

GEOLOGICAL NOTES

Regional Paleoprecipitation Records from the Late Eocene and Oligocene of North America

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ABSTRACT

The marine $\delta^{18}\text{O}$ record indicates precipitous cooling or significant expansion of Antarctic ice volume across the Eocene-Oligocene epoch boundary at 33.7 Ma. This climatic step is also found in estimates of decreased paleoprecipitation from paleosols in Oregon, Montana, and Nebraska but is only a part of a long-term decline across the Eocene-Oligocene transition rather than a sudden shift. This suggests that climate was driven by a gradual process rather than a single perturbation. Paleoprecipitation levels stabilized during the Oligocene, coincident with a stabilization in $\delta^{18}\text{O}$ values, possibly indicating a new climatic steady state.

Introduction

Cenozoic climate has changed markedly from the warm, wet latest Paleocene thermal maximum (Katz et al. 1999) and Eocene "Cenozoic global climatic optimum" (Clyde et al. 2001) to the present cool, dry glacial-interglacial cycles. This shift has been accompanied by the appearance and then geographic expansion of grasslands in North America and elsewhere to some 25% of the world's land area (Retallack 2001). That floral change has been accompanied by evolution of increased cursoriality and hypsodonty in mammals (Janis et al. 2002). Cenozoic climate is characterized by continuous evolution with a few rapid precipitous changes and short-term aberrations such as the latest Paleocene thermal maximum (Zachos et al. 2001). The largest of these steps is thought to have taken place across the Eocene-Oligocene transition. Zachos et al. (2001) reviewed and compiled a $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ record for the past 65 m.yr. on the basis of benthic foraminifera extracted from over 40 Ocean Drilling Program and Deep Sea Drilling Program drilling sites. Their $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ data spanning the Eocene-

Oligocene transition are shown in figure 1. There is a $>1.0\%$ $\delta^{18}\text{O}$ shift across the epoch boundary that has been attributed in large part (0.6%–1.0%) to increased Antarctic ice volume (Zachos et al. 1994, 2001). Marine Mg/Ca ratio (Lear et al. 2000) and fish otolith data (Ivany et al. 2000) suggest minimal cooling during the Eocene-Oligocene transition and support the notion that most of the $\delta^{18}\text{O}$ shift is due to ice volume changes. There is also a concomitant $\delta^{13}\text{C}$ shift of $\sim 0.8\%$ across the Eocene-Oligocene epoch boundary (Zachos et al. 2001), which represents a significant perturbation of the global carbon cycle. However, the timing of cooling and faunal turnover in the marine and terrestrial records are not synchronous, with terrestrial cooling apparently occurring at ~ 33.2 Ma (Prothero and Heaton 1996; Terry 2001) and most of the faunal turnover even later (Prothero and Heaton 1996; Prothero 1999) rather than at 33.7–33.4 Ma as in the ocean (Zachos et al. 2001). This study presents long-term climatic records from paleosols across the Eocene-Oligocene transition from Oregon, Montana, and Nebraska and discusses possible mechanisms for the observed changes. The new rainfall records are consistent with a gradual forcing mechanism rather than a catastrophic one (e.g., a bolide impact; McGhee 2001).

