Late Miocene climate and life on land in Oregon within a context of Neogene global change

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Abstract

Clarendonian (12 Ma) fossil soils, plants, molluscs, fish, and mammals of eastern Oregon allow reconstruction of late Miocene paleogeography, paleoclimate, and paleoecology on land between the global thermal maximum of the middle Miocene (16 Ma) and global cooling and drying of the late Miocene (7 Ma). Six different pedotypes of paleosols recognized near Unity and Juntura allow reinterpretation of local mammalian paleoecology. Fossil beavers dominated gleyed Entisols of riparian forest. Abundant camels and common hipparionine horses dominated Alfisols of wooded grassland and grassy woodland. Diatomites overlying mammal-bearing beds have bullhead catfish \( Ictalurus (Ameiurus) vespertinus \), as well as fossil leaves dominated by live oak \( Quercus pollardiana \). Fossil plants and soils of Unity and Juntura are most like those of grassy live oak woodland and savanna on the western slopes of the Sierra Nevada in northern California today. Fossil plants and soils indicate mean annual temperature of 12.9 (7.7–17.7) °C and mean annual precipitation of 879 (604–1098) mm. Miocene paleoclimatic changes in eastern Oregon show no relationship to changes in oxygen isotopic composition of marine foraminifera, usually taken as an index of global paleoclimatic change. Mismatch between land and sea paleoclimatic records is most likely an artefact of global ice volume perturbation of oxygen isotopic values. Instead, Miocene paleoclimatic change in eastern Oregon parallels changes in carbon isotopic composition of marine foraminifera, presumably through fluctuations in greenhouse gases. © 2004 Elsevier B.V. All rights reserved.

Keywords: Miocene; Paleosols; Plants; Mammals; Oregon

1. Introduction

Because Miocene soils, plants, and animals were close to our own time and modern in many respects, the surprisingly large amplitude of Miocene paleoclimatic change is of interest for understanding future global change. Evidence of global middle Miocene (16 Ma) warmth includes lateritic paleosols as far north and south as South Australia, Japan, Oregon, and Germany (Schwarz, 1997). Middle Miocene thermophilic trees such as sweetgum \( Liquidambar pachyphylla \) grew as far north as Alaska (Wolfe and Tanai, 1980), and coconuts \( Cocos zeylanica \) dispersed as far south as New Zealand (Fleming, 1975). Middle Miocene thermo-
philic large foraminifera, corals, and molluscs extended as far north as Oregon, Alaska, Japan, and Kamchatka (Moore, 1963; Itoigawa and Yamanoi, 1990; Oleinik and Marinovich, 2001), and as far south as South Australia and New Zealand (Li and McGowran, 2000; Fleming, 1975). These indicators of warmth disappeared from such high latitudes by later middle Miocene (15 Ma), when cold water minerals (glendonite pseudomorphs of ikaite) formed in marine rocks of Oregon (Boggs, 1972). Benthic foraminiferal oxygen isotopic values also indicate a global middle Miocene marine thermal maximum, followed by late Miocene global cooling (Zachos et al., 2001).

These global trends also were felt in eastern Oregon, where fossil leaves indicate marked cooling and drying between about 16 and 15 Ma (Wolfe et al., 1997; Graham, 1999), and a second cooling and drying to a modern flora by latest Miocene (7 Ma; Chaney, 1948). Paleosols of eastern Oregon show similar changes between middle Miocene mesic forests and late Miocene sagebrush steppe (Bestland and Krull, 1997; Retallack et al., 2002a; Sheldon, 2003). Associated mammal faunas include a transition from three-toed, mixed-feeding horses of the middle Miocene (15 Ma, Barstovian) to modern monodactyl grazing horses by late Miocene (7 Ma, Hemphillian; Downs, 1956; Shotwell, 1970).

Fig. 1. Late Miocene fossil localities of Juntura, Ironside, and Unity, eastern Oregon, with detailed geological map and cross-section around Unity (Thayer, 1957; Thayer and Brown, 1973; Reef, 1983; Walker and MacLeod, 1991). Numbered fossil localities and fossils are listed in Tables 4–5; unnumbered localities have not yet yielded identifiable specimens.
This study examines paleoclimate and paleoecology of the intervening part of the late Miocene within the context of Miocene global paleoclimatic extremes, using evidence from fossil soils, plants, molluscs, fish, and mammals in the eastern Oregon. Previous study of mammalian communities of this age from near Juntura (Shotwell, 1963; Shotwell and Russell, 1963), Ironside (Merriam, 1916), and Baker City (Downs, 1952) are here supplemented with additional collections from near Unity, including reappraisal of a locality for a previously collected gomphothere skull and jaws (Orr and Orr, 1999). New collections of fossil plants and fish are also reported, as well as new mapping of the area around Unity (Fig. 1) and an assessment of the paleoenvironmental significance of paleosols associated with these fossils (Figs. 2 and 3). The record of Neogene paleoclimatic fluctuation from these and other paleosols in eastern Oregon and Washington (Retallack, 1997b) does not correlate with oxygen isotopic records from deep-sea foraminifera but is in phase with foraminiferal variations in carbon isotopic composition (Zachos et al., 2001). This may indicate a closer relationship with the carbon cycle and greenhouse gases,

Fig. 2. Measured section of the Ironside Formation in Windlass Gulch, 3 miles northeast of Unity, Baker County, Oregon, showing paleosols and their development, calcareousness, and hue (using conventions of Retallack, 1997a, 2001b).
such as CO₂ and CH₄ (Retallack, 2002), than with deep-sea temperature or polar ice volume tracked by oxygen isotopic composition.

