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Stable isotopic evidence for a pre–late Miocene elevation gradient in the Great Plains–Rocky Mountain region, USA

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ABSTRACT

In order to investigate if high elevations existed in the Rocky Mountains before the late Miocene, we examined oxygen isotope ratios of 63 Tertiary smectite samples as a proxy for the isotopic composition of precipitation. Of these samples, 51 were also analyzed for hydrogen isotope ratios. These smectites were formed as a result of the weathering of volcanic air-fall deposits that blanketed much of western North America during the Tertiary. Smectite-bearing ashfall samples were collected from Eocene, Oligocene, and Miocene deposits along a transect that extends from the western Great Plains to Yellowstone National Park at modern elevations from ~900 to ~2800 m. In general, oxygen and hydrogen δ values of smectite lie along a line parallel to the meteoric water line, which suggests that the isotopic composition of these ash-derived smectites records the meteoric water composition during its formation. There is little evidence for postdepositional exchange with basinal brine fluids, evaporative effects, or diagenesis of these smectites. The $\delta^{18}O$ values of Oligocene and Miocene samples increase ~6‰ linearly from sample sites located at the crest of the Rocky Mountains to sites in western Nebraska and South Dakota. These results mimic the distribution and values of calculated oxygen isotope ratios of theoretical modern smectite over this same geographic traverse of decreasing elevation. This result suggests modern atmospheric circulation patterns and that the resulting distribution of $\delta^{18}O_{precipitation}$ has persisted since the Oligocene. The $\delta^{18}O$ values of Eocene samples increase ~8‰ between the Yellowstone region and central Wyoming, a result that does not correlate with modern $\delta^{18}O_{precipitation}$ trends. Our Eocene results may be explained by climate conditions extant at that time, but tectonic modification in the region between 50 Ma and 37 Ma cannot be excluded as the cause of our results. Because the modern climate system requires interaction with and modification by high-elevation areas, our results suggest that the Rocky Mountains have been at high elevation since at least 50 Ma.

Keywords: stable isotopes, smectite, Rocky Mountains, paleoclimate, paleotopography.

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INTRODUCTION

There is considerable debate as to when the Rocky Mountains formed as a significant topographic feature. The traditional view holds that the modern relief that exists throughout the area is a recent feature that formed as a result of late- and post-Miocene uplift. According to this interpretation, the low-relief topography and shallow-dipping Tertiary sedimentary deposits that are a common feature at high elevations throughout the Rocky Mountain region are interpreted as evidence for Early and Middle Tertiary deposition on a broad, flat, low-elevation plain (Bradley, 1987; Burchfiel et al., 1992). The present topography resulted from Late Tertiary tectonic activity, uplift, and incision (Epis and Chapin, 1975; Steven et al., 1997; McMillan et al., 2002). In contrast, recent paleofloral and stable isotopic studies suggest that elevations as high or higher than present may have existed in the Rocky Mountains since, at the latest, the end of the Laramide orogeny, and much of the present relief is the result of increased erosion and exhumation due to Late Tertiary climate change (Molnar and England, 1990; Gregory and Chase, 1992; Drummond et al., 1993; Gregory and McIntosh, 1996; Norris et al., 1996; Chase et al., 1998; Wolfe et al., 1998; Dettman and Lohmann, 2000; Peizhen et al., 2001).

To address this debate, we conducted an oxygen and hydrogen isotopic study of authigenic smectites from Eocene (50–49 Ma), Oligocene (37–32 Ma), and Miocene (20–12 Ma) volcanic ashes that are exposed along a transect from the Continental Divide to the Great Plains (Fig. 1). Our study focuses on smectite because its oxygen and hydrogen isotopic values can be used to reconstruct past climate and topographic histories of mountain belts (Lawrence and Taylor, 1971; Stern et al., 1997; Chamberlain and Poage, 2000; Poage and Chamberlain, 2002). In addition, Tertiary volcanic ashes containing abundant smectite are exposed along the eastern flanks of the Rocky Mountains and extend into the western Great Plains. Because these ash layers were initially deposited as volcanic air-fall, time-equivalent deposits blanket a large geographic area. Thus, it is possible to determine isotopic gradients in paleoprecipitation over a broad area, which can be diagnostic of topography.

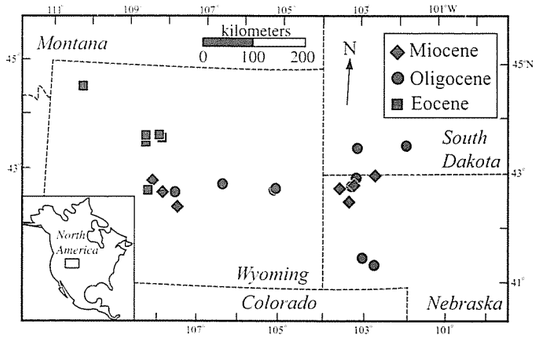

Figure 1. Locations of Tertiary smectite sample sites in the western Great Plains and eastern Rocky Mountain region. All samples were collected from air-fall ashes of Miocene, Oligocene, and Eocene age.

