Investigation of early Eocene water-vapor transport and paleoelevation using oxygen isotope data from geographically widespread mammal remains

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ABSTRACT

The oxygen isotope composition ($\delta^{18}O$) of apatite from mammalian tooth enamel can be used to infer the $\delta^{18}O$ value of ingested water, which is in turn related to that of precipitation stored in surface reservoirs. Therefore, the $\delta^{18}O$ value of phosphate from fossil tooth enamel can be used to infer the $\delta^{18}O$ value of these reservoirs in the past. In this paper, tooth enamel from a semiaquatic mammal taxon (*Coryphodon*) collected from five early Eocene localities in North America is used to construct patterns in $\delta^{18}O$ values of river water for this time period.

At all localities, the $\delta^{18}O$ value of river water ($\delta^{18}O_r$) is estimated to have been higher during the early Eocene relative to present-day North American rivers, although the $\delta^{18}O$ vs. latitude gradient was shallower during the Eocene. Higher $\delta^{18}O_r$ values are consistent with warmer Eocene air masses being able to hold more water vapor and with an increase in the poleward transport of both moisture and latent heat. The regular decrease in $\delta^{18}O_r$ with latitude indicates that global atmospheric circulation patterns and hydrological transport were not much different from those of the present, although the shallower $\delta^{18}O$ vs. latitude gradient during the Eocene may reflect regional differences in precipitation, evaporation, and river recharge.

At a more regional scale, the $\delta^{18}O$ value of river water can provide insight into topographic relief during the early Eocene. In the case of intermontane basins of Wyoming, differences in average $\delta^{18}O_r$ values between basins indicate that Laramide mountain relief was on the order of 475 m. It is suggested that anomalously low $\delta^{18}O_r$ values reported previously do not provide unambiguous evidence for permanent snow at higher elevations and may instead reflect brief episodes of cooler winters and/or altered atmospheric circulation patterns.

Keywords: oxygen isotopes, paleoelevation, paleohydrology, Eocene, climate.

INTRODUCTION

The way in which oxygen isotope compositions ($\delta^{18}O$) of modern precipitation vary over the surface of the Earth has been a focus of research for almost 40 yr and has led to an increased understanding of the link between oxygen isotope fractionation and hydrological processes. The primary reason for such a link is the preferential incorporation of ^{18}O into condensate as water is precipitated and removed from cooling air masses. As more precipitation is removed from an air mass, the $\delta^{18}O$ value of the remaining vapor becomes progressively lower. Resulting patterns in the $\delta^{18}O$ value of precipitation ($\delta^{18}O_{pt}$) include a regular decrease in the $\delta^{18}O_{pt}$ value as air masses cool while rising over mountains, moving away from coastal areas, or moving from tropical source areas to polar sinks (e.g., Dansgaard, 1964; Epstein and Mayeda, 1953; Rozanski et al., 1993; Gat, 1996). In tropical regions, where vertical convection results in cooling, a correlation is also observed between the amount of precipitation and $\delta^{18}O_{pt}$ (e.g., Dansgaard, 1964; Araguas-Araguas et al., 1998).

Relationships between $\delta^{18}O_{pt}$ and these variables hold tremendous potential for the study of past terrestrial environments. If spatial patterns in $\delta^{18}O$ of ancient precipitation can be reconstructed, then it is possible to constrain geologically important events such as the timing, height, and uplift rates of mountains (e.g., Norris et al., 1996; Chamberlain et al., 1999; Dettman and Lohmann, 2000; Garzione et al., 2000; Rowley et al., 2001; Poage and Chamberlain, 2002), the onset of monsoonal atmospheric circulation patterns (e.g., Stern et al., 1997), and the global transport of heat and water vapor (e.g., Yapp, 1998; White et al., 2001). In addition, past patterns of $\delta^{18}O_{pt}$ can be used to test the predictions of global climate models that include oxygen isotope tracers in their simulations (e.g., Joussaume et al., 1984; Joussaume and Jouzel, 1993; Gedzelman and Arnold, 1994; Jouzel et al., 1994).

To estimate the $\delta^{18}O$ value of ancient precipitation, it is necessary to measure the $\delta^{18}O$ value of a material that forms in isotopic equilibrium with surface water and then preserves this primary $\delta^{18}O$ value over geologic time. If the temperature of formation and the isotope-fractionation equations are known, then it is possible to calculate the isotopic composition of the ancient water by using such proxy materials. These conditions are met in various degrees by a wide variety of minerals and organic material that form at the surface of the Earth (Table 1). In this paper, oxygen isotope compositions of biogenic phosphate ($\delta^{18}O_{bp}$) from mammalian tooth enamel are used to infer the $\delta^{18}O$ value of the surface water ingested by the mammal. More specifically, the $\delta^{18}O_{bp}$ values of fossil remains of the semiaquatic mammal *Coryphodon* are used to infer the $\delta^{18}O$ values of the Eocene rivers in which the individuals lived, and both local and regional patterns in the $\delta^{18}O$ values of precipitation over North America are reconstructed. These patterns provide information regarding vapor transport and the hydrologic cycle during this warm time period, the $\delta^{18}O$ values of Eocene ocean waters in high latitudes, and the height of Laramide mountain ranges.

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TABLE 1. COMMON PRE-QUATERNARY PROXIES FOR OXYGEN ISOTOPE RATIOS OF PRECIPITATION

Material	Occurrence	Precipitation reservoir
Carbonate	Lakes	Lake water
Clays	Paleosols	Soil/ground waters
Carbonate	(Arid environments)	
Siderite	(Wet environments)	
Hydroxide	(Wet/organic-rich)	
Oxides	(Arid/organic rich)	
Biogenic aragonite	Invertebrate shells	River water
Biogenic apatite	Vertebrate skeletons	Animal body water

Note: None of these materials forms in direct equilibration with precipitation; rather, they form from precipitation stored in different surface reservoirs. Excluding biogenic apatite from homeothermic vertebrates, all of these materials form over a potentially wide range of temperatures that may vary seasonally or from year to year. Paleosol minerals may take thousands or more years to form, lakes can precipitate minerals on a yearly basis, and biogenic minerals form over some part of the lifetime of the animal.