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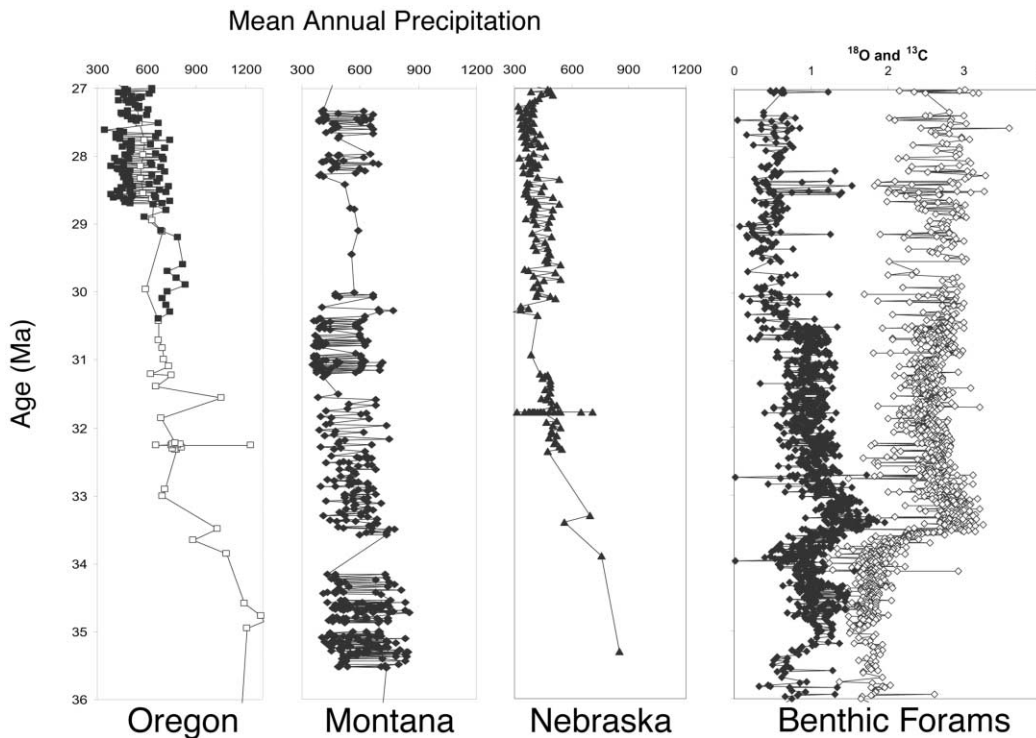


Figure 1. Plot of paleoprecipitation, $\delta^{18}\text{O}$, and $\delta^{13}\text{C}$ versus time. In the left-hand graph, data from Oregon are plotted with boxes, with open symbols for data derived from chemistry and closed symbols for data derived from Bk depths. In the center left graph, Bk depth data from Montana are plotted with diamonds. In the center right graph, Bk depth data from Nebraska are plotted with triangles. In the right-hand graph, carbon isotopes are plotted with closed diamonds and oxygen isotopes are plotted with open diamonds. The standard error on Bk depth paleoprecipitation estimates is 156 mm, and the standard error on CIA-K paleoprecipitation estimates is 182 mm.

Methods

Paleoprecipitation estimates from paleosols are based on two independent proxies: depth to a calcic horizon and the degree of chemical weathering. Both approaches yield results consistent with independent estimates of paleoenvironmental conditions from fossil plants and animals and with each other (Retallack et al. 2000; Sheldon et al. 2002). Soils forming in areas receiving higher mean annual precipitation generally have Bk (carbonate-bearing) horizons lower in their profiles. This relationship quantified for modern soils (Retallack 1994) has been used in investigations of paleosols (Retallack 2001). Bk depths are measured from the top of a paleosol profile using a tape measure and are corrected for any structural dip. The top of a profile is typically markedly finer grained than the overlying sediments, and other evidence of pedogenesis (e.g., root traces) is used to confirm the location of the profile top. For a discussion of the applicability of this approach and its limitations, see articles by Royer (1999, 2000), Retallack (2000),

and Sheldon et al. (2002). Royer (1999) found only a weak relationship between Bk depth and mean annual precipitation, but as Retallack (2000) pointed out, there were a number of conditions that had to be met (e.g., evidence for no erosion of the soil prior to burial) in his original work (Retallack 1994) that were not met by Royer's (1999) data set. Further, as noted above, paleoprecipitation estimates from Bk horizons have been shown to correspond well ($\pm < 200$ mm) to estimates from fossil plants (Retallack et al. 2000) and also to estimates from the degree of chemical weathering (Sheldon et al. 2002; this study) when the criteria of Retallack (1994) are met. The original transfer function (Retallack 1994) is reconfigured here to have depth to the Bk horizon (as a positive number) as the input variable as follows:

$$\text{precipitation (mm)} = -0.013D^2 + 6.39D + 139.6, \quad (1)$$

where D is the depth (cm), $R^2 = 0.63$, and the standard error is 156 mm/yr.