MODERN CLIMATE OF THE WESTERN GREAT PLAINS AND EASTERN ROCKY MOUNTAINS

In contrast to the relatively simple climate systems used in previous isotopic paleoelevation studies, such as Poage and Chamberlain (2001) and Rowley et al. (2001), the modern distribution of the isotopic composition of precipitation in the eastern Rocky Mountains and western Great Plains is fairly complex. The annual average isotopic values of precipitation depends on the annual timing of precipitation, average annual temperature, and the interaction among three isotopically discrete air masses that originate from three different areas (Bryson and Hare, 1974; Nativ and Riggio, 1990; Thordsen et al., 1992; Harvey and Welker, 2000). The modern climate dynamics of the region as described in the following tend to make the change in isotopic composition with elevation difference appear much greater in the western Great Plains and eastern Rocky Mountains than those predicted by the models of Poage and Chamberlain (2001) and Rowley et al. (2001).

The modern climate of the Rocky Mountain region and the western Great Plains is influenced by the direct competition of three distinct air masses that originate over the Arctic, the Gulf of Mexico, and the Pacific Ocean, and by the mountains that create rain shadows for these air masses (Bryson and Hare, 1974). The interplay of these three air masses causes strong seasonal variations in the amount of precipitation and temperature. In the high elevations of the northern Rocky Mountains, dry summer, wet winter conditions dominate as a result of the influence of the eastern Pacific subtropical high-pressure system. During the summer, the westerlies are relatively weak and displaced to the north, causing the summer months to be relatively dry at high elevations. Heating of the continental interior creates a thermal low-pressure system that draws moisture north from the Gulfs of Mexico and California (Bryson and Hare, 1974). This summer monsoon system dominates the low-elevation areas of the northern Rocky Mountains, which are characterized by relatively wet summers and dry winters compared to high-elevation areas (Despain, 1987; Whitlock and Bartlein, 1993). Areas of high elevation in the Rocky Mountains interact with Pacific-sourced air masses, which causes these areas to receive relatively abundant precipitation in the winter (Despain, 1987). These air masses pass over low-elevation areas unaffected, and the lack of orographic precipitation results in lower precipitation amounts in the winter (Despain, 1987). These local climate trends influence the pattern of seasonal precipitation over relatively small geographic areas, in some cases on the scale of only a few kilometers (Despain, 1987; Whitlock and Bartlein, 1993).

East of the Rocky Mountains, the effects of local topography are much reduced. The climate is dominated by interaction of warm, moist air masses that originate over the Gulfs of Mexico and California, cold, dry arctic air masses that originate over the Arctic Ocean, and mild, dry westerlies from the Pacific (Bryson and Hare, 1974). The seasonal dominance of the southern air masses in the summer results in relatively wet, humid summers, with increasing amounts of moisture to the east. In the winter, the Arctic and Pacific air masses dominate, which results in cold, dry winters, especially in the rain shadow of the Rocky Mountains.

The climate variability of the Yellowstone–Rocky Mountain region is reflected in the stable isotope composition of precipitation from the area; $\delta^{18}O_{precipitation}$ ranges from −17‰ to −22‰ $\{\delta^{18}O = 1000[(^{18}O/^{16}O)_{sample}/(^{18}O/^{16}O)_{standard}] - 1\}$; we report all $\delta^{18}O$ and δD values relative to Vienna standard mean ocean water (V-SMOW) (Thordsen et al., 1992; Horton et al., 1999). The dominant control on isotopic composition of precipitation in the Yellowstone area is elevation; lighter isotopic values occur in areas of high elevation (Thordsen et al., 1992). To the east of the Rocky Mountains, oxygen isotopic values of precipitation are more enriched in ^{18}O. In the Wind River Basin of central Wyoming, oxygen isotope values of soil water in equilibrium with modern carbonates range from ~−18‰ to ~−12‰ (Amundson et al., 1996). In the western Great Plains of western Nebraska and South Dakota, oxygen isotope values of precipitation average ~−9.8‰ (Harvey and Welker, 2000). East of the Rocky Mountains, the average oxygen isotope value of precipitation is highly dependent on the annual timing of precipitation. Areas in the eastern part of the study area (western Great Plains) receive more summer precipitation than western areas, and therefore have higher average $\delta^{18}O_{precipitation}$ values (Amundson et al., 1996; Harvey and Welker, 2000). This results in a present-day average $\delta^{18}O_{precipitation}$ difference of ~10‰ between Yellowstone National Park (elevation ~2500 m) and North Platte, Nebraska (elevation, ~900 m) or $\Delta\delta^{18}O_{precipitation} = -6.25$‰/km elevation difference. This is a much more dramatic $\delta^{18}O_{precipitation}$ gradient than the global average (~−2.8‰/km elevation difference) reported by Poage and Chamberlain (2001). Similarly high isotopic gradients (−4.2‰/km) have also been observed in nearby areas of the eastern Rocky Mountains (Drummond et al., 1993). These greater than average isotopic gradients are the result of elevation increase to the west in our study area, which results in relatively light $\delta^{18}O_{precipitation}$. This is coupled with a much greater average annual amount of isotopically light winter precipitation in the western part of the study area than to the east.

METHODS

Approximately 160 samples were collected from previously described and age-dated Eocene, Oligocene, and Miocene ash layers on an east-west transect from western Nebraska and South Dakota to western Wyoming (Fig. 1; Table 1). Samples were ground with an iron mortar and pestle and dispersed in distilled water using a commercial blender. The samples were centrifuged to a particle size of <0.5μm to remove nonsmectite clay and other minerals with a larger grain size. X-ray analysis was conducted on a Siemens D-500 diffractometer.