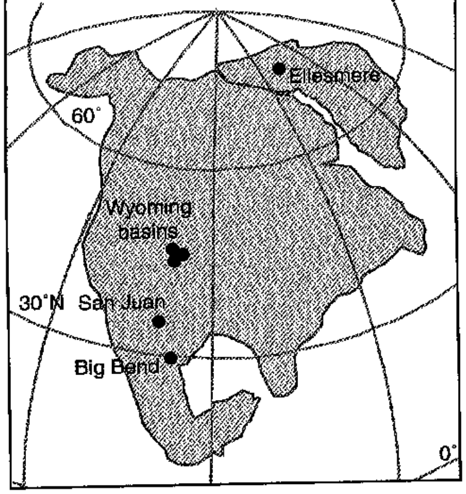

Figure 1. Paleogeographic map of North America during the early Eocene. Sample localities are shown by circles. The Mississippi embayment extended ~5° latitude farther north during the Eocene relative to the present day.

BIOGENIC APATITE AND $\delta^{18}O$ VALUE OF SURFACE WATERS

$\delta^{18}O$ values of biogenic phosphate can be used to estimate $\delta^{18}O$ values of local precipitation because water ingested by animals from surface reservoirs of precipitation such as streams, ponds, or plants plays a major role in determining the $\delta^{18}O$ value of their body water and, hence, the $\delta^{18}O$ value of biogenic phosphate in the skeletal apatite that forms from body water (Longinelli, 1984; Luz and Kolodny, 1985; Bryant and Froelich, 1995; Kohn, 1996). This relationship between $\delta^{18}O_{bp}$ and $\delta^{18}O$ of ingested waters is not one-to-one for land-dwelling animals owing to the incorporation of additional oxygen from organic matter and the atmosphere, but it can be quantified by using physiological models that account for the fluxes and fractionations of oxygen during the formation of biogenic phosphate from ingested waters (Bryant and Froelich, 1995; Kohn, 1996).

Of the different types of biogenic phosphate, mammalian tooth enamel has several characteristics that make it one of the best proxies available for estimating of $\delta^{18}O$ of surface water. Mammals are homeothermic; thus skeletal apatite forms at a constant temperature of ~37 °C. Therefore, any change in $\delta^{18}O_{bp}$ over time or space should reflect a change only in the $\delta^{18}O$ of ingested water. This constant, known temperature of formation of skeletal apatite contrasts with that of minerals forming in pedogenic, lacustrine, or fluvial environments where temperatures are variable and not known precisely (Table 1), which is important because oxygen isotope fractionation during mineral formation can vary significantly with temperature and must be known to make precise estimates of $\delta^{18}O$ of ancient surface waters. Another advantage of biogenic phosphate is its resistance to oxygen isotope exchange during diagenesis, particularly when the phosphate occurs as part of the large and tightly packed apatite crystals in mammalian tooth enamel (Lee-Thorpe and Van der Merwe, 1991; Ayliffe et al., 1994). Finally, mammalian tooth enamel is ideal for reconstructing past patterns in $\delta^{18}O_{pt}$ because it is common throughout the Cenozoic time period and over broad geographic regions.

Like all minerals used in principle to infer $\delta^{18}O$ of past precipitation, the greatest difficulty in using biogenic phosphate is the fact that surface water ingested by mammals may be isotopically modified relative to local precipitation (Table 1). Small ponds and streams may hold local precipitation with little isotopic modification, but larger lakes and soil waters, especially in arid regions, may undergo evaporation that modifies their oxygen isotope composition. Similarly, $\delta^{18}O$ values of leaf water ingested by herbivorous mammals can be shifted to higher values relative to precipitation as a result of evaporation at the surface of the leaf (Sternberg, 1989), and the influence of evaporation can then be observed in $\delta^{18}O_{bp}$ (Luz et al., 1990). Another factor to be considered is the homogenization of precipitation that occurs during the formation of lakes, soil waters and groundwaters, and larger rivers. These larger bodies of water may incorporate precipitation that formed over a wide area or over a long period of time. Given the limited spatial and temporal resolution of sampling inherent to this study of early Eocene localities, however, such a homogenization is not a major concern.

SAMPLES

This study relies on the isotopic analysis of tooth enamel from *Coryphodon* (Mammalia, Pantodonta), a large terrestrial mammal that was common throughout the entire Holarctic region during the Paleogene (Lucas, 1998). Samples have been collected from lower Eocene rocks at localities in North America that span a large latitudinal range including Big Bend, Texas; the San Juan Basin of New Mexico; the Green River, Big Horn, and Powder River Basins of Wyoming; and Ellesmere Island in Arctic Canada (Fig. 1; Table 2). All localities date from the Wasatchian (Wa) North American Land Mammal Age (ca. 55.0 to 50.0 Ma; Lucas, 1998). The Bighorn and Green River Basin samples can be given more precise age estimates within mammalian faunal zones Wa-6 (ca. 53 to 52.4 Ma), whereas fossil assemblages from the Powder River Basin are restricted in age to mammalian faunal zone Wa-2 (ca. 54 to 54.5 Ma; Clyde et al., 1994).

Except for the very earliest and latest Wasatchian, marine oxygen isotope data indicate that the $\delta^{18}O$ value of the early Eocene oceans varied by only ~0.25‰ and the oceans' temperature varied by ~1 °C during this time period (Zachos et al., 1994; Clyde et al., 2001). A large and brief $\delta^{18}O$ excursion at the beginning of the Wasatchian (Wa-0) is associated with a unique fauna and is unlikely to have been sampled unexpectedly, and the $\delta^{18}O$/temperature shift at the end of the Wasatchian is only an additional ~0.25‰ and ~1 °C decrease relative to rest of the time period. Given this relative isotopic and climatic stability, it is assumed that no major error is introduced by comparing isotope data from these different localities and that they can be combined to represent the early Eocene in general.