2. Materials and methods

Mapping and measurement of two stratigraphic sections near Unity, Baker County, Oregon (Fig. 3) at fossil quarries of Shotwell (1963). Petrographic thin sections were cut of all the collected rocks in order to support field identifications of lithologies, and 36 thin sections were point-counted to determine paleosol textures and mineral composition (following the methods of Retallack, 1997a, 2001b). Clays and evaporite minerals were identified by X-ray diffraction using a Rigaku computer-automated goniometer.

Six distinct kinds of paleosols, or pedotypes, were recognized (Table 1) throughout the mapped area (Figs. 1 and 2) and also near Juntura (Fig. 3) and Ironside. Each pedotype represents a distinct ancient environment (Table 2). The pedotype names are from the Sahaptin Native American language (Rigsby, 1965; Delancey et al., 1988) and are part of a wider classification of Oregon Cenozoic paleosols (Retallack et al., 2000). Descriptive and analytical data for these paleosols are presented in Fig. 5 and Table 1. Chemical analyses of 21 paleosol specimens by Bondar Clegg of Vancouver, Canada, used X-ray fluorescence of a borate-fused bead for major elements, titration for ferrous iron, and gravimetry for loss on ignition. Bulk density for each specimen was determined using the clod method. Of 109 paleosols logged near Unity and 42 near Juntura, 45 had pedogenic carbonate, and 5 were chemically analyzed for paleoclimatic interpretation. These data are added to 754 pedogenic carbonate measurements and 92 chemical analyses of other paleosols from Oregon and Washington ranging back in age to 28.7 Ma (Gustafson, 1978; Bestland and Krull, 1997; Tate, 1998; Retallack, 2004), in order to assess local Neogene paleoclimate and its correspondence with global paleoclimatic change (Zachos et al., 2001).

3. Geological background

Only rocks of the Clarendonian North American Land Mammal “Age” around Unity are described and mapped in detail here (Figs. 1 and 2). Few fossils have been recovered from poor exposures of the Ironside Formation near Ironside (Merriam, 1916) and Baker City, Oregon (Downs, 1952). The Unity, Ironside, and Baker City sequences and local faunas are similar to
better-known sequences and faunas of the Juntura Formation near Juntura, Malheur County (Fig. 3; Bowen et al., 1963; Greene, 1973; Fiebelkorn et al., 1982; Johnson et al., 1996).

3.1. John Day Formation (late Eocene–early Miocene)

The oldest rocks exposed near Unity are 300 m of rhyodacitic tuff breccias, interpreted as lahars by Reef (1983). He considered them Eocene in age, based on fossil leaves (*Rhamnidium chaneyi*) and permineralized palm wood (*Palmoxylon* sp. indet.). However, Walker (1990) reports K–Ar ages from hornblende and plagioclase of 19.6 ± 0.8 and 19.5 ± 0.6 Ma, respectively (early Miocene), and regarded these rocks as outliers of the late Eocene to early Miocene John Day Formation. The laharic tuff breccia is overlain by a 100-m-thick rhyodacitic flow of similar petrographic and geochemical composition to the tuff breccia, additional tuff breccia, and local basalt flows (Reef, 1983; Walker, 1990). These volcanic rocks are overlain by about 20 m of red claystone, including numerous paleosols of the Luca pedotype, which are late middle Eocene to early Miocene.

**Table 1**

<table>
<thead>
<tr>
<th>Pedotype</th>
<th>Sahaptin meaning</th>
<th>Diagnosis</th>
<th>Australian classification (Stace et al., 1968)</th>
<th>F.A.O. World Map (F.A.O., 1974)</th>
<th>U.S. Soil Taxonomy (Soil Survey Staff, 1999)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Abiaxi</td>
<td>Bitterroot</td>
<td>Near-mollie surface (A) horizon over weakly weathered subsurface (Bw)</td>
<td>Brown clay</td>
<td>Eutric Cambisol</td>
<td>Haplosalicl</td>
</tr>
<tr>
<td>Cmti</td>
<td>New Underneath</td>
<td>Shale and siltstone with root traces</td>
<td>Alluvial soil</td>
<td>Eutric Fluvisol</td>
<td>Fluvental</td>
</tr>
<tr>
<td>Monana</td>
<td></td>
<td>Thin impure lignite (O) on siltstone (A) with shallow root traces</td>
<td>Humic gley</td>
<td>Eutric Histosol</td>
<td>Fibrst</td>
</tr>
<tr>
<td>Skaw</td>
<td>Scare</td>
<td>Thin dark gray to black spotted siltstone with root traces (A) over bedded siltstone</td>
<td>Wiesenboden</td>
<td>Eutric Gleysol</td>
<td>Molllic Endoaquent</td>
</tr>
<tr>
<td>Tutanik</td>
<td>Hair</td>
<td>Near-mollie brown thick siltstone (A) with deep calcareous and siliceous rhizoconcretions and nodules (Bk)</td>
<td>Brown earth</td>
<td>Chromic Luvisol</td>
<td>Typic Natrixeralf</td>
</tr>
<tr>
<td>Xaus</td>
<td>Root</td>
<td>Sandstone with root traces (A) often ferruginized over bedded sandstone</td>
<td>Alluvial soil</td>
<td>Eutric Fluvisol</td>
<td>Psamment</td>
</tr>
</tbody>
</table>