Of the 160 samples that were collected, 63 consisted of pure smectite; these pure smectite samples were selected for isotopic analysis. None of the 160 diffractograms showed the presence of significant mixed layering with other clay minerals by inspection and/or by using the $\Delta 2\theta$ (002/003) – $\Delta 2\theta$ (001/002) method of Moore and Reynolds (1997). This includes mixed layering of illite and smectite that would be indicative of diagenetic alteration. The diffractogram of the glycolated sample shown in Figure 2 is typical of the smectite samples selected for isotopic analysis.

Oxygen extraction procedures followed those outlined in Poage and Chamberlain (2002). An internal quartz standard (BX-88) was analyzed with each loading of the extraction line, and an internal smectite standard (01DS069) was analyzed in several of the batch runs. Extraction line runs for which ratios of the standards were outside of the range of ±0.2‰ for quartz or ±0.4‰ for smectite were discarded, and the samples were reanalyzed.

We also collected hydrogen isotopic values of 51 of the smectite samples using a continuous flow online combustion method similar to that described in Sharp et al. (2001). Our analysis of hydrogen isotope standards gave a precision and accuracy of the following: (1) on NBS 22 (oil) the accepted δD value is −118.5‰; our value was −118.6 ± 1.7‰; (2) Clay Mineralogy Source clays were: kga-1 (kaolinite) −64.1 ± 1.5‰; kga-2 (kaolinite) −65.9 ± 0.8‰; STx-1 (smectite) −99.2 ± 1.0‰; and SAz-1 (smectite) −109.2 ± 2.2‰. The δD values of these clays have not been previously reported.

RESULTS

Five results follow from our analyses:

1. With the exception of a few samples, the combined oxygen and hydrogen isotope values of smectite cluster along a line parallel to and displaced from the meteoric water line (Fig. 3; Table 1). Only a small number of the samples (notably samples: 01DS39, 01DS16, 01DS129, and 01DS124) are significantly displaced from this line. This result is similar to results originally described by Lawrence and Taylor (1971), who found that the isotopic composition of kaolinite, montmorillonite, and gibbsite from Quaternary soils throughout North America correlate with the present-day meteoric waters. In fact, the oxygen and hydrogen isotopic values of smectite presented in this paper are similar to the isotopic values of montmorillonite in modern soils from Montana and Idaho (Lawrence and Taylor, 1971).
2. There is a wide range of both hydrogen and oxygen isotope ratios of smectite, and the isotopic values increase toward the east. The δD values of smectite range from −126‰ to −88‰, and $\delta^{18}O$ values of smectites range from 8.7‰ to 20.2‰. In general, both δD and $\delta^{18}O$ values of smectite from eastern sections are ~40‰ and ~6‰

TABLE 1. LOCATION, GEOLOGIC GROUP OR FORMATION, AGE, ISOTOPE VALUE, MODERN ELEVATION, FIELD LOCATION, AND AGE REFERENCE FOR SMECTITE SAMPLES