Paleontological evidence suggests that *Cor-*

TABLE 2. SAMPLE LOCATION, PALEOLATITUDE, AGE, $\delta^{18}O_{bp}$, AND ESTIMATED $\delta^{18}O_r$

Locality	Paleolat. (°N)	Tooth (position)	Age (Wa-biozone)	$\delta^{18}O_{bp}$ (‰)	$\delta^{18}O_r$ (‰)
Big Bend	29		Wa-6		
Cory 1		?		19.9	−0.1
Cory 2		Molar		18.6	−1.8
Cory 3		?		20.6	0.9
Cory 4		Molar		20.1	0.2
Cory 5		Molar		18.8	−1.5
San Juan Basin	36		Wa-6		
Cory 1		?		17.8	−2.8
Cory 2		?		18.4	−2.0
Cory 3		Molar		18.2	−2.3
Cory 4		?		17.4	−3.3
Cory 5		Molar		18.2	−2.3
Cory 6				18.6	−1.8
Cory 7–1		Incisor		17.9	−2.7
Cory 7–2				17.1	−3.7
Cory 7–3				17.8	−2.8
Cory 7–4				17.1	−3.7
Cory 7–5				17.0	−3.9
Cory 7–6				17.2	−3.6
Cory 7–7				18.1	−2.4
Green River Basin	44		Wa-6		
Cory 1*		Molar		15.9	−5.3
Cory 2*		Molar		15.4	−6.0
Cory 3*		Molar		18.2	−2.3
Bighorn Basin	45		Wa-6		
Cory 1†		Canine		14.7	−6.9
Cory 2†		Canine		13.2	−8.9
Cory 3†		Canine		13.2	−8.9
Cory 1		Molar		12.8	−9.4
Cory 2		Incisor		14.7	−6.9
Cory 3		?		13.8	−8.1
Powder River Basin	45		Wa-2		
Cory 1‡		Canine		15.2	−6.2
Cory 2‡		Molar		14.8	−6.8
Cory 3‡		Molar		13.8	−8.1
Ellesmere Island	73		Wasatchian		
Cory 1–1		Incisor		6.1	−18.2
Cory 1–2				5.9	−18.5
Cory 1–3				5.3	−19.3
Cory 1–4				5.1	−19.5
Cory 1–5				6.5	−17.7
Cory 1–6				6.0	−18.3
Cory 1–7				6.2	−18.1
Cory 2–1		Incisor		7.5	−16.4
Cory 2–2				5.6	−18.9
Cory 2–4				6.3	−17.9
Cory 2–5				7.3	−16.6
Cory 2–6				4.8	−19.9
Cory 2–7				5.3	−19.2
Cory 3		Incisor		4.4	−20.4
Cory 4		Incisor		4.5	−20.3
Cory 5		Incisor	Outlier	2.8	−22.6
Cory 6		Molar		6.0	−18.3

Notes: Approximate paleolatitudes of each locality are from Scotese (1999). All samples are from fragments of tooth enamel except in the case in which additional numbers represent the position of multiple samples taken along the length of single teeth. When possible, the tooth position is identified. Ages are given in mammalian biozones of the Wasatchian North American Land Mammal Age, and in millions of years when possible. $\delta^{18}O_r$ is estimated by using the physiological model for mammalian herbivores of Kohn (1996).

†Median corrected $\delta^{18}O_r$ value of intra-tooth data from Fricke et al. (1998a).

‡$\delta^{18}O_{bp}$ values obtained by using high-temperature reduction. Outliers not included in the computation of averages are noted.

yphodon was herbivorous and most likely lived near, or perhaps in, rivers that flowed through Paleogene basins. As discussed in Fricke et al. (1998a), the sources of water ingested by *Coryphodon* may have included river water and water from stems and leaves of terrestrial and aquatic plants. Of these, only water from leaves has the potential to be significantly different from river water because evaporation at the leaf surface can result in shifts in $\delta^{18}O$ of leaf water (Sternberg, 1989). This effect, however, would most likely have been muted for aquatic plants and for terrestrial plants living in humid early Eocene environments characterized by significant amounts of precipitation (e.g., Wilf et al., 1998; Wilf, 2000), and a study of modern hippopotamuses indicates that this semiaquatic taxon ingested water with lower $\delta^{18}O$ values compared to other, nonsemiaquatic vertebrate taxa (Bocherens et al., 1996). Given the likely habitat occupied by *Coryphodon*, the $\delta^{18}O$ value of its tooth enamel is best considered a proxy for the $\delta^{18}O$ value of river water ($\delta^{18}O_r$) rather than a direct reflection of the $\delta^{18}O$ value of precipitation.

A final factor to consider when sampling mammalian tooth enamel is the incremental formation of enamel over time scales of months that can capture much of any seasonal variations in $\delta^{18}O$ of ingested water (e.g., Koch et al., 1989; Fricke and O'Neil, 1996; Fricke et al., 1998b). For example, the oxygen isotope compositions of precipitation and, hence, rivers often vary seasonally (Rozanski et al., 1993; Coplen and Kendall, 2000), and in the case of the Bighorn Basin, isotopic variations of 1‰ to 4‰ have been observed along the length of single *Coryphodon* teeth (Fricke et al., 1998a). To reduce any possible sampling bias when multiple samples were taken from one tooth, median $\delta^{18}O_{bp}$ values are used when comparing data between teeth and/or localities. For the majority of localities, only fragments of teeth were available for analysis, and only bulk samples of tooth enamel were taken. Fortunately, statistical studies indicate that seasonal variations in $\delta^{18}O_{bp}$ are reflected by the isotopic variability between multiple bulk samples from a single locality (Clementz and Koch, 2001), and therefore average $\delta^{18}O_{bp}$ values of the multiple samples are used when comparing data from different localities.

RESULTS

Oxygen isotope compositions of biogenic phosphate were measured by two different methods. One is the technique of O'Neil et al. (1994) that involves the isolation of the phosphate radical as Ag_3PO_4, the reaction of this material with graphite at 1400 °C in a sealed quartz tube to form CO_2, and the introduction of this gas to the inlet system of a mass spectrometer. A second method involves the reduction of Ag_3PO_4 in a graphite furnace at 1400 °C to produce CO that is then introduced by means of continuous-flow techniques into a mass spectrometer (Kornexl et al., 1999; Vennemann et al., 2002). A rigorous comparison of these methods (Venneman et al., 2002) reveals that the O'Neil et al. (1994) method is characterized by a scale compression, thus necessitating a correction of raw data. This correction was not made for data originally published by Fricke et al. (1998a), but has been applied to data presented in Table 2.