During 2001 and 2002, 1199 Bk depths were measured in various paleosol-bearing formations in Montana, of which a subset ($n = 448$) at ~36–27 Ma from Pipestone Creek, Little Pipestone Creek, McCartney's Mountain, Cook Ranch, Easter Lily, and Mill Point is presented here (fig. 1). An additional 360 Bk depth measurements ($n = 118$ at 36–27 Ma) were made in the John Day Formation of Oregon and are tabulated elsewhere (Retallack et al. 2000; Retallack 2004). In addition to new data, further Bk depth measurements of Eocene and Oligocene paleosols in Nebraska are given by Retallack (1997). A total of 582 Cenozoic Bk depths have been measured in Nebraska ($n = 173$ at 36–27 Ma). In each case, measured Bk depths were adjusted using the decompaction equations of Sheldon and Retallack (2001).

Because the stratigraphic sequences are typically nearly flat-lying and dated units (e.g., tuffs) can be correlated between sections, stratigraphic superposition can be used to infer the age of a given paleosol. Geological ages of the paleosols in Oregon were inferred from paleomagnetic estimates (Prothero and Rensberger 1985) and radiometric ages regressed against stratigraphic level (Retallack et al. 2000; Retallack 2004). Geological ages of paleosols in Montana come from bio- and magnetostratigraphic data (Tabrum et al. 1996) regressed against the composite stratigraphic section of Hanneman and Wideman (1991). Geological ages of paleosols in Nebraska were determined using bio- and magnetostratigraphic data (Prothero and Swisher 1992; Prothero and Whittlesey 1998) regressed against stratigraphic sections. The standard errors of the age models are typically ~200 Ka, ranging from ~160 Ka for the Oligocene of Oregon to ~400 Ka for the Eocene of Nebraska. These age-depth relationships assume a constant sedimentation rate between the age determinations, though not for a given sequence as a whole. Nonetheless, the long-term sedimentation rates for each of the three areas are nearly constant, and the individual age-depth relationships all show very strong correlations ($R^2 = 0.95$ – 0.99).

Our second paleoclimatic proxy is based on the observation that soils receiving higher mean annual precipitation show a greater degree of chemical weathering than soils receiving lower precipitation. This general relationship is due to two factors: (1) higher precipitation leads to greater potential for runoff and for dissolution of soil mineral matter and (2) higher precipitation facilitates greater plant productivity, which augments chemical weathering as plants remove nutrients from the soil. Results are generally consistent with paleo-

precipitation estimates from fossil plants (e.g., Wolfe 1995) where the proxies can be directly compared (e.g., middle Eocene to Late Oligocene; Sheldon et al. 2002). Data on the degree of chemical weathering of Oregon paleosols were compiled elsewhere (Retallack et al. 2000; Sheldon et al. 2002; Retallack 2004). Paleoprecipitation can be estimated using the chemical index of alteration without potassium (CIA-K; Maynard 1992) of paleosol B horizons and the following equation derived by relating the CIA-K values of modern soils to their measured mean annual precipitation (mm/yr):

$$P = 221.12e^{0.0197(\text{CIA-K})}, \quad (2)$$

with $R^2 = 0.72$ and a standard error of 182 mm/yr (Sheldon et al. 2002). For further discussion of the applicability of the proxy, see the article by Sheldon et al. (2002).