Fig. 4. Badlands of the late Miocene Ironside Formation in central Windlass Gulch (95–135 m of measured section Fig. 2), northeast of Unity. A gray airfall tuff is visible to the lower right and two thin white tuffs in the middle of these badlands.
Oligocene (Uintan to Orellan) in age near Clarno and the Painted Hills, Oregon (Bestland et al., 1999; Retallack et al., 2000). Comparable red paleosols also are found in the upper John Day Formation east of Kimberly, Oregon, where they formed on well-drained uplands (Retallack et al., 2000).

3.2. Strawberry and Dooley Volcanics (middle Miocene)

Platy, 2-pyroxene, basaltic andesites are characteristic of middle Miocene (radiometric ages 15–10 Ma), Strawberry Volcanics, with smaller amounts of basalt, rhyolite, rhyolitic tuff, vent breccia, and micronorite (Robyn, 1977). The Strawberry Volcanics are calcalkaline and were not comagmatic with voluminous tholeiitic basalts of the Columbia River Basalt Group, which also are middle Miocene (radiometric age 17–15 Ma; Sheldon, 2003). Calcalkaline volcanism is produced by melting above subduction zones, but Strawberry Volcanics and other coeval andesitic volcanoes (Walker, 1990) were far inland from subduction of the Farallon Plate during the middle Miocene (Orr et al., 1992). Middle Miocene Strawberry stratovolcanoes rose at least 1000 m above valley floors of the Paleozoic and Mesozoic basement (Robyn, 1977).

Dooley Volcanics near Unity are 150 m thick and include three distinct units of rhyolitic, welded, ash flow tuffs (Reef, 1983). The ash flow tuffs are light gray with abundant pumice, rock fragments, and phenocrysts of plagioclase, biotite, and quartz. Dooley Volcanics reach a thickness of 2400 m around volcanic centers only a few kilometers to the north and northwest of the mapped area. One of the last eruptive events of the Dooley Volcanics was a rhyolite dome dated by K–Ar at 14.7 ± 0.4 Ma (Evans, 1992).

3.3. Ironside Formation (late Miocene)

The Ironside Formation is at least 250 m of conglomerates, siltstones, diatomites and lignites
Fig. 5. Petrographic and chemical data for newly described paleosols of Tutanik and Xaus paleosols from 4 to 6 m, Tutanik, Cmti, Xaus, Skaw, and Abiaxi paleosols from 41 to 46 m, and Abiaxi, Cmti and Skaw paleosols from 229 to 232 m in reference section of Ironside Formation in Windlass Gulch (Fig. 2).
named for the small village of Ironside, 15 km east on U.S. highway 26 from Unity (Thayer, 1957; Thayer and Brown, 1973). Sediments and paleosols along the highway east of Ironside are identical with those near Unity and particularly with the lower part of a measured section in Windlass Gulch (0–7 m in Fig. 2). Diatomites like those in the upper part of the Ironside Formation near Unity, also crop out in the hills north of Ironside. A fossil locality southeast of Baker City also exposes tuffs and conglomerates like those of the lower Ironside Formation (95–100 m in Fig. 2) in Windlass Gulch (Downs, 1952; Evans, 1992).

These three areas are linked by late Miocene fossil mammals of the Clarendonian North American Land Mammal "Age" (8.4–12.3 Ma of Prothero, 1998), although nomenclatural difficulties remain. For example, a fossil proboscidean skull from Unity on display at the Oregon Museum of Science and Industry in Portland (Fig. 6C–D) has remained undescribed beyond a photograph and account of its collection (Baldwin, 1964; Orr and Orr, 1999). The excavation is still visible within an Abiaxi paleosol on a low knoll southeast of a prominent bluff, east of Unity (locality 8 of Table 4; Walker, 1990, figs. 5 and 9). George Gaylord Simpson examined the specimen in 1953 and suggested it was “Miomastodon merriami” (Orr and Orr, 1999). A more appropriate name is “Gomphotherium cingulatum” of Downs (1952), based on a lower jaw discovered by Albert Werner at a locality 15 km east of Baker City and about 80 km northeast of Unity. This and
other validly established taxa were not recognized by Lambert and Shoshani (1998), who considered the genus oversplit. Tobien (1972) went so far as to include all North American *Gomphotherium* in “*Gomphotherium productum*”. Here, I follow Madden and Storer (1985) in using their broadly defined *Gomphotherium osborni* (Barbour) for the Unity skull. Another gomphothere tooth, probably conspecific, was found by Elmer Molthan in the Ironside Formation 400 m south of Ironside Post Office, which is only 15 km east of Unity (“*Tetrabelodon*” sp. of Merriam, 1916). *Gomphotherium* and two additional proboscidean species are recognized from the Clarendonian Juntura Formation to the south in Malheur County (Shotwell and Russell, 1963): a two-tusker “*Mammuth furlongi*” (a species not recognized by Lambert and Shoshani, 1998), and a shovel-tusker similar to *Platybelodon barnumbrowni* (but regarded as a new genus by Lambert and Shoshani, 1998).