Physiological models used to relate $\delta^{18}O_{bp}$ to $\delta^{18}O$ of ingested water have been construct-ed by estimating the fluxes and fractionations of oxygen moving into and out of the bodies of mammals having a constant body temperature (Longinelli, 1984; Luz and Kolodny, 1985; Bryant and Froelich, 1995; Kohn, 1996). These models indicate that although there is a linear relationship between $\delta^{18}O_{bp}$ and $\delta^{18}O$ of ingested water, it is not a one-to-one relationship owing to the influence of oxygen from ingested organic matter and from the atmosphere on the $\delta^{18}O$ value of body water. Furthermore, the intercepts of the relationship will vary slightly for different animals depending on the fluxes of H_2O, O_2, and CO_2 into and out of their bodies. In this paper, the model of Kohn (1996) for mammalian herbivores is used: $\delta^{18}O_{bp} = 0.76 \times \delta^{18}O_{ingested\ water} + 19.94$ (if a relative humidity of 75% is assumed). Estimates of $\delta^{18}O$ of river water for each locality are given in Table 2 and have an uncertainty of ± 1.39‰ that is due to small taxonomic differences in physiology, diet, and behavior (Kohn, 1996). Most error introduced is related to absolute estimates of $\delta^{18}O$ using data from a single mammalian taxon. For any given taxon, however, physiology, diet, and behavior are not likely to vary significantly, and *differences* in estimated $\delta^{18}O$ values of river water between localities (i.e., $\delta^{18}O$ vs. latitude gradients) are likely to have much smaller errors associated with them.

Isotopic offsets between *Coryphodon* and another ecologically and physiologically distinct taxon, such as a freshwater fish, provide test of whether $\delta^{18}O_{bp}$ values have been altered during diagenesis (Fricke et al., 1998a; Barrick et al., 1999; Fricke and Rogers, 2000). Such isotopic offsets between taxa would not occur if isotopic alteration was extensive, as isotopic exchange with groundwaters or secondary precipitation of apatite during diagenesis would result in uniform $\delta^{18}O$ values of phosphate. Systematic offsets of ~1.5% between biogenic phosphate from *Coryphodon* and biogenic phosphate from associated fish scales are observed from the Bighorn Basin (Fricke et al., 1998a).

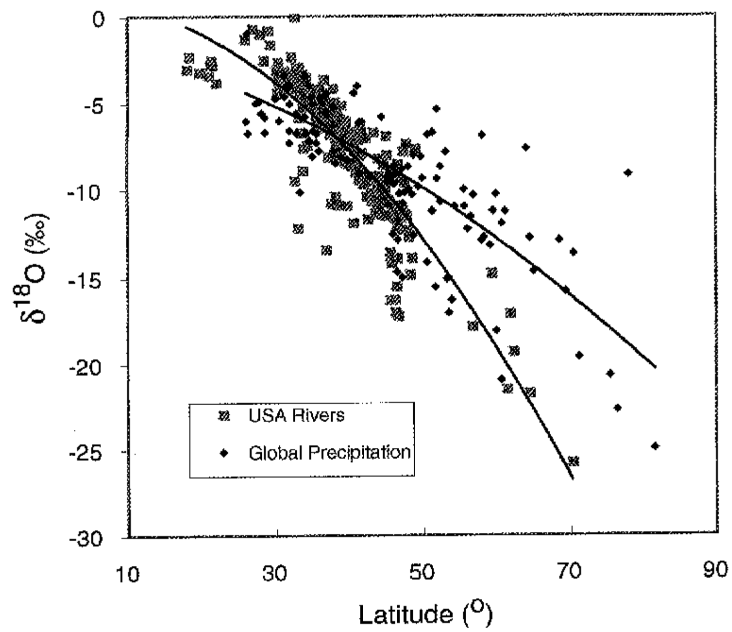

Figure 2. Oxygen isotope compositions of global precipitation (diamonds) and United States rivers under 300 m in elevation (squares) vs. latitude for the present day (Rozanski et al., 1993; Coplen and Kendall, 2000). There is a regular decrease in $\delta^{18}O$ values with increasing latitude because of the preferred incorporation of ^{18}O into precipitation as air masses cool from tropical to polar regions. A steeper gradient for rivers most likely reflects evaporation at lower latitudes and seasonal recharge at higher latitudes.

DISCUSSION

Paleohydrology

Global-Scale Patterns in $\delta^{18}O$ and Paleohydrology

Variations in the $\delta^{18}O$ value of river water at the global scale during the early Eocene can provide insight into variations in the hydrologic cycle (e.g., water fluxes and circulation patterns) under different climatic conditions. Global isotope patterns and hydrology are linked by the preferential incorporation of ^{18}O into condensate during adiabatic processes of cooling and a subsequent lowering of the $^{18}O/^{16}O$ ratios in vapor as air masses move along surface-temperature gradients (1) from tropical to polar latitudes or (2) from low to high elevations (Fig. 2). Models of this isotopic distillation indicate that warmer air masses have higher initial $\delta^{18}O_{pt}$ values compared to cooler air masses because the warmer air is able to hold more moisture and thus retain a larger fraction of ^{18}O (e.g., Dansgaard, 1964; Rowley et al., 2001). As a result of these relationships, patterns of $\delta^{18}O$ for the early Eocene can be used to help infer rainout histories of air masses and thus the distribution of atmospheric water vapor at that time.

Because *Coryphodon* is assumed to have ingested river water, however, it is necessary to understand how $\delta^{18}O$ of river water is related to $\delta^{18}O$ of local precipitation. One important difference is the steeper $\delta^{18}O$ vs. latitude gradient observed for North American rivers relative to the global precipitation network (Fig. 2). A plausible explanation for why surface waters have high $\delta^{18}O$ values relative to $\delta^{18}O_{pt}$ at latitudes below ~35°N is the intense evaporation associated with high-pressure belts. These regions are characterized by a mean annual evaporative input (E) of water from the surface to the atmosphere that is greater than water loss from air masses via precipitation (P; i.e., $P - E < 0$; Peixoto and Oort, 1992). This circumstance leads to a net transfer of lighter ^{16}O to the atmosphere, thus increasing the $\delta^{18}O$ value of surface waters. In contrast, precipitation exceeds evaporation in middle latitudes ($P - E > 0$; Peixoto and Oort, 1992), and a likely reason for rivers having lower $\delta^{18}O$ values, compared to precipitation, at higher latitudes is a seasonal bias in river recharge, with isotopically heavier summer precipitation being quickly returned to the atmosphere by means of plant transpiration, while isotopically lighter winter and spring precipitation is able to infiltrate and recharge groundwater and surface waters.

