
Flagstaff Basin (UT)	38.92	-111.81	49-39	21.8	Calcite	13.2*	-8.7	5.7	56	Davis et al., 2008
Galisteo Basin (NM)	35.50	-106.00	49-39	21.9	Calcite	7.5*	-10.0	1.2	36	Davis et al., 2009
Great Plains/Rocky Mts	43.50	-108.58	47.9	14.9	Smectite	10*	-12.9	0.5	3	Sjostrom et al., 2006
Great Plains/Rocky Mts	42.76	-107.59	49	14.2	Smectite	10.9*	-13.4		2	Sjostrom et al., 2006
Great Plains/Rocky Mts	42.71	-108.18	49	14.4	Smectite	10.9*	-13.2		2	Sjostrom et al., 2006
Great Plains/Rocky Mts	42.60	-108.29	49	15.7	Smectite	10.9*	-11.9		1	Sjostrom et al., 2006
Great Plains/Rocky Mts	44.85	-110.16	49	9.1	Smectite	10.9*	-18.5		2	Sjostrom et al., 2006
Great Plains/Rocky Mts	44.87	-111.05	49	10.2	Smectite	10.9*	-17.4		2	Sjostrom et al., 2006
Great Plains/Rocky Mts	44.89	-111.33	49	9.5	Smectite	10.9*	-18.1		2	Sjostrom et al., 2006
Greater Green River Basin (WY	41.75	-108.80	48	22.5	Calcite	17.2†	-7.2	1.2	57	Carroll et al., 2008
Ibapah Basin (UT)	40.12	-114.12	46-39	19.8	Calcite	13.2*	-10.8	2.2	21	Davis et al., 2009
Sage Creek Basin (MT/ID)	44.66	-112.58	48.0-39.1	13.5	Calcite	13.4*	-17.1	1.8	58	Kent-Corson et al., 2006
Sierra Nevada (CA)	39.35	-120.98	44.9		Kaolinite	17.1*	-9.3	0.5	25	Mulch et al., 2006
Sierra Nevada (CA)	39.18	-120.85	44.9		Kaolinite	17.1*	-10.1		2	Mulch et al., 2006
Sierra Nevada (CA)	39.69	-120.93	44.9		Kaolinite	17.1*	-10.0	0.0	5	Mulch et al., 2006
Sierra Nevada (CA)	39.69	-120.98	44.9		Kaolinite	17.1*	-9.8		2	Mulch et al., 2006
Uinta Basin (UT)	40.05	-110.53	49-39	27.0	Calcite	13.2*	-3.6	2.8	76	Davis et al., 2008

* Chase et al., 1998

† Wolfe et al., 1997

§ Used author's calculated water value

Supplemental Table DR1. Isotopic Values used in Compilation

Bin 3: 39 to 28 Ma

Region	Latitude	Longitude	Age Ma	Mean $\delta^{18}O$ VSMOW	Material	Mean T °C	$\delta^{18}Op$ VSMOW	$1\sigma\ \delta^{18}Op$	n	Study
Claron Basin (UT)	37.68	-112.83	39-28	14.8	Calcite	11*	-16.3	1.4	34	Davis et al., 2008
Copper Basin (NV)	41.76	-115.47	38.7	9.6	Smectite	10.5†	-18.2	3.2	8	This study
Elko Basin (NV)	40.55	-115.96	36.0-28.0	14.5	Calcite	13*	-16.1	1.5	10	Horton et al., 2004
Flastaff Basin (UT)	38.93	-111.95	39-28	15.4	Calcite	11*	-15.7	0.4	11	Davis et al., 2008
Galisteo Basin (NM)	35.50	-106.00	39-28	21.8	Calcite	0*	-11.8	0.9	16	Davis et al., 2009
Great Plains/Rocky Mts	42.83	-101.70	35	17.7	Smectite	11.5*	-9.8		1	Sjostrom et al., 2006
Great Plains/Rocky Mts	43.66	-102.66	35.8	19.4	Smectite	11.5*	-8.1	0.8	5	Sjostrom et al., 2006
Great Plains/Rocky Mts	41.60	-103.11	31.25	18.2	Smectite	11.5*	-9.2	1.0	6	Sjostrom et al., 2006
Great Plains/Rocky Mts	41.70	-103.34	31.25	18.9	Smectite	11.5*	-8.6	0.4	2	Sjostrom et al., 2006
Great Plains/Rocky Mts	42.92	-103.49	34.5	16.9	Smectite	11.5*	-10.6		1	Sjostrom et al., 2006
Great Plains/Rocky Mts	42.77	-103.56	30.3	17.0	Smectite	11.5*	-10.5		2	Sjostrom et al., 2006
Great Plains/Rocky Mts	42.76	-105.01	35.5	16.6	Smectite	11.5*	-10.9		1	Sjostrom et al., 2006
Great Plains/Rocky Mts	42.71	-105.37	31.25	14.7	Smectite	11.5*	-12.8		1	Sjostrom et al., 2006
Great Plains/Rocky Mts	42.64	-106.76	35.65	14.1	Smectite	11.5*	-13.3	0.5	12	Sjostrom et al., 2006
Great Plains/Rocky Mts	42.76	-107.56	32	10.9	Smectite	11.5*	-16.6		1	Sjostrom et al., 2006
Great Plains/Rocky Mts	42.58	-108.29	32	10.6	Smectite	11.5*	-16.9		1	Sjostrom et al., 2006
Sage Creek Basin (MT/ID)	44.76	-112.55	38.8-32.0	14.1	Calcite	10.8*	-17.0	1.0	36	Kent-Corson et al., 2006
Sierra Nevada (CA)	39.70	-120.99	28.5		Glass		-14.3		1	Cassel et al., 2009
Sierra Nevada (CA)	39.85	-120.35	28.5		Glass		-16.8	0.6	8	Cassel et al., 2009
Sierra Nevada (CA)	39.25	-120.93	28.5		Glass		-13.2	0.3	8	Cassel et al., 2009
Sierra Nevada (CA)	39.46	-120.49	28.5		Glass		-15.3	0.8	11	Cassel et al., 2009
Sierra Nevada (CA)	39.56	-120.50	28.5		Glass		-16.2		2	Cassel et al., 2009
Sierra Nevada (CA)	38.19	-120.85	28.5		Glass		-10.6		2	Cassel et al., 2009

Supplemental Table DR1. Isotopic Values used in Compilation

Uinta Basin (UT)	40.33	-109.53	39-28	17.7	Calcite	11*	-13.4	3.5	40	Davis et al., 2008
Wind River Range (WY)	42.70	-108.19	36.8-28.3	16.8	Calcite	11.3*	-14.1	1.7	36	This study
Wind River Range (WY)	43.34	-107.48	37.0-34.0	19.8	Calcite	11.5*	-11.1	1.4	8	This study

* Chase et al., 1998

† Wolfe et al., 1997

§ Used author's calculated water value