There are also nomenclatural problems with Clarendonian horse teeth from Unity and Juntura, because among the numerous teeth and jaws, there is no skull with facial fossae or other diagnostic...
features. The names given to these teeth ("Hipparion anthonyi" of Merriam, 1916 and "Hipparion condoni" of Shotwell and Russell, 1963) are considered by MacFadden (1984) likely junior synonyms of Cormohipparion occidentale or Cormohipparion sphenodus, the latter most likely on the basis of size. Shotwell and Russell (1963) considered that the large collection of teeth and jaws from Juntura represented only one species.

Late Miocene geological age is also indicated by fossil plants from near Unity and Juntura. A new fossil plant assemblage in Juniper Gulch near Unity (Fig. 7; locality 6 of Table 4, at 160 m in Fig. 2) is here designated the Unity flora. Entire to weakly dentate oak leaves like those dominating this assemblage have been identified as "Quercus hannibali", but this name has been included within Quercus parishiana (Knowlton) by Axelrod (1995) or identified with the similar, but more dentate, living Quercus chrysolepis (Wolfe for Leopold and Wright, 1985). The Unity flora is taxonomically most like the late Miocene Stinkingwater flora in the Juntura Formation near Juntura (Chaney and Axelrod, 1959) and the Weiser flora from the Poison Creek Formation of Idaho (Dorf, 1936; Leopold and Wright, 1985).

The thick gray volcanic ash of the Ironside Formation at 171 m in Windlass Gulch is mineralogically most similar to the 11.31 Ma CPTXI Ash of the Great Basin (Perkins et al., 1998), and another gray tuff at 101 m is most like the 11.59 Ma CPTIX Ash (Figs. 2 and 3). A geological age of 11–12 Ma is similar to the age of the Juntura Formation, 78 km to the south in Malheur County. A whole-rock K–Ar date for 12-m-thick basalt 152 m stratigraphically below a fossil mammal quarry (OU locality 2337 of Fig. 3) and 85 m above the Stinkingwater flora (Chaney and Axelrod, 1959) near Juntura is 12.4 Ma (Evernden and James, 1964; corrected using Dalrymple, 1979). A tuff 76 m stratigraphically above the mammal quarry has been dated by K–Ar on shards at 11.5 ± 0.6 Ma (Fiebelkorn et al., 1982), and the Devine Canyon ashflow tuff at the base of the Drewsey Formation, 107 m above the quarry, has been dated by $^{39}$Ar/$^{40}$Ar of sanidine as 9.7 Ma (Johnson et al., 1996; five similar K–Ar dates are tabulated by Greene, 1973). Also comparable in age and lithology are volcanic ashes, conglomerates, and diatomites of the Coal Valley Formation of Nevada dated by the K–Ar technique at 11.4–9.4 Ma (Golia and Stewart, 1984).

4. Late Miocene paleogeography of eastern Oregon

4.1. Sedimentological evidence

The depositional basin of the Ironside Formation was bounded by volcanic terranes in the Strawberry and Dooley Volcanics to the north and west, and fault blocks of Paleozoic and Mesozoic basement to the north and southwest. There was erosional paleorelief on the underlying Dooley Volcanics and Clarno Formation, as indicated by several inliers cropping out south of Juniper Gulch. The stepped system of northeast-extensional faults (Fig. 1) did not exist during deposition, when this broad alluvial and lacustrine basin extended southeast from highlands to the north and west (Fig. 8).

Basin-marginal boulder breccias and conglomerates interfinger with diatomites and tuffaceous siltstones of the Ironside Formation in road cuts over Dooley Mountain and north of Whited Reservoir (Fig. 1). Paleocurrent directions measured from trough cross bedding in conglomeratic paleochannels in the Windlass Gulch section (Fig. 9) indicate that streams flowed southeast. This early phase of river channels and floodplains was followed by extensive oligotrophic lakes, which eventually accumulated diatomites out to the hilly margins of the depositional basin. A final depositional phase of diminished lake area is recorded by fluvial gravels and lignites at the top of the exposed sequence near Unity (Fig. 2).

4.2. Paleosol evidence

Various paleosols represent different sedimentary environments and communities within these river valleys and lake shores. Weakly developed paleosols (Abiaxi, Cmti, Xaus) represent ephemeral communities on the landscape, such as vegetation colonizing stream and lake margins after flooding. Weakly developed carbonaceous paleosols (Skaw) represent riparian or lake-margin marsh, and moderately
developed peaty paleosols (Monana) represent swamp woodlands, where decay of organic debris was suppressed by seasonally high water table.

In contrast, Tutanik paleosols represent stable floodplains and alluvial terraces. Tutanik paleosols are thick, brown, and have little relict bedding due to bioturbation (including siliceous rhizoconcretions; Fig. 10A–B) and to development of soil structure (fine subangular blocky peds) and microfabric (skel-masepic porphyroskelic plasmic fabric of Brewer, 1976). They are also more aluminous and less rich in
alkalis and alkaline earths than associated paleosols (Fig. 5). Although volcanic ash was an important part of Tutanik paleosol parent material, few recognizable shards have escaped weathering. This degree of weathering is found after 8–27 ka in tropical highland soils of New Guinea (Ruxton, 1968). Tutanik paleosols have common clay skins in thin section and a higher proportion of fine pedogenic clay than associated paleosols (Fig. 5). The degree of enrichment of clay is intermediate between that seen in upper Modesto soils dated at 10 Ka and lower Modesto soils dated at 40 Ka in alluvium of the Merced River, San Joaquin Valley, California (Harden, 1982, 1990).