Paleohydrology of North America

The early Eocene was characterized by a regular decrease in $\delta^{18}O_r$ with latitude and thus was broadly similar to the present day (Fig. 3). This similarity indicates that atmospheric circulation patterns influencing air-mass movement and cooling (e.g., Hadley circulation cells) were not very different, despite records from eolian sedimentary deposits that suggest globally weaker winds and less in-

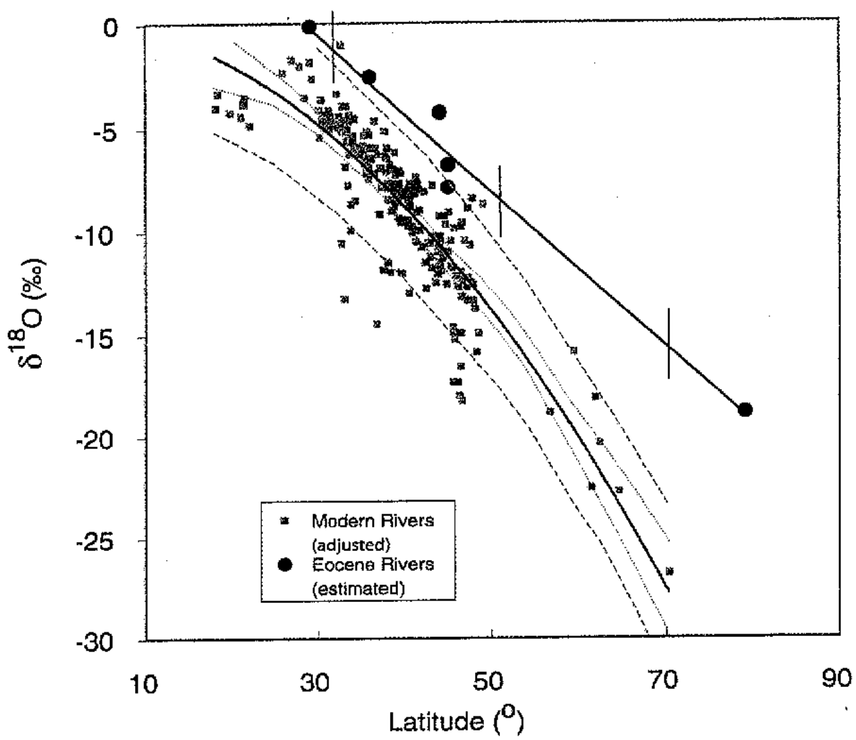

Figure 3. Oxygen isotope compositions of modern rivers (squares) vs. those estimated for the early Eocene (circles). To facilitate comparison with pre–ice-sheet time periods, modern data have been shifted by 1‰ to lower values. For modern data, short-dashed black curves are 95% confidence intervals around regression lines, and long-dashed curves are 95% confidence intervals around the data set. For Eocene data, solid vertical lines represent uncertainty of the position of the regression line associated with the physiological model utilized. The slope of the regression line for Eocene rivers and the $\delta^{18}O$ values themselves are significantly different from the present situation. Higher $\delta^{18}O$ values and a shallower $\delta^{18}O$ vs. latitude gradient indicate that more moisture was being transported from tropical to polar latitudes during the Eocene.

tense atmospheric circulation during the Eocene (e.g., Rea, 1998). Eolian records have also been used to infer a position of the Intertropical Convergence Zone (ITCZ) north of the present position by ~20° latitude during the Eocene (Rea et al., 2000). The data presented here cannot address this possible "expansion" of the tropics, but any such movement of the ITCZ does not appear to have altered extratropical circulation patterns to a large degree.

What is noticeably different between the Eocene and the present are the significantly higher Eocene $\delta^{18}O_r$ estimates. Because Eocene oceans had $\delta^{18}O$ values ~1‰ lower than at present (Miller et al., 1987), $\delta^{18}O$ values of precipitation and therefore the rivers' isotopic patterns would have been offset by ~1‰ (Fig. 2). Taking this offset into account, the relative difference between modern and Eocene $\delta^{18}O_r$ values is even greater (Fig. 3). The difference in $\delta^{18}O_r$ is greatest at higher latitudes, and the result is that a shallower $\delta^{18}O_r$ vs. latitude gradient is also observed.

Higher $\delta^{18}O$ values in the subtropics indicate that there was an increase in evaporation and net flux of water into the atmosphere; i.e., a larger proportion of moisture and thus of ^{18}O was retained in air masses as they began to move poleward. This interpretation is consistent with warmer subtropical temperatures during the Eocene (e.g., Greenwood and Wing, 1995; Markwick, 1995) and the increased capacity of warmer air masses to retain moisture.

A shallower $\delta^{18}O_r$ vs. latitude gradient at higher latitudes during the Eocene implies that a smaller proportion of precipitation was lost from air masses as they continued to cool and move out of the tropics. Relative to the present-day situation, the Eocene would have been characterized by a net increase in the transportation of water vapor to higher latitudes. Another possible explanation for this shallower $\delta^{18}O_r$ vs. latitude gradient is a less pronounced seasonal discrimination in river recharge at higher latitudes, where summer as well as winter precipitation ultimately is incorporated into rivers. Although both of these factors may play a role, there is paleobotanical evidence to support the hypothesis that in the Eocene, more water vapor was transported to higher latitudes where it condensed as precipitation. In general, humid, subtropical forests dominated the Eocene of North America, and leaf-area analyses of megafloras indicate that mean annual precipitation in the Bighorn and Green River Basins was on the order of 120–150 cm during this time (Wilf et al., 1998; Wilf, 2000). This amount is similar to the amount of rainfall received near the present-day Gulf Coast of North America. In polar regions, floral remains from Ellesmere Island imply high amounts of rainfall similar to those of the present-day Pacific Northwest (Kalgutkar and McIntyre, 1991).

At a more regional hydrologic scale, a comparison of adjacent Laramide basins from Wyoming reveals higher $\delta^{18}O_r$ values in the Green River Basin relative to the other basins (−3.9‰ vs. −7.0‰ and −8.1‰ for the Powder River and Big Horn Basins, respectively). These higher values most likely represent enhanced evaporation in the Green River Basin and thus signal an increase in aridity that culminated in the formation of evaporites in this basin during the late Eocene.