Sample	Latitude (°N)	Longitude (°W)	Formation	Age (Ma)	$\delta^{18}O$	δD	Elevation (m)	Reference
Miocene samples								
01DS37	42.4373	103.4184	Arikaree	21.9	19.7	–127	1340	2
01DS28	42.769	103.5354	Arikaree	22	18.8	–93.0	1300	2
01DS36	42.4271	103.5419	Arikaree	21.9	16.2	–121	1340	2
01DS44	42.9492	103.6854	Arikaree	22	18.8	–97	1100	2
01DS35	42.4234	103.7525	Arikaree	21.9	15.0	–94.0	1340	2
01DS33	42.7425	103.8113	Arikaree	22	16.7	–93.0	1100	2
01DS34	42.7154	103.8828	Eagle Crag Ash	19.2	16.8	–95.0	1100	2
01DS104	42.446	107.5316	Split Rock	Middle Miocene	15.7	–107	1770	6
01DS102	42.4446	107.5588	Split Rock	Middle Miocene	12.7	–110	1770	6
01DS103	42.435	107.5628	Split Rock	Middle Miocene	12.8	–118	1770	6
01DS116	42.7288	107.5826	Split Rock	Middle Miocene	12.8	–112	2250	6
01DS115	42.7308	107.5845	Split Rock	Middle Miocene	11.9	–113	2250	6
01DS121	42.6817	108.1239	Split Rock	Middle Miocene	14.5	–116	1800	6
Oligocene samples								
01DS55	43.8288	101.6981	Peanut Peak	35	17.7	–107	760	4
01DS69	43.863	102.8605	Chadron	37–34	18.6	–103	850	5
01DS70	43.8614	102.8631	Chadron	37–34	20.2	–	850	5
01DS72	43.8614	102.8631	Chadron	37–34	18.4	–101	850	5
01DS74	43.8614	102.8631	Chadron	37–34	20.0	–101	850	5
01DS75	43.8608	102.8636	Chadron	37	19.6	–99	850	5
01DS12	41.5968	103.1131	White River	31.9–30.6	18.3	–94.0	1170	1
01DS14	41.5968	103.1131	White River	31.9–30.6	18.0	–88.0	1170	1
01DS15	41.5968	103.1131	White River	31.9–30.6	18.0	–92.0	1170	1
01DS16	41.5968	103.1131	White River	31.9–30.6	20.2	–110	1170	1
01DS17	41.5968	103.1131	White River	31.9–30.6	17.6	–102	1170	1
01DS18	41.5968	103.1131	White River	31.9–30.6	17.3	–123	1170	1
01DS19	41.6991	103.3415	White River	31.9–30.6	19.2	–100	1290	1
01DS20	41.6991	103.3415	White River	31.9–30.6	18.6	–90.0	1290	1
01DS39	42.9194	103.4913	Peanut Peak	34.5	16.9	–123	1160	4
01DS31	42.7791	103.5621	Brule	30.6–30	17.2	–93.0	1170	3
01DS30	42.7688	103.5835	Brule	30.6–30	16.8	–107	1170	3
01DS81	42.7625	105.0128	Chadron	37–34	16.6	–101	1600	1
01DS84	42.7125	105.371	White River	31.9–30.6	14.7	–112	1520	1
01DS89	42.6386	106.7545	White River	35.9–35.4	13.9	–107	1830	1
01DS90a	42.6386	106.7545	White River	35.9–35.4	15.5	–122	1830	1
01DS91	42.6386	106.7545	White River	35.9–35.4	14.4	–118	1830	1
01DS92	42.6386	106.7545	White River	35.9–35.4	14.2	–109	1830	1
01DS93	42.6386	106.7545	White River	35.9–35.4	14.1	–122	1830	1
01DS94	42.6386	106.7594	White River	35.9–35.4	13.9	–121	1830	1
01DS95	42.6386	106.7594	White River	35.9–35.4	13.7	–124	1830	1
01DS97	42.6386	106.7594	White River	35.9–35.4	14.3	–117	1830	1
01DS98	42.639	106.765	White River	35.9–35.4	13.4	–126	1830	1
01DS99	42.639	106.765	White River	35.9–35.4	14.3	–124	1830	1
01DS100	42.639	106.765	White River	35.9–35.4	13.6	–120	1830	1
01DS101	42.639	106.765	White River	35.9–35.4	14.0	–113	1830	1
01DS113	42.7574	107.5635	White River	ca. 32	10.9	–126	2200	6
01DS129	42.5848	108.287	White River	ca. 32	10.6	–104	2500	6
01DS129	42.5848	108.287	White River	ca. 32	10.6	–104	2500	6
Eocene samples								
01DS110	42.7575	107.5874	Wagon Bed	ca. 49	13.6	–122	2150	6
01DS111	42.7608	107.59	Wagon Bed	ca. 49	14.7	–122	2150	6
01DS123	42.7091	108.1809	Wagon Bed	ca. 49	16.1	–107	1900	6
01DS124	42.7091	108.1809	Wagon Bed	ca. 49	12.7	–101	1900	6
01DS128	42.5964	108.2933	Wagon Bed	ca. 49	15.7	–113	2500	6
6C-1	43.83	108.55	Lower Aycross	ca. 50	15.8	–	2150	7
6C-3	43.83	108.55	Lower Aycross	ca. 50	15.8	–	2152	7
6A-1	43.501	108.58	Upper Aycross	47.9 ± 0.5	14.5	–	2193	7
6A-2	43.501	108.58	Upper Aycross	47.9 ± 0.5	15.4	–	2195	7
6A-3	43.501	108.58	Upper Aycross	47.9 ± 0.5	14.8	–	2197	7
Mnau	44.8542	110.1567	Lamar River	49	9.4	–	2160	8
MNI	44.8542	110.1567	Lamar River	49	8.7	–	2160	8
CC2	44.8655	111.0488	Lamar River	49	10.2	–	2651	8
CC3	44.8655	111.0488	Lamar River	49	10.1	–	2651	8
SR1	44.8946	111.3345	Lamar River	49	9.2	–	2329	8
SR2	44.8946	111.3345	Lamar River	49	9.7	–	2329	8

Note: 1—MacFadden and Hunt (1998); 2—Love (1970); 3—Larson and Evanoff (1998); 4—Terry (1998); 5—Evans and Welzenbach (1998); 6—LaGarry (1998); 7—Sundell et al. (1984); 8—Fritz (1980).

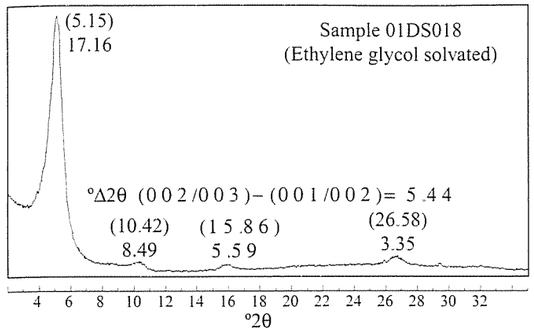

Figure 2. Representative X-ray pattern of glycolated smectite.

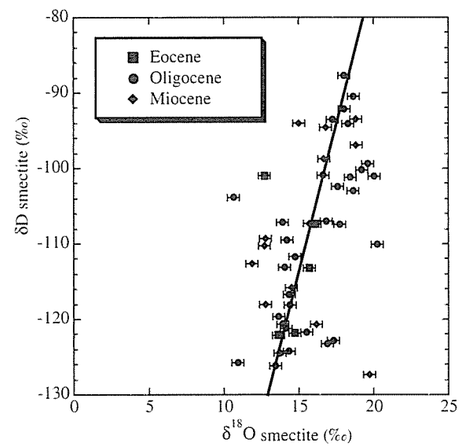

Figure 3. Combined hydrogen and oxygen isotope analyses of smectites examined in this study. Error bars represent analytical uncertainty of oxygen isotope analysis; analytical uncertainty for hydrogen isotopes falls within the range covered by each symbol. Trend of isotopic data is approximately parallel to the meteoric water line, which suggests these smectites formed in equilibrium with meteoric water and have not been modified isotopically by postformation events.

higher, respectively, than δD and $\delta^{18}O$ values of smectite from western sections, irrespective of geologic age.