4.3. Paleontological evidence

Diatomites near Unity have not been studied in detail, but of 108 diatom taxa described from coeval diatomites of Juntura, only two are found in estuaries and the sea: “Coscinodiscus miocenicus” and “Coscinodiscus subtilis” of Van Landoningham (1967; perhaps better referred to Actinocyclus according to Bradbury et al., 1985). Melosira granulata, Melosira distans, and Fragilaria construens dominate individual collections (58–88%) and indicate fresh water of low alkalinity, low in dissolved solids, neutral to slightly acidic in pH, and low in nutrient levels (Bradbury et al., 1985). Significant (8%) amounts of C. subtilis in one sample (Van Landoningham, 1967) may indicate connection to the sea by way of low-gradient streams (Bradbury et al., 1985). Fossil aquatic molluscs from Juntura and Unity (Tables 4–5; Taylor, 1985) are evidence for a Miocene course of the Snake River southwest into the Pit River of northern California, before Basin and Range faulting and northward capture by the Columbia River (Van Tassell et al., 2001). Lakes were also needed by many of the fossil birds (Brodkorb, 1961): cormorants, coots, mergansers, teals, and extinct straight-billed flamingos (Table 5).

5. Late Miocene paleoclimate of eastern Oregon

5.1. Evidence from paleosols

Only Tutanik paleosols are useful paleoclimatic proxies, because other pedotypes were too weakly developed to show a paleoclimatic signature. Differences between Tutanik and other comparably developed Cenozoic paleosols from Oregon may reflect changing Neogene paleoclimate. Tutanik paleosols are generally similar in their brown color and clayey subsurface (argillic horizons) to Skwiskwi paleosols of the Oligocene John Day and latest Miocene
Rattlesnake Formations (Retallack et al., 2000, 2002a), and to early Miocene Tima paleosols of the John Day Formation (Retallack, 2004). Tutanik paleosols lack thick silcretes of Tima paleosols and have silica–micrite rhizoconcretions (Fig. 8A–B) and chaledony septarian nodules (Fig. 8C–I) unknown in Skwiskwi paleosols. The large amount of silt (never less than 27% by volume) and sodium (soda/potash ratios >1) also ally Tutanik paleosols with Tima rather than Skwiskwi paleosols. Tutanik paleosols also differ from paleosols of the late Miocene (7 Ma) Rattlesnake Formation, such as the Tatas pedotype, which was crumb-structured, moderately leached, and calcareous (Retallack et al., 2002a). Tutanik paleosols thus show greater affinities with interpreted summer-dry (xeric) paleosols, such as Tima, than summer-wet (ustic) paleosols, such as Skwiskwi and Tatas.

Modern soils like Tutanik paleosols are found near Redding, northern California (map unit Le3-2a with duric phase of F.A.O., 1975), where mean annual temperature is 17.7 °C, and mean annual precipitation is 1040 mm, January and July temperatures are 8 and 29 °C, respectively, and precipitation is 216 mm in January, but only 5 mm in July (Ruffner, 1985). In contrast, for Baker City near Unity, mean annual temperature is 7.6 °C, mean annual rainfall is 270 mm, temperature is more seasonal (−4 °C in January, 19 °C in July), but precipitation is less seasonal (35 mm in January, 12 mm in July; Ruffner, 1985).

Independent estimates of former precipitation can be gained from chemical composition of clayey horizons and from carbonate in rhizoconcretions of Tutanik paleosols at depths of 79–91 cm. This depth (D in cm) can be used to estimate mean annual rainfall (P in mm ± 156) using a transfer function derived from study of modern soils (Retallack, 1994), once allowance is made for burial compaction (Sheldon and Retallack, 2001) due to at least 60 m of overlying Pleistocene fanglomerate (Evans, 1992). A large database of chemical analyses of North American soil B horizons has been used to derive yet other transfer functions for mean annual precipitation (P in mm ± 182) and mean annual temperature (T in °C ± 4.4) from two chemical measures: S, the molar ratio Na₂O+K₂O/Al₂O₃; and C, molar Al₂O₃/(Al₂O₃+CaO+Na₂O) times 100 (following Sheldon et al., 2002), as follows:

\[ P = 139.6 + 6.388D - 0.01303D^2 \]

\[ P = 221e^{0.02C} \]

\[ T = -18.516(S) + 17.298. \]

Paleoclimatic results of these calculations (Table 3) are 450–800 mm mean annual precipitation from depth to carbonate and 600–1100 mm from chemical composition of clayey horizons. Because the carbonate-depth transfer-function was designed for use with nodules (Retallack, 1994, 2000), rather than rhizoconcretions, these estimates are less likely to be reliable than the subhumid regime estimated from chemical composition of B horizons. Geochemical composition of paleosol parent rhyodacitic and andesitic airfall ash shows insignificant change through time compared with paleosol modification (Perkins et al., 1998; Retallack et al., 2000).