A final transfer function relates salinization (S) to mean annual temperature (MAT; °C):

$$\text{MAT} = -18.5S + 17.3, \quad (3)$$

where S is the molar ratio of potash and soda to alumina, $R^2 = 0.37$, and the standard error is 4.4°C for the relationship (Sheldon et al. 2002).

Paleoprecipitation Changes

The paleoprecipitation record from Montana shows considerable variation but little or no secular trend (fig. 1). Because climatic variation is subject to short-term variations as well as long-term trends, it is useful to subdivide the Montana data into time slices. The mean paleoprecipitation at 36–34 Ma is 596 ± 118 (1σ) mm/yr (range 404–845 mm), at 34–32 Ma it is 572 ± 89 mm/yr (range 393–777 mm), at 32–30 Ma it is 498 ± 111 mm/yr (range 388–697 mm), and at 30–27 Ma it is negligibly wetter at 508 ± 94 mm/yr (range 387–697 mm). The younger time intervals are slightly drier but virtually indistinguishable from the older times (i.e., the differences are not statistically significant at the 95% level). Thus, little overall change is indicated across the Eocene-Oligocene transition or into the Oligocene in Montana.

Within each cluster of data, paleoprecipitation estimates vary by ~200–300 mm/yr (fig. 1) between wet and dry intervals, and there appears to be some cyclicity (fig. 2), consistent with Milankovitch-scale variation. In general, each cycle consists of two dry paleosols and one wet paleosol. This apparent cyclicity, coupled with the muted long-term

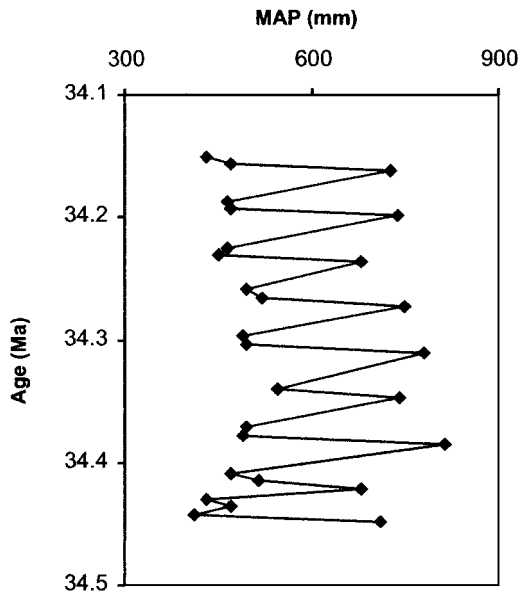


Figure 2. Short-term cyclicality of paleoprecipitation. Each Milankovitch-scale cycle typically consists of two dry and one wetter paleosol.

changes in mean annual precipitation, suggests that Montana's intermontane rain-shadow paleoclimate was affected only to a minor extent by whatever was driving the directed changes in oceanic paleoclimatic proxies.

The data from Nebraska, while sparse, show a possible long-term drying trend from mean annual precipitation of 852 mm at 35.3 Ma to mean annual precipitation of 260 mm at 30.3 Ma with a short-term return to wetter conditions at ~ 31.8 Ma and low variability after ~ 30.5 Ma (fig. 1). Across the Eocene-Oligocene transition, precipitation falls from 755 mm/yr at 33.9 Ma to 562 mm/yr at 33.4 Ma, but this drop is smaller than the return to drier conditions from the later transient wetting event at ~ 31.8 Ma. Climate became very dry with little variability at ~ 31.8 –27 Ma when mean precipitation was 422 ± 64 mm/yr (range 293–707 mm).