3. The $\delta^{18}O$ values of smectites from Oligocene and Miocene samples show a strong correlation with modern altitude, with lower values at higher altitudes (Figs. 4 and 5). However, there is essentially no correlation between $\delta^{18}O_{smectite}$ and modern elevation for the Eocene samples (Fig. 6). The δD data from smectite of Oligocene and Miocene samples show a similar correlation with altitude, however, the correlation is not as good as the correlation of oxygen isotope ratios and altitude. Presumably the lower correlation results from the increased scatter in hydrogen isotopes of smectite; scatter which may reflect some degree of post-depositional exchange, evaporative effects, and exchange with basinal fluids, all of which affect hydrogen isotopes to a greater degree than oxygen isotopes. For this reason, we largely restrict our analysis to oxygen isotopes, and simply use the hydrogen isotopes of smectite to assess the role of evaporation and exchange with basinal brines.

4. The $\delta^{18}O_{smectite}$ values for Oligocene and Miocene samples are similar to $\delta^{18}O$ for modern hypothetical smectites calculated to be in isotopic equilibrium with modern precipitation. The dashed lines in Figures 4, 5, and 6 represent the theoretical oxygen isotope value of smectite formed in equilibrium with modern precipitation and was calculated with a smectite-water fractionation factor (Savin and Lee, 1988), modern meteoric precipitation isotope values from North Platte, Nebraska (Harvey and Welker, 2000), central and eastern Wyoming (Coplen and Kendall, 2000), and Yellowstone National Park (Thordsen et al., 1992), and modern annual average temperature at these three sites.

5. No temporal trends in isotopic values were evident where multiple samples were collected throughout a stratigraphic section. The variation in $\delta^{18}O_{smectite}$ was typically much less than 3‰ at these locations. Considering our analytical error of ±0.4‰, there is essentially no difference in oxygen isotope ratios temporally between Miocene and Oligocene samples. Eocene samples have slightly lower oxygen isotope values than Miocene and Oligocene samples overall (Fig. 7).

DISCUSSION

There are several possible explanations for the geographic variation of $\delta^{18}O_{smectite}$ from Tertiary volcanic airfall deposits of the eastern Rocky Mountains and western Great Plains, which include: (1) the modern elevation gradient and climate system on the east side of the Rocky Mountains has been in place since the Oligocene; (2) the smectites do not record paleoprecipitation, but instead have equilibrated with modern waters; (3) the smectites formed relatively recently in these volcanic air-fall ashes; and/or (4) the oxygen isotope values of smectites have been altered by later diagenesis or exchange with basinal fluids. For the following reasons, we believe that the possible effects of points 2 through 4 did not significantly influence the results presented in this paper. Instead, we suggest that the oxygen isotope values of smectites record the paleoprecipitation and elevation gradient on the east side of the Rocky Mountains that has existed since at least Oligocene times.

First, it could be argued that the similarity between the spatial distribution of $\delta^{18}O_{smectite}$ from Early Tertiary air-fall deposits and the $\delta^{18}O$ values of modern precipitation is the result of

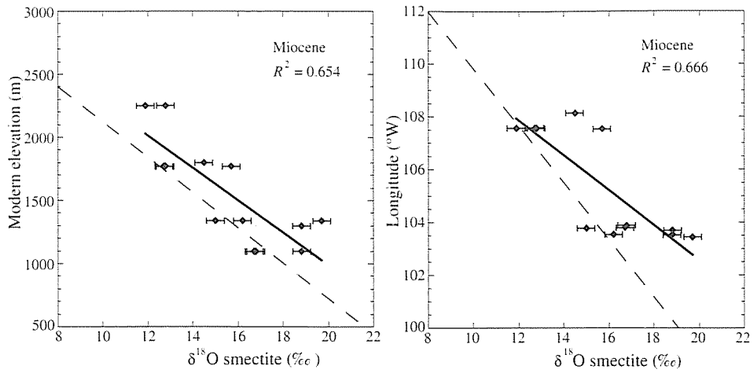

Figure 4. Results of oxygen isotope analysis of Miocene smectite versus modern elevation and longitude. Error bars represent analytical uncertainty of ±0.4‰. The dashed line represents the theoretical oxygen isotope value of smectite formed in equilibrium with modern precipitation (see text). Heavy solid line represents the linear regression best fit to the measured smectite isotopic data. Miocene isotopic results correlate well with modeled, modern smectites.

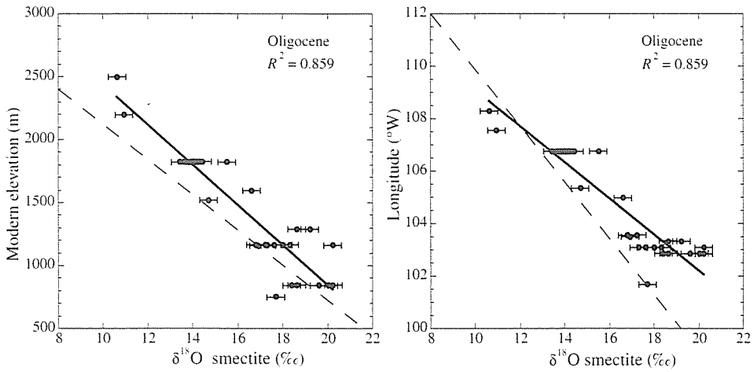

Figure 5. Results of oxygen isotope analysis of Oligocene smectite versus modern elevation and longitude. Error bars represent analytical uncertainty of ±0.4‰. The dashed line represents the theoretical oxygen isotope value of smectite formed in equilibrium with modern precipitation (see text). Heavy solid line represents the linear regression best fit to the measured smectite isotopic data. Oligocene isotopic results correlate well with modeled, modern smectites. Together with the Miocene results from Figure 4, these data suggest that the oxygen isotope gradient in the Rocky Mountain region has remained relatively unchanged since at least 37 Ma.