A highly seasonal climate is indicated by concentrically banded rhizoconcretions with silica and micrite in Tutanik paleosols (Fig. 10A–B). Seasonality is also indicated by the unusually abundant silt and soda in Tutanik paleosols, especially considering their likely mean annual precipitation (Table 3). In such rainy climates, there would not be so much salt or dust unless summers were very dry, and most rain and snow were during winter and spring. Furthermore, the scarce carbonate and distinctive silica–micritic rhizoconcretions (Fig. 10A–B) are like those found in modern soils of Mediterranean (summer-dry) climates (Chadwick et al., 1987, 1995). Siliceous rhizoconcretions from a few Tutanik paleosols (45, 59, and 69 m in Fig. 2) grade insensibly through cigar- and tear-shaped forms into ellipsoidal septarian nodules of a form not previously reported from paleosols (Fig. 10C–F). Some of these contain fossil wood or lumpy root traces morphologically comparable to rhizobially nodulated roots. Others are hollow and cracked, with void-filling dolomite rhombs (Fig. 10G, I). Microfabric of the nodules is radially oriented, dendritic chaledony, creating an external fabric reminiscent of cone-in-cone structure. The ellipsoidal nodules are superficially similar to chaledony nodules of the “button beds”, a lacustrine facies of the middle
My own examination of the Barstow nodules found conspicuous relict bedding and microcrystalline textures, unlike the septarian and radial-dendritic crystal form of the Unity nodules. Nevertheless, highly alkaline late-summer aridity could have played a role in the formation of both kinds of siliceous nodules.

5.2. Evidence from fossil plants

Close floristic composition of Unity (Table 5) and Stinkingwater floras (Chaney and Axelrod, 1959) with the modern flora of California supports evidence from associated paleosols for extension of Mediterranean climate into eastern Oregon during the late Miocene. The paleoclimatic affinities of these floras are quite distinct from more humid and warm climates evident during the middle Miocene thermal maximum, represented locally by the Austin, Tipton, Baker, and Mascall floras (Oliver, 1934; Brown for Gilluly, 1937; Chaney and Axelrod, 1959), and before the advent of modern sagebrush and riparian communities during the Pliocene (Chaney, 1948; Ashwill, 1983, Retallack et al., 2002a). Mean annual precipitation of around 900–1000 mm and mean annual temperature of 13 (8–18) °C have been estimated from other late Miocene fossil floras of the Great Basin (Wolfe, 1994; Wolfe et al., 1997; Graham, 1999). A summer-dry paleoclimatic regime for the Unity flora is also compatible with frequent wildfires indicated by charcoal associated with middle to late Miocene fossil floras of southeastern Oregon (Taggart et al., 1982; Taggart and Cross, 1990).

5.3. Evidence from fossil animals

Abundance of bullhead catfish in siltstones with the Unity flora (Fig. 7; Table 4) can be contrasted with the overlying diatomites, which have a sparse fish fauna of other kinds of fish. The leaves and catfish probably represent a nearshore lacustrine habitat distinct from that of the open lake. Bullhead catfish lived in western North America from the Eocene to Pliocene, but after that time became restricted to eastern North America. They have been artificially reintroduced to Oregon and Idaho within the last 120 years. Today, they live in low-gradient streams and lakes at elevations less than 1000 m in warm temperate climates with at least 230 frost-free days and at least 400 mm mean annual precipitation (Van Tassell et al., 2001).

Brodkorb (1961) invokes Bergmann’s rule (Brown and Lomolino, 1998) to argue for a warmer than present climate from the small size of Juntura cormorants, coots, and teals compared with living relatives. Oxygen isotopic studies of fossil mammal teeth from Juntura, Oregon, indicate mean annual temperatures of about 18 °C, and a mean annual range
of temperature of some 25 °C (Kohn et al., 2002), compared with 7.6 and 23 °C, respectively, for Baker City today (Rufiner, 1985).

Indications of a late Miocene climate at Juntura, wetter than at present, come from the sewerelf (Tardontia in Table 5). Similar aplodontids are now restricted to Oregon and Washington, west of the Cascades (Shotwell, 1958). The western American mole (Scapanus) is also now found well west of Juntura and Unity (Hutchison, 1968). On the other hand, subhumid to semiarid rangeland conditions are indicated by ground squirrels, pocket gophers, and extinct mylagaulines (Shotwell, 1958, 1970).

6. Late Miocene palaeoecology of eastern Oregon

6.1. Evidence from paleosols

A detailed local mosaic of vegetation (Fig. 8) can be inferred from the relationship of paleosols to local sedimentary facies and from plant-induced features of the paleosols (Figs. 2 and 3; Table 3). Tutank paleosols are finely structured (granular peds about 1 cm across) with abundant fine root traces as well as scattered rhizoconcretions, as in soils supporting grass with scattered woody shrubs or trees (Retallack, 1997b, 2001b). The paleosols lack fine crumb peds found in sod-forming grass-
Table 5