Vapor Transport and Global Warming

A persistent problem in trying to understanding global warmth during the Eocene has been explaining why paleotemperature records from polar regions indicate that warming relative to the present day was greatest in polar relative to tropical regions (e.g., Zachos et al., 1994). In general, changes in the concentration of atmospheric CO_2 (pCO_2) are considered to play a major role in warming. Because of the uniform distribution of CO_2 in the atmosphere, however, changes in pCO_2 should have led to a uniform warming at all latitudes and cannot account for a shallower latitudinal temperature gradient (Sloan et al., 1995). One possible means of focusing warming at the high latitudes is the efficient removal and transfer of latent heat in water vapor from tropics to higher latitudes by atmospheric circulation (Sloan et al., 1995). It has also been suggested that stratospheric clouds form preferentially in polar regions, thus increasing heat retention in these areas relative to the tropics (Sloan and Pollard, 1998).

Both of these mechanisms for high-latitude warming depend on the enhanced movement of atmospheric water vapor toward the poles and are supported by the oxygen isotope data presented here. Higher $\delta^{18}O_r$ values and a shallower Eocene $\delta^{18}O_r$ vs. latitude gradient provide direct evidence for more transport of atmospheric water vapor and, hence, latent heat to middle and higher latitudes from the tropics. In addition, modern observations indicate that higher concentrations of water vapor in the polar troposphere can "leak" into the overlying stratosphere (Kirk-Davidoff et al., 1999), thus providing the moisture necessary for cloud formation in these regions. Overall, these results demonstrate that estimates of past $\delta^{18}O_r$ values represent a valuable tool in accounting for the transport of atmospheric water vapor in the past, and future comparisons of isotopic data with global climate model simulations should prove profitable in quantifying the role of water as a cause of climate change during the Eocene.

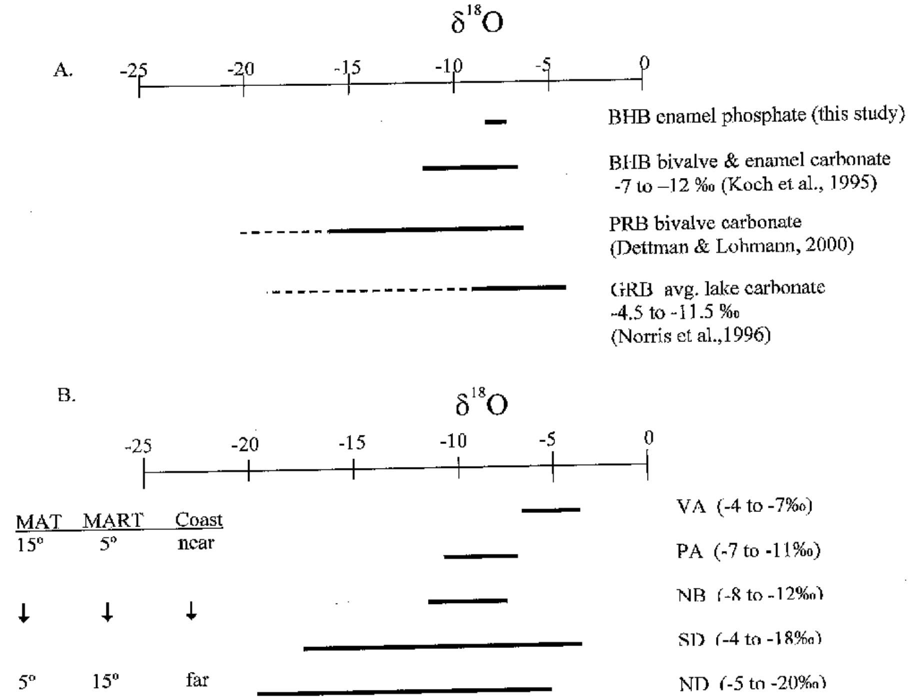

Figure 4. (A) $\delta^{18}O$ values of precipitation estimated by using mammalian tooth enamel, riverine bivalve shells, and lake carbonates from the Bighorn Basin (BHB), Green River Basin (GRB), and the Powder River Basin (PRB). All samples are Eocene age (55 to 53.7 Ma) except Green River Basin lake carbonates, which are late Eocene. Dark lines illustrate ranges in $\delta^{18}O$ values for the majority of the data from each study, whereas dashed lines illustrate the range including the 10% of the data with exceptionally low values. The majority of the total data do not provide unequivocal evidence for permanent snow fields in high mountains. (B) Seasonal ranges in $\delta^{18}O$ values for selected U.S. watersheds with elevations under ~650 m. None of these rivers or their associated climate conditions is directly analogous to the Eocene Bighorn Basin; however, these data illustrate how the $\delta^{18}O$ values of river water have the potential to be quite variable and low in the absence of significant topographic relief within the watershed. Rivers include the following: Holiday Creek, Blackwater River, and New River from Virginia (VA); Young Womans Creek, Juniata River, and Susquehanna River from Pennsylvania (PA); Niobrara River from Nebraska (NB); White River, Cheyenne River, and Moreau River from South Dakota (SD); and Beaver Creek, Heart River, Cannonball River, and Knife River from North Dakota (ND). Data are from Coplen and Kendall (2000). MAT—mean annual temperature; MART—mean annual range in temperature.

Using $\delta^{18}O_r$ to Estimate Topographic Relief of Laramide Mountains

Turning from hydrology and climate, patterns in $\delta^{18}O$ values of precipitation and rivers can also be used to study paleotopographic relief in continental settings. This application takes advantage of the relationship between $\delta^{18}O$ and elevation that has its basis in the orographic cooling of air masses as they are forced over mountains and the preferential removal of ^{18}O into the resulting precipitation (for studies demonstrating this effect, see Table 1 in Chamberlain and Poage, 2000). The resulting relationship between $\delta^{18}O$ and elevation is termed the lapse rate (unit = ‰/100 m), and it can be applied to the difference in $\delta^{18}O$ between authigenic or biogenic minerals on the windward and leeward sides of a mountain range in order to estimate the amount of relief associated with that range (e.g., Norris et al., 1996, 2000; Chamberlain et al., 1999; Dettman and Lohmann, 2000; Garzione et al., 2000; Poage and Chamberlain, 2002). Theoretical models of isotopic distillation of precipitation from orographically uplifted air masses can also be used to study paleotopography (e.g., Rowley et al., 2001). Application of these of methods to Laramide basins of Wyoming has the potential to answer long-standing questions of timing of uplift and height of the Rocky Mountain region of North America.