The paleoprecipitation record from Oregon reveals both significantly wetter conditions than Montana and a long-term secular trend toward drier conditions similar to Nebraska (fig. 1). Paleoprecipitation changed from >1200 mm/yr at 35 Ma to <600 mm/yr at 28.5 Ma, a change also reflected in the flora and fauna (Retallack et al. 2000). While there appears to have been significant drying at the Eocene-Oligocene transition, paleoprecipitation had been decreasing for at least 1.5 m.yr. before the epoch boundary and continued to fall gradually after the boundary (fig. 1). Furthermore, there are also

transient returns to wetter conditions suggested by single data points at 33.5, 32.3, 31.6 Ma and by multiple data points at 29.9 and 29.6 Ma (fig. 1) superimposed on the ~ 8 Ma secular drying trend. Data from Bk depths give a mean precipitation of 722 ± 65 mm/yr (range 587–832 mm) for the interval at 30.4–28.7 Ma, dropping off to just 527 ± 94 mm/yr (range 385–742 mm) for the interval at 28.7–27 Ma. This difference is statistically significant at the 99% confidence level. A decrease in chemical weathering following the Eocene-Oligocene transition in Oregon is also consistent with climatic cooling (Bestland 2000; Retallack et al. 2000). Though the method employed for estimating paleotemperatures gives statistically significant results that are consistent with floral evidence (Sheldon et al. 2002), it has large uncertainties associated with it (fig. 3). Again, there seems to be a gradual trend toward cooler conditions rather than a sudden change across the Eocene-Oligocene transition.

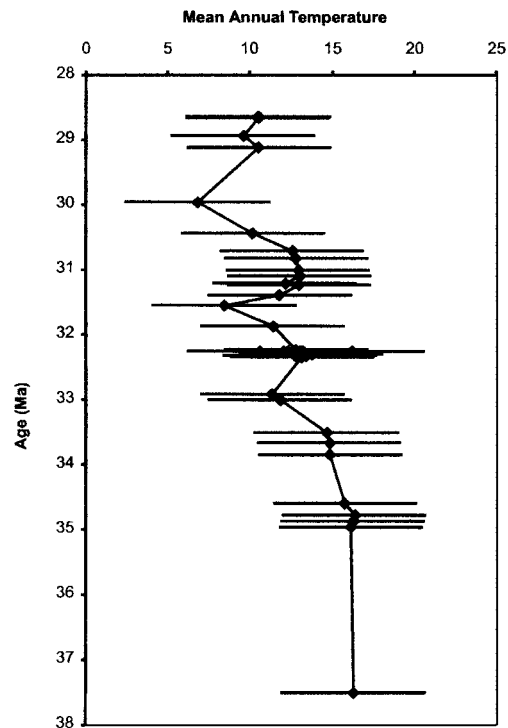


Figure 3. Plot of paleotemperature versus time. Paleotemperatures are calculated using the transfer function of Sheldon et al. (2002) and have a standard error of 4.4°C. Cooling was gradual across the Eocene-Oligocene boundary in central Oregon.