Late Miocene (Clarendonian) megafossils of the Juntura Formation near Juntura

<table>
<thead>
<tr>
<th>Fossil taxon</th>
<th>Common name</th>
</tr>
</thead>
<tbody>
<tr>
<td><em>Sphaerium</em> sp. cf. <em>S. lavernense</em></td>
<td>Aquatic lake orb</td>
</tr>
<tr>
<td>Herrington (Herrington and Taylor, 1958)</td>
<td>mussel</td>
</tr>
<tr>
<td><em>Pisidium</em> sp. cf. <em>P. clessini</em></td>
<td>Aquatic pea clam</td>
</tr>
<tr>
<td>Neumayr and Paul (1875)</td>
<td></td>
</tr>
<tr>
<td><em>Pisidium</em> sp. indet.</td>
<td>Aquatic pea clam</td>
</tr>
<tr>
<td><em>Fluminicola junturae</em> Taylor, 1963</td>
<td>Aquatic pea clam</td>
</tr>
<tr>
<td><em>Viviparus turneri</em> Hannibal, 1912</td>
<td>Aquatic pea clam</td>
</tr>
<tr>
<td>Radix junturae Taylor, 1963</td>
<td>Aquatic pea clam</td>
</tr>
<tr>
<td><em>Carinifex shotwelli</em> Taylor, 1963</td>
<td>Aquatic pea clam</td>
</tr>
<tr>
<td><em>Promenetus</em> sp. indet.</td>
<td>Aquatic pea clam</td>
</tr>
<tr>
<td><em>Italurus peregrinus</em> Lundberg, 1975</td>
<td>Aquatic pea clam</td>
</tr>
<tr>
<td><em>Cyprinidae</em> gen. indet.</td>
<td>Extinct stork</td>
</tr>
<tr>
<td><em>Catastomidae</em> gen. indet.</td>
<td>Aquatic pea clam</td>
</tr>
<tr>
<td><em>Phalacrocoryx leptopus</em> Brodkorb, 1961</td>
<td>Aquatic pea clam</td>
</tr>
<tr>
<td><em>Ciconiidae</em> gen. indet.</td>
<td>Extinct stork</td>
</tr>
<tr>
<td><em>Megapalaeolodus opsigonus</em> Brodkorb, 1961</td>
<td>Extinct stork</td>
</tr>
<tr>
<td><em>Eutamias</em> sp. indet.</td>
<td>Aquatic pea clam</td>
</tr>
<tr>
<td><em>Hystricops</em> sp. indet.</td>
<td>Aquatic pea clam</td>
</tr>
<tr>
<td><em>Eucastor malheurensis</em> Shotwell and Russell, 1963</td>
<td>Extinct stork</td>
</tr>
<tr>
<td><em>Eremochen russelli</em> Brodkorb, 1961</td>
<td>Aquatic pea clam</td>
</tr>
<tr>
<td><em>Ammospermophilus junturensis</em> Shotwell and Russell, 1994</td>
<td>Ground squirrel</td>
</tr>
<tr>
<td><em>Ammospermophilus</em> sp. indet.</td>
<td>Ground squirrel</td>
</tr>
<tr>
<td><em>Eutamias</em> sp. indet.</td>
<td>Extinct chipmunk</td>
</tr>
<tr>
<td><em>Hystricops</em> sp. indet.</td>
<td>Extinct large beaver</td>
</tr>
<tr>
<td><em>Eucastor malheurensis</em> Shotwell and Russell, 1963</td>
<td>Extinct beaver</td>
</tr>
<tr>
<td><em>Microtus</em> sp. indet.</td>
<td>Extinct vole</td>
</tr>
<tr>
<td><em>Copemys</em> sp. cf. <em>C. esmeraldensis</em> (Wood) Lindsay, 1972</td>
<td>Extinct small field mouse</td>
</tr>
<tr>
<td><em>Copemys</em> sp. indet.</td>
<td>Extinct large field mouse</td>
</tr>
<tr>
<td><em>Copemys</em> sp. indet.</td>
<td>Extinct large field mouse</td>
</tr>
<tr>
<td><em>Ammospermophilus junturensis</em> Shotwell and Russell, 1994</td>
<td>Ground squirrel</td>
</tr>
<tr>
<td><em>Perognathus</em> sp. indet.</td>
<td>Extinct pocket mouse</td>
</tr>
<tr>
<td><em>Dipronyomys</em> sp. indet.</td>
<td>Extinct pocket mouse</td>
</tr>
<tr>
<td><em>Pliosaccomyx</em> sp. indet.</td>
<td>Extinct pocket gopher</td>
</tr>
<tr>
<td><em>Macrogynathus namus</em> Hall, 1930</td>
<td>Extinct birch mouse</td>
</tr>
<tr>
<td><em>Borophagus</em> sp. indet.</td>
<td>Short-face bone-crushing dog</td>
</tr>
<tr>
<td><em>Leptodonthomys</em> sp. indet.</td>
<td>Long-face bone-crushing dog</td>
</tr>
<tr>
<td><em>Vulpes</em> sp. indet.</td>
<td>Extinct fox</td>
</tr>
<tr>
<td><em>Leptarctus</em> sp. indet.</td>
<td>Extinct badgerlike mustelid</td>
</tr>
<tr>
<td><em>Hopiclis</em> sp. indet.</td>
<td>Extinct like honey badger</td>
</tr>
<tr>
<td><em>Sthenictis junturensis</em> Shotwell and Russell, 1963</td>
<td>Extinct otterlike musteline</td>
</tr>
<tr>
<td><em>Perognathus</em> sp. indet.</td>
<td>Extinct small cat</td>
</tr>
<tr>
<td><em>Gomphotherium</em> sp. indet.</td>
<td>Extinct four-tusk proboscidean</td>
</tr>
<tr>
<td><em>Mammut furlongi</em> Shotwell and Russell, 1963</td>
<td>Extinct two-tusk mastodon</td>
</tr>
<tr>
<td><em>Platybelodon barnumbrowni</em> (Barbour) Barrour, 1932</td>
<td>Extinct shovel-tusker</td>
</tr>
<tr>
<td><em>Cormophippurion sphenodus</em> (Cope) MacFadden and Skinner (1977)</td>
<td>Extinct three-toed horse</td>
</tr>
<tr>
<td><em>Tapiridae</em> gen. indet.</td>
<td>Extinct hornless rhinoceros</td>
</tr>
<tr>
<td><em>Aphelops</em> sp. indet.</td>
<td>Extinct piggly oreodon</td>
</tr>
<tr>
<td><em>Prosthenops</em> sp. indet.</td>
<td>Extinct camel</td>
</tr>
<tr>
<td><em>Merychys major</em> Leidy, 1858</td>
<td>Extinct large camel</td>
</tr>
<tr>
<td><em>Procamelus grandis</em> Gregory, 1942</td>
<td></td>
</tr>
<tr>
<td><em>Megatylopus</em> sp. indet.</td>
<td></td>
</tr>
</tbody>
</table>