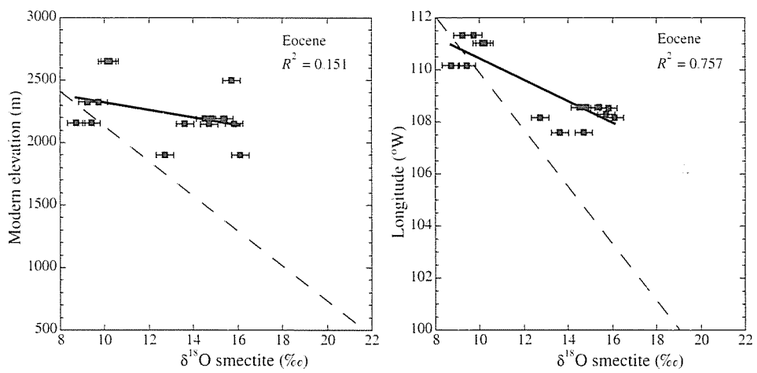

Figure 6. Results of oxygen isotope analysis of Eocene smectite versus modern elevation and longitude. Error bars represent analytical uncertainty of ±0.4‰. The dashed line represents the theoretical oxygen isotope value of smectite formed in equilibrium with modern precipitation (see text). Heavy solid line represents the linear regression best fit to the measured smectite isotopic data. Eastern, low-elevation samples are heavier than our modeled modern smectites, which is likely due to a warmer climate in the Eocene. The western, high-elevation samples, which correlate well with modeled, modern smectites, cannot be explained by global climate change alone.

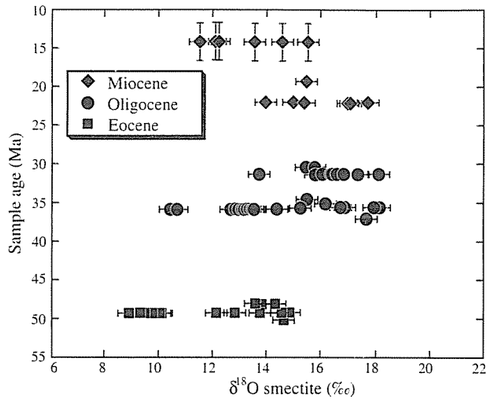

Figure 7. Oxygen isotope results versus age of sample. No correlation exists between age and isotope ratio (the R^2 for this relationship is <0.01). Error bars represent analytical uncertainty of ±0.4‰. Error for the ages is reported within the cited reference listed in Table 1 and falls within the range covered by the symbol where absent. Eocene-aged samples are slightly lighter than younger samples; these smectites formed in locations of relatively high modern elevation and low average modern $\delta^{18}O_{precipitation}$. It appears that geographic location rather than age is a much more important factor for $\delta^{18}O_{smectite}$, which suggests long-term stability of atmospheric circulation patterns in the sample area.

isotopic equilibrium with modern precipitation, and may not be representative of Early Tertiary isotopic composition. It has been shown, however, that pedogenic clay minerals retain their original structural oxygen isotopic composition and do not reequilibrate isotopically after formation (Lawrence and Taylor, 1971; Yeh and Savin, 1977; Stern et al., 1997; Savin and Hsieh, 1998). This work is corroborated by studies showing that smectite and calcite (which is extremely resistant to postdepositional isotopic exchange; Cerling and Quade, 1993) record similar isotopic trends in rocks where both minerals have been examined (Stern et al., 1997; Poage and Chamberlain, 2002).

Second, it is possible that the smectite we analyzed may be much younger than the age of the sedimentary deposits, and the isotope values preserved in these minerals therefore record relatively recent formation. We consider this possibility unlikely, because there is abundant evidence that smectite, as an alteration product, forms relatively rapidly in the shallow subsurface. Strontium isotopic studies and petrographic analysis of calcite, smectite, and silica cements of sandstones from the White River Group, Arikaree Group, and Miocene-Pliocene Ogallala Group suggest that these authigenic minerals formed in the Tertiary soon after deposition of the sediments (Stanley and Benson, 1979; Stanley and Faure, 1979). Additionally, studies of glacial moraines as young as 10^4 yr old in the Rocky Mountains contain smectite as a weathering product (Blum and Erel, 1997), Quaternary soils in this region contain authigenic smectite (Lawrence and Taylor, 1971), and ash from the 1980 Mt. St. Helens eruption has already converted to clay minerals in many locations (G. Thompson, 2003, personal commun.).