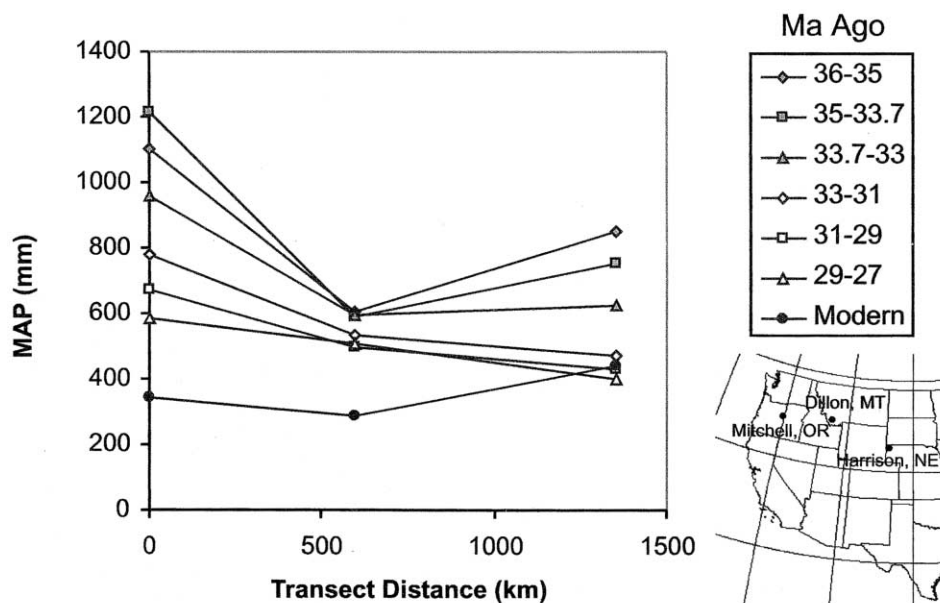


Figure 4. Geographic distribution of precipitation through time. Map modified after the U.S. National Atlas (<http://www.nationalatlas.gov>). Field sites in Oregon are centered around Mitchell, field sites in Montana are centered around Dillon, and field sites in Nebraska are centered around Harrison. There is a marked decrease in east-west precipitation gradients at 36 and 27 Ma. Uncertainties are variable; see table 1.

Discussion

Precipitation is driven in part by proximity to orographic impedances, such as the rain shadow of the Cascade Range that has created the high desert of central Oregon. This may explain the large initial differences in Eocene paleoprecipitation estimates from the dry continental interior setting of Montana compared with Oregon paleosols setting near the Pacific slope during the middle Eocene and in a back-arc basin from the Late Eocene (~39 Ma) on. At 28 Ma, both central Oregon and Montana were receiving similar amounts of precipitation (figs. 1, 4, 5). Paleosols with pedogenic carbonate-like soils of the dry (<760 mm/yr) western United States (Royer 1999) do not appear in the Oregon record until ~30.4 Ma (fig. 1), well after the main growth period of the Antarctic ice sheets at 33.5 Ma. In South Dakota and Nebraska, pedocals appeared at ~35 Ma (Retallack 1983; Terry 2001). In the Montana record, such soils go back to the middle Eocene (~40 Ma; Tabrum et al. 1996). Thus, these dry climatic conditions are not coincident with the opening of the Southern Ocean.

Permanent Antarctic ice sheets are thought to have been present throughout the time interval examined by this study, with the primary period of ice sheet growth occurring across the Eocene-

Oligocene epoch boundary and later Oligocene glacier expansion to sea level (Zachos et al. 2001). Antarctic shelf sediments have an increased proportion of illite to smectite taken to indicate a reduction

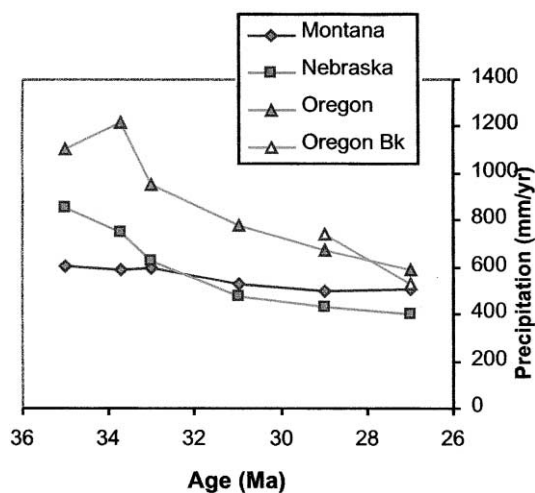


Figure 5. Temporal paleoprecipitation trends. The data have been binned as in table 1 and plotted as the youngest age in the bin. Oregon and Nebraska show a similar pattern of aridification, while Montana exhibits little change.