Low $\delta^{18}O$ and High Elevation

In two previous studies, anomalously low $\delta^{18}O$ values of lake carbonates from the Green River Basin (Norris et al., 1996, 2000) and low $\delta^{18}O$ values of riverine bivalves from the Bighorn and Powder River Basins (Dettman and Lohmann, 2000) were used to infer the existence of permanent snow fields in the adjacent mountains. Because mean annual temperatures of the basins are such that snowfall in them was precluded (e.g., Wing et al., 2000), these authors concluded that low $\delta^{18}O_r$ values must reflect the addition of mountain snow melt with low $\delta^{18}O$ values. Norris et al. (1996, 2000) assumed that this snow melt came from permanent snow packs where mean annual temperatures were less than zero and then used a temperature vs. altitude lapse rate of 6 °C/km to estimate that mountain heights were 2.5 to 3 km above the basin floor. Dettman and Lohmann (2000) also assumed that snow melt came from permanent snow packs, in this case with very low average $\delta^{18}O$ values (less than −21‰), and then used a theoretically calculated $\delta^{18}O$ vs. elevation lapse rate of 4‰/km to estimate a similar amount of Laramide mountain relief.

Despite the consistency of these results, the equivocal nature of the majority of lake carbonate and bivalve data makes it worthwhile to reexamine the question of mountain relief during the Eocene. In the case of bivalves, all estimates of $\delta^{18}O_r$ from Bighorn basin rivers are between −7 and −12.2‰, and ~90% of

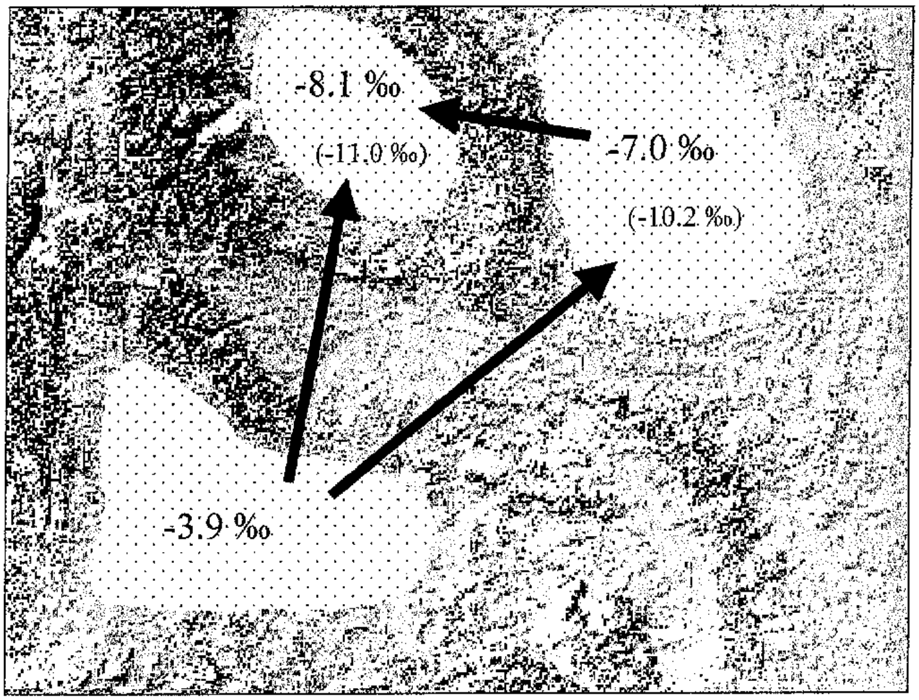

Figure 5. Detailed physiographic map of Wyoming showing the locations of the Bighorn Basin (avg. $\delta^{18}O_r = -8.2‰$), Powder River Basin (avg. $\delta^{18}O_r = -7.0‰$), and Green River Basin (avg. $\delta^{18}O_r = -3.9‰$). Estimates of $\delta^{18}O_r$ obtained from bivalve carbonate of the same age are shown in parentheses (Dettman and Lohmann, 2000). Arrows represent possible transport directions of air masses during the Eocene. Intervening topographic relief in the form of the Bighorn, Owl Creek, Wind River, and Casper Arch uplifts (BHU, OCU, WRU, and CAU respectively) would have caused air-mass rainout and a lowering of $\delta^{18}O_r$ values. The heights of these uplifts can be estimated by using the difference in $\delta^{18}O_r$ values between basins and an isotopic lapse rate (see text).

the Powder River basin deposits have average estimates of $\delta^{18}O_r$ above $-16‰$ (Fig 4A; Dettman and Lohmann, 2000). Considering temporal differences in $\delta^{18}O$ of ocean water, values between -7 and $-16‰$ during the Eocene are comparable to modern values of -6 to $-15‰$, and this range is extremely common in modern watersheds that are *not* associated with extreme topographic relief (Fig. 4B). Thus it is arguable that only ~10% of the bivalve data provide more conclusive evidence for permanent snow packs in mountain highlands. The percentage of lake carbonates with exceptionally low $\delta^{18}O$ values is similarly small (Norris et al., 1996; 2000). Both Norris et al. (1996, 2000) and Dettman and Lohmann (2000) attributed the rarity of low $\delta^{18}O_r$ values to hydrological complexities of large watersheds, in particular the mixing of snow melt with other water sources. Although the interpretations based on this assumption may indeed be correct, they are not the only interpretations possible.