Third, it is possible that the oxygen isotopes of the smectites we measured record some later period of diagenesis or exchange with basinal brines. However, during burial diagenesis, smectite converts to mixed-layer illite/smectite to pure illite with increasing burial depth. The conversion of smectite to illite begins to take place at temperatures above 35 °C, and most of the conversion takes place between 55 °C and 95 °C (Perry and Hower, 1970). X-ray diffraction analysis conducted on the air-fall ashes examined in this study showed only a small percentage of mixed layering of illite/smectite (<5% illite) in a small number of samples. Thus, it is unlikely that the ashes reached burial temperatures of ~50 °C, whereas temperatures greater than 100 °C are necessary to cause isotopic exchange between smectite and water (O'Neil and Kharaka, 1976; Yeh and Savin, 1977). Assuming the modern geothermal gradient of the western Great Plains, which is ~28 °C/km, and an average annual surface temperature of ~10 °C, temperatures of 35 °C would be achieved at a burial depth of ~900 m. However, estimations of the original thickness or preserved thickness of the Tertiary and Quaternary deposits do not exceed 800 m (Love, 1970; Hunt, 1990; Larson and Evanoff, 1998). Moreover, there is little isotopic evidence for later exchange with basinal brines, as the combined δD and $\delta^{18}O$ values of smectite cluster along a line parallel to the meteoric water line (Fig. 3) rather than forming a field with a significantly shallower slope due to water-rock interaction (see Longstaffe, 1987).

We, therefore, suggest the observed geographic trends in smectite oxygen isotope values record the long-term precipitation history of the eastern Rocky Mountains. Our data suggest that the Oligocene and Miocene paleoprecipitation patterns were similar to that of today; and there was likely a significant difference in the climate of this region in the Eocene. As mentioned already, the isotopic gradient on the east side of the Rocky Mountains is primarily controlled by the interaction of three distinct air masses, each with a different isotopic signature. Climate models show that this unique climate regime is related to the topography of the region (Kutzbach et al., 1989; Huber and Sloan, 1999; Sewall et al., 2000). The fact that the Miocene and Oligocene isotopic gradients are similar to modern isotopic gradients, therefore, suggests that the Rocky Mountains as a significant topographic feature have existed since at least that time.

However, since the Oligocene, there has been a continuous ~2–4 °C decrease in global temperature (Wolfe, 1978; Prothero and Heaton, 1996; Terry, 2001) and an increase in the oxygen isotope composition of ocean waters due to development of significant ice sheets since ca. 25 Ma (Zachos et al., 2001). Both of these effects could also influence the isotopic composition of precipitation. To assess these effects, we calculated the isotopic composition of smectite, assuming that the isotopic changes in ocean water (Zachos et al., 2001) cause a direct effect on $\delta^{18}O_{precipitation}$, and the temperature changes affect both the $\delta^{18}O_{precipitation}$ (Savin and Hsieh, 1998) and the isotopic

fractionation between smectite and water (Savin and Lee, 1988). The results of these calculations show that the smectite in the Oligocene and Miocene samples would have essentially the same $\delta^{18}O$ values as modern samples (our results for these calculations were less than our analytical error of ±4 for our oxygen isotope analysis), and they plot essentially on top of the calculated modern smectite values represented in Figures 4–6 by the dashed lines. Therefore, we suggest that magnitude of change in annual temperature and oxygen isotope composition of ocean water as reported in the literature does not significantly influence the isotopic gradient observed in our Oligocene and Miocene samples when compared to present values, and the stability of atmospheric circulation patterns is a much more important consideration. This hypothesis is in agreement with the conclusions of Amundson et al. (1996), who suggested that atmospheric circulation, not average temperature, determined the $\delta^{18}O$ of precipitation in the eastern Rocky Mountains of Wyoming during the Pleistocene.

Unlike the Oligocene and Miocene samples, Eocene smectites from central Wyoming have $\delta^{18}O$ values that are higher than would be expected for their current elevation. There are several possible explanations for the trends observed in the Eocene data from central Wyoming: (1) these smectites formed at lower elevations and have been uplifted prior to Oligocene time; (2) the central Wyoming smectites formed from evaporative waters during the Eocene; (3) these smectites formed at warmer temperatures; and/or (4) the precipitation in central Wyoming during the Eocene had higher $\delta^{18}O$ values than modern precipitation. Our hydrogen and oxygen isotopic data show no evidence for increased evaporation in the basins of central Wyoming during the Eocene, and there is no compelling evidence for uplift of the basins during the Oligocene. We prefer the interpretation that the relatively high $\delta^{18}O_{smectite}$ values are directly linked to the climatic history, because there were significant climatic changes during the Eocene that would influence the isotopic composition of precipitation in this region. We cannot, however, exclude tectonic modification between the time of deposition of our Eocene and Oligocene samples as a possible explanation of our results.