in chemical weathering across the Eocene-Oligocene epoch boundary due to ice sheet growth (Robert and Kennett 1997). The timing of this change has been thought to be coincident with the opening of the Tasmania-Antarctic passage, which may have pushed the global hydrologic cycle across an ice growth threshold. However, recent studies (Exon et al. 2002) suggest that the Tasmania-Antarctic passage was open at ~37 Ma, so the link between oceanic circulation changes and climatic cooling is not a perfect one. The Drake Passage, in contrast, opened to deep water circulation at ~31 Ma (Lawver and Gahagan 2003), significantly after the Eocene-Oligocene transition, suggesting that it cannot be implicated in the climatic changes at that time either. There was a change in terrestrial paleoclimatic conditions across the Eocene-Oligocene transition but not nearly as marked as the $\delta^{18}\text{O}$ shift at that time in the sea. The $\delta^{18}\text{O}$ shift is probably due to a change in Antarctic ice volume (Zachos et al. 1994, 2001; Lear et al. 2000), which would have had a less profound effect on life and climate on land than the other primary explanation for the observed $\delta^{18}\text{O}$ shift, a dramatic global cooling event. Lack of a large cooling is also borne by the mammalian record, which suggests that the first faunal and floral interchange came after the Eocene-Oligocene epoch boundary by 300–500 Ka (Prothero and Heaton 1996), with the largest changes much later (Prothero 1999).

Between 36 and 27 Ma, east-west precipitation gradients decreased markedly (fig. 4), and differences between wet and dry intervals became muted (note the decline in standard deviation of mean precipitation values, especially after 33 Ma; table 1). That the gradients decreased both eastward and westward suggests the development and expansion of rain shadows on either side of the Rocky Mountains. The simplest tectonic explanation for this pattern is that the western rain shadow formed in response to the uplift/emplacement of the Cascade Range (active since 40 Ma; Verplanck and Duncan 1987), and the eastern rain shadow formed when regional climatic conditions (and circulation pat-

terns) homogenized in response to the post-Laramide subsidence of the Rocky Mountains.

Aridification and the Evolution of Grasslands. Our paleoprecipitation records show long-term aridification of North America across the Eocene-Oligocene transition and into the Early Oligocene (figs. 4, 5; table 1). Throughout the Cenozoic, this change was accompanied by significant expansion of grassy vegetation from a small component of the ecosystem in the Eocene (Clyde et al. 2001) to coverage of 25% of the earth's vegetated surface (Retallack 2001). Modern North American grassland soils (Mollisols) receive 582 ± 181 mm/yr, while modern North American forest soils (Alfisol) receive mean annual precipitation of 991 ± 196 mm/yr (Sheldon et al. 2002). By definition, modern aridland soils (Aridisols) receive less than 500 mm/yr and typically much less (limited data of Sheldon et al. [2002] gives a value of 273 ± 73 mm/yr). The spread of grasslands into areas previously inhabited by arid shrublands (Aridisol belt) and dry woodlands (dry end of the Alfisol belt) suggests that grassland ecosystems adapted more successfully than other vegetation to cool and dry regions, which expanded because of climate change (Retallack 2001). This model is appropriate for Oregon and Nebraska, which show broadly similar trends toward long-term aridification (fig. 5) preceding the incursion of scrubland/grassland ecosystems.

In contrast, paleoclimatic conditions in Montana were steady throughout the interval we studied (figs. 1, 4, 5), and grasslands could have evolved locally in Montana from arid scrubland or expanded their coverage to this and other arid areas. Paleosol evidence (near-mollic Aridisols) suggests shifts at times of paleoclimatic change to bunch grasslands or to woody ecosystems with a higher proportion of grasses (Retallack 2001), but there is little supporting fossil plant evidence at this point, with the exception of ongoing studies of grass phytoliths in Nebraska from the Late Eocene onward (Strömberg 2001) and tooth wear studies of horses (Solounias and Semprebon 2002). Mollic paleosols (fossil grass-