Alternative Explanations for Low $\delta^{18}O$

An equally plausible explanation of the oxygen isotope data is that low $\delta^{18}O_r$ values reflect *episodic events* rather than *permanent features* and that mountains were not as high as estimated previously. For example, seasonal temperature variations could have resulted in the formation of winter precipitation with significantly lower than average $\delta^{18}O$ values in the *absence* of extreme relief in adjacent highlands. Mean annual temperature in the Bighorn Basin was ~16.7 °C during the Wasatchian (Wing et al., 2000), and cold monthly mean temperature was ~6.3°C (Greenwood and Wing, 1995). Applying the modern relationship between seasonal changes in temperature and seasonal changes in the $\delta^{18}O$ value of precipitation of 0.40‰/°C (Rozanski et al., 1993) to this seasonal difference in temperature [0.40‰/°C × (16.7–6.3 °C)], the $\delta^{18}O$ of winter precipitation can be estimated to have been ~4‰ lower than that of the annual mean. These seasonal extremes in $\delta^{18}O$ may have been extenuated periodically owing to changes in orbital parameters that may have decreased mean annual and cold monthly mean temperatures even further (e.g., Sloan and Morrill, 1998) and thus resulted in even lower $\delta^{18}O$ values of winter precipitation. In both cases, extreme topographic relief is not required to explain low $\delta^{18}O_r$ values.

An additional factor that may have resulted in episodically low $\delta^{18}O_r$ values is the introduction into Wyoming of air masses originating over the Pacific. Global climate model simulations of the modern situation indicate that reduced ice cover over the Arctic Ocean (as would have been common during the Eocene) strengthens the Arctic cyclone so that Pacific air masses are transported great distances overland at high latitudes before moving over Wyoming (Sewall, personal commun. 2002). At the present time, snow from such western and polar air masses can easily result in seasonal $\delta^{18}O_r$ values of lower than $-20‰$ in the absence of drastic differences in elevation within a watershed (Fig. 4B, South and North Dakota), and may also explain uncommonly low $\delta^{18}O_r$ values during the Eocene.

Interbasinal Comparisons of $\delta^{18}O$ and Topographic Relief

If the infrequent occurrence of low $\delta^{18}O$ values from a single basin cannot be used as *unambiguous* evidence for either permanent snow fields or episodic precipitation events, then an alternative approach of estimating the topographic relief of Laramide mountains is needed. One possibility is to determine the difference in average $\delta^{18}O_r$ values from basins lying on the windward and leeward sides of these mountains, and to apply an $\delta^{18}O$ vs. elevation lapse rate to estimate the relief associated with them. The Powder River and Bighorn Basins provide an excellent opportunity to apply this method to the Bighorn Mountains, as global climate simulations indicate that predominant wind directions over Eocene Wyoming during rainy seasons were from the east and southeast (Sewall et al., 2000), and no intermediate highlands stood between these areas and the source of air masses over the Mississippi embayment (Figs. 1 and 5).

Estimates from *Coryphodon* yield an interbasinal difference in average $\delta^{18}O_r$ values of 1.1‰, which is similar to the inter-basinal difference in *average* $\delta^{18}O_r$ values of 0.8‰ observed for bivalves (Dettman and Lohmann, 2000) of the same age (55 to 53.7 Ma). The fact that these dissimilar isotopic proxies are both recording a difference between basins of ~1‰ is encouraging and lends confidence that a rainout signal is being observed (Fig.

5). In applying a $\delta^{18}O$ vs. elevation lapse rate to this difference of 1‰, it must be considered that lapse rates vary regionally, ranging from 0.1‰/100 m to 0.4‰/100 m with a global average of 0.21‰/100 m (Chamberlain and Poage, 2000). Because of the limited amount of modern data, it is not clear which lapse rate to use for the present-day Rocky Mountains, nor which lapse rate to use for the Laramide mountains of the Eocene. However, a range in estimates of relief from 1000 to 250 m can be obtained for the Bighorn Mountains of the Eocene by using end-member lapse rates; by using the global average lapse rate, topographical relief between basin floors is estimated to have been 475 m. Despite the obvious uncertainties, all estimates of Bighorn mountain relief obtained by using average $\delta^{18}O_r$ values are much lower than those of Norris et al. (1996, 2000) and Dettman and Lohmann (2000).

Interpreting oxygen isotope data from adjacent basins requires some knowledge of transport directions of air masses in the past and can be complicated by the relative position of Eocene highlands (Fig. 5). For example, $\delta^{18}O_r$ differences between the Green River and Bighorn Basins are more difficult to interpret because winds may have come from directions other than east and south. Furthermore, *both* the Wind River and Owl Creek Mountains may have stood between them, and two orographically induced rainout events may have been superimposed on one another. Finally, it must be remembered that the oxygen isotope compositions may have been shifted to higher $\delta^{18}O_r$ values as a result of evaporation, thereby causing an overestimate of paleotopography. Such a shift is likely in the Green River Basin, and thus sedimentological and fossil evidence that might constrain paleohumidity also needs to be considered.

Laramide Paleoelevation—A Multidisciplinary Approach

Regardless of the exact approach taken, the preceding discussion underscores the fact that estimating topographic relief in the North American Cordillera using oxygen isotope methods is less straightforward than in the cases of the Himalaya (e.g., Garzione et al., 2000; Rowley et al., 2001), Sierra Nevada (e.g., Poage and Chamberlain, 2002), or Southern Alps (e.g., Chamberlain et al., 1999). Overcoming these difficulties will require a combination of (1) more oxygen isotope data from modern precipitation and rivers from the Rocky Mountain region, (2) detailed isotopic studies of fossil material collected from different parts of many basins, and (3) integration of isotope data with global climate model simulations. Modern data are needed to demonstrate clearly that isotopic rain shadows can and do exist between basins and to characterize any seasonal variations in the nature of these rain shadows. In addition, it is important to correlate isotopic offsets between modern basins with predominant directions of air mass transport.

A similar approach is also needed when studying the past. Many intermontane basins are characterized by an excellent spatial distribution of fossils (i.e., north, south, east, and west margins), and a detailed collection of isotope data from these different margins may allow for the isotopic identification of windward and leeward sides of adjacent mountain ranges. When integrated with modern data, it is hoped that paleowind directions and topographic relief between basins can be estimated more accurately.

Last of all, integration of oxygen isotope studies with global climate model simulations is essential in constraining how data are interpreted. Global climate models provide the only practical way of putting isotope data from a single basin into a context of regional and global hydrology and atmospheric circulation patterns. In particular, the incorporation of an oxygen isotope "tracer" into these simulations (e.g., Joussaume et al., 1984; Joussaume and Jouzel, 1993; Gedzelman and Arnold, 1994; Jouzel et al., 1994) should provide a means of comparing isotope data and the models directly. Until this type of multidisciplinary research has been conducted, conclusions regarding the elevation of Laramide mountains during the Paleogene should be made with care.

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