Previous paleoclimate studies suggest the average temperatures in the Eocene in North America were 10–20 °C warmer than present before a rapid, global cooling event in the early Oligocene (33.2 Ma), when average temperatures in North America fell to 2–4 °C warmer than present (e.g., Wolfe, 1978; Prothero and Heaton, 1996; Terry, 2001). In addition to temperature changes and the lack of significant ice sheets, differences in past oceanic circulation patterns could potentially have affected the stable isotopic composition of precipitation. During the Eocene, oceanic basins were not as restricted in the equatorial regions as they are presently. Because no land mass was extant between North and South America prior to the late Miocene–early Pliocene, seawater exchange took place between the Pacific and Atlantic basins. Additionally, prior to the Oligocene closing of the Tethys Seaway, the northern proto-Indian and northern Atlantic Oceans were open to circulation through the Mediterranean basin. In spite of these circulation differences, modeling of oceanic circulation for the early Eocene Pacific Ocean suggests circulation patterns that closely resembled present-day circulation (Bice et al., 2000). Mean annual temperatures for surface waters in the early Eocene North Pacific (30°–60°N latitude) are thought to have been ~1–2 °C warmer than present, and the global averaged mean annual atmosphere-ocean moisture flux has been calculated to be 10% greater than present, with most of the increase in tropical latitudes (Bice et al., 2000). Based on modeled early Eocene salinity, Bice et al. (2000) predicted the $\delta^{18}O$ value of early Eocene surface water in the North Pacific to be −1.0‰ to −1.2‰, or ~0.5‰ to 1.5‰ lighter than present-day values. Using this difference in $\delta^{18}O$ values for surface waters in the North Pacific for the early Eocene, annual average temperatures for the interior of North America 10–20 °C warmer in the Eocene, and the relationship between average annual temperature and $\delta^{18}O_{precipitation}$ (Gat, 1980), we predict the Eocene $\delta^{18}O_{precipitation}$ in western Nebraska to be approximately −5‰ to 0‰, or 5‰ to 10‰ heavier than present $\delta^{18}O_{precipitation}$ values. Assuming this value for $\delta^{18}O_{precipitation}$ and the fractionation factor for smectite of Savin and Lee (1988) ($1000\ln \alpha_{smectite\text{-}water} = 2.58 \times 10^6 T^{-2} - 4.19$), we would expect that smectite formed in the Eocene throughout our study area would be 4‰–6‰ higher than modern smectites. In addition, general circulation models for the Late Tertiary suggest that as a result of a relatively warm climate, the monsoonal effects were intensified along the east side of the Rocky Mountains (Huber and Sloan, 1999; Sewall et al., 2000). This would add to the increase in the $\delta^{18}O$ values of precipitation in central Wyoming relative to today, but the magnitude of this effect is difficult to quantify. Since these effects can cause the isotopic shift observed in our Eocene samples relative to modern from the eastern, low-elevation part of our study area, it is not necessary to invoke tectonics as the cause of the relatively high $\delta^{18}O$ values of these samples, although we do not rule out Eocene uplift as a possible explanation for our results. The preceding discussion does not explain our results for samples from western Wyoming, because these Eocene smectites are not isotopically heavier than our modeled modern smectites. We propose these relatively light isotopic values in western Wyoming may be the result of cooler than global average temperatures in this area (rather unlikely without high elevations), a relatively large amount of isotopically light winter precipitation in the early Cenozoic Rocky Mountain region, which has been suggested by Norris et al. (1996) and Dettman and Lohmann (2000), uplift or topographically high elevations in western Wyoming in the Eocene, or isotopic rainout effects due to a topographic barrier to the west of western Wyoming. All of these possibilities suggest high elevations or topography in the Early Tertiary Rocky Mountains, which is consistent with other isotopic studies from the region (Norris et al., 1996; Dettman and Lohmann, 2000; Fricke, 2003). We cannot, however, determine from the isotopic data from our Eocene samples if tectonic or climate modification caused the shift in isotopic gradient between our 50–49 Ma and 37 Ma samples.

CONCLUSIONS

We conclude that an isotopic gradient, which suggests areas of high elevation, has existed in the Rocky Mountain–Great Plains region since at least 37 Ma. This conclusion is similar to those of previous isotopic studies of older rocks in the area (e.g., Dettman and Lohmann, 2000; Fricke, 2003). Because this isotopic gradient reflects the unique climatic regime that results from the interaction of three air masses with the high topography of the Rocky Mountains, we suggest it is very unlikely that high elevations in the Rocky Mountains could have formed in the late Miocene without significant modification of the atmospheric circulation patterns of western North America and the resulting distribution of pre–late Miocene $\delta^{18}O_{precipitation}$ values compared to the modern distribution of $\delta^{18}O_{precipitation}$. If there were not a topographic barrier before the late Miocene, we predict unobstructed precipitation originating over the Pacific Ocean would have extended deep into the continental interior of North America. Additionally, with no orographic rainout effects due to interaction with a topographic barrier, the oxygen isotope composition of precipitation in the western part of our study area would have been relatively enriched in ^{18}O. Therefore, we would expect the west-to-east gradient of $\delta^{18}O_{smectite}$ to be shallower and $\delta^{18}O_{precipitation}$ to decrease from west to east in western Wyoming if there were no pre–late Miocene Rocky Mountains. Our $\delta^{18}O_{smectite}$ results, however, increase from west to east, which suggests that pre–late Miocene high-elevation areas existed in the Rocky Mountain region, and for the Oligocene and Miocene samples, the values mimic the modern distribution and values of $\delta^{18}O_{precipitation}$. We propose that this indicates Oligocene and later high elevation in the Rocky Mountains (in other words, an elevation gradient between the Rocky Mountains and the western Great Plains) approximately equal to modern elevations.

Results from our Eocene samples can be explained by climate and/or tectonic modification since the time of deposition for these samples, but both of these explanations suggest high elevations in the Rocky Mountain region ca. 50 Ma. Our results do not agree with the traditional view that the surface uplift of the Rocky Mountains occurred in the late Miocene and early Pliocene. There is additional evidence that the Basin and Range and Sierra Nevada Mountains were at high elevations during much of the Tertiary as well (Wolfe et al., 1997; Wolfe et al., 1998; Poage and Chamberlain, 2001). The idea of late Cenozoic surface uplift in many areas of the American west should be reconsidered, because evidence used to support this view may be the result of changes in global temperature rather than changes in topography, a concept originally proposed by Molnar and England (1990).

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