Table 1. Precipitation (mm/yr) through Time by Area

Area	36–35 (Ma)		35–33.7 (Ma)		33.7–33 (Ma)		33–31 (Ma)		31–29 (Ma)		29–27 (Ma)	
		<i>n</i>		<i>n</i>		<i>n</i>		<i>n</i>		<i>n</i>		<i>n</i>
Montana	603 ± 121	91	589 ± 116	107	596 ± 89	42	531 ± 103	97	496 ± 117	54	506 ± 95	57
Nebraska	852 ± 156	1	752 ± 156	1	628 ± 156	2	474 ± 67	51	433 ± 60	34	399 ± 49	84
Oregon (1)	1104 ± 182	1	1217 ± 89	5	957 ± 182	2	778 ± 142	19	671 ± 40	6	587 ± 38	10
Oregon (2)									744 ± 54	11	532 ± 97	107

Note. Uncertainties are 1σ for all time periods with at least five observations and the standard error of the method where there are fewer observations. Oregon (1) data are based on paleosol chemistry, and Oregon (2) data are based on Bk depths.

land soils) have been identified in the Early Oligocene (Terry 2001), and future studies may push their origin back into the Eocene, but the evidence available at present favors the crossing of an aridification threshold in the Oligocene as selective pressure on the evolution of the ecosystem type.

Rate of Climate Change and Oligocene Climate. All of these records point to the predominance of long-term Cenozoic climatic forcing rather than singular events such as impacts. While there were three impacts near the Eocene-Oligocene epoch boundary (McGhee 2001), there are not significant extinctions associated with them. It has been suggested that the Late Eocene greenhouse interval resulted from those impacts, which all struck carbonate rocks that may have vaporized to carbon dioxide (McGhee 2001). The slight Late Eocene decrease in the $\delta^{13}\text{C}$ value could be indicative of such a post-apocalyptic greenhouse (fig. 1). However, the dramatic reversal to icehouse in the isotopic proxies following the boundary is unlikely to have been caused by impacts. The effect of impacts was only a slight and temporary reversal of longer-term trends. Furthermore, changes in precipitation at that time are of smaller magnitude than later wetting events. While short-term perturbations are apparent in our records, it is unlikely that they are the driving force behind long-term directed climatic change and then stability between 36 and 27 Ma in North America.

The most striking feature of Oligocene paleoclimate in all three records is the declining variability between wet and dry intervals (fig. 1) and the continent-scale homogenization of precipitation (fig. 4). Paleoprecipitation stabilized in Nebraska and Montana early in the Oligocene and somewhat later in Oregon. At ~30 Ma, semiarid climate zones covered much of the central and western United States, with paleoprecipitation remaining steady until 27 Ma. Following the Eocene-Oligocene transition, in spite of secular variability, the trend in

$\delta^{18}\text{O}$ values stabilized as well (fig. 1). The rough synchrony of the stabilization of the $\delta^{18}\text{O}$ values and our regional paleoprecipitation records suggests that the global hydrologic cycle may have stabilized at this time as well, probably in response to the presence of permanent ice sheets (Zachos et al. 2001) or the stabilization of atmospheric CO_2 levels, which were declining throughout the Eocene-Oligocene transition before leveling off in the Oligocene (e.g., Retallack 2002; unpublished modeling and paleosol results by N. D. Sheldon).

Conclusions

Long-term continental paleoprecipitation records from paleosols in Oregon, Nebraska, and Montana indicate gradual aridification across the Eocene-Oligocene transition, with much of the drying occurring well after the epoch boundary. The distribution and regional gradients of precipitation are consistent with tectonically controlled changes in orographic impedances and their resulting rain shadows. This gradual climatic shift was likely accompanied by the expansion of grassland ecosystems into ecological niches previously occupied by dry woodland vegetation. Gradual climatic drivers such as large southern hemisphere ice sheets and mountain uplift or declining atmospheric CO_2 levels are better able to explain the gradual nature of the Eocene-Oligocene transition and the low amplitude of Oligocene climatic change than more abrupt geological perturbations such as bolide impacts.

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