land soils (Mollisols, Phaeozems, Chernozems; Retallack et al., 2002a), so that grasses were most likely bunch grasses.
Other paleosols with prominent relict bedding and large woody root traces supported plant formations early in ecological succession after disturbance. For example, Xaus paleosols of in-channel sandy bars probably supported riparian woodland, and Cmti and Abiaxi paleosols of levee siltstones probably supported riparian gallery woodland. Cmti paleosols are thinner, less sodic, and have more relict bedding than Abiaxi paleosols (Fig. 7), indicating a shorter period of vegetation growth and successional development. Early successional marsh ecosystems are represented by carbonaceous clayey paleosols with small root traces and relict bedding (Skaw pedotype). Swamp ecosystems are represented by lignitic paleosols with large woody root traces (Monana pedotype). The swamp paleosols have impure coals and charcoal and were probably seasonally dry, but there is no comparable evidence of seasonal exposure of marsh paleosols, which may have been associated with perennial watercourses.

6.2. Evidence from fossil plants

Only one kind of paleosol contained preserved fossil plants: weakly developed paleosols of the Skaw pedotype (localities 1–3 of Table 4). The best preserved fossil plants were found within laminated diatomites and siltstones (localities 4–7 of Table 4). These are interpreted as leaf litter of marsh soils (Skaw pedotype) and leaves blown into an oligotrophic lake, respectively. In all cases, there is a bias toward aquatic plants, such as water chestnut, arrowhead, and horsetail.

Despite these taphonomic limitations, the Stinkingwater, Weiser, and Unity floras include locally abundant fossil grasses and live oak comparable to those of summer-dry parts of California today. The Unity flora is most similar to interior live oak woodland of Barbour and Major (1988) and canyon live oak woodland of Sawyer and Keeler-Wolf (1995). These oak woodlands are at moderate elevation below the zone of abundant conifers on the southwestern slopes of the Sierra Nevada flanking the northern California Great Valley. The oak trees are largely evergreen and up to 15 m high in both closed and open formation. Their grassy understorey is brown and dry for most of the summer. Mesic elements in the Unity flora (Carya, Ptelea) are not found in modern Californian live oak woodlands. These middle Miocene holdover taxa are evidence that late Miocene dry woodlands of Unity were not exactly like Californian vegetation today.

No taxodiaceous conifers were found near Unity, where they would be expected considering charcoal and woody root traces in Monana paleosols. These swamp conifers are common in the coeval Stinkingwater flora of Juntura (Chaney and Axelrod, 1959) and the Weiser flora of Idaho (Dorf, 1936).

Fossil grasses are represented by sheathing leaves near Unity (Table 4) and by abundant pollen in coeval rocks of Idaho (Leopold and Wright, 1985). Miocene grasses of the Great Plains and Oregon are unlikely to have been floristically similar, given differences between fossil dicot floras of Oregon and the Great Plains (Thomasson, 1979; Graham, 1999; Strömberg, 2002) and differences between sodic–silicic Miocene paleosols of Oregon (Retallack, 2004) and calcareous Miocene paleosols of the Great Plains (Retallack, 1997b, 2001b). A comparable modern vegetation distinction is the western wheat grass (Agropyron spicatum) rangeland province of northern California and the Great Basin, and blue grama (Bouteloua gracilis) grasslands of the western Great Plains (Leopold and Denton, 1987).

6.3. Evidence from fossil animals

Quarry collections from Juntura formed the basis for Shotwell’s (1963) pioneering study of Clarendonian mammalian paleocommunities (Fig. 11), and these quarries were revisited in order to examine their paleosols (Fig. 3). Shotwell’s (1963) “pond bank community” came from Skaw paleosols (the only pedotype found), whereas his “savanna community” came largely from sandstones immediately overlying a Tutanik paleosol. His “savanna community” was mainly camels, hipparionine horses, two-tusker mastodon, and rhinoceros. The “pond bank community”, which from paleosol and sedimentological evidence appears more likely to have been riparian woodland, was mainly fish and turtles, but also had many beavers, as well as rabbits and a variety of rodents and insectivores (Table 5). The weak development and carbonaceous and manganiferous composition of Skaw paleosols are
Evidence of seasonally inundated streamside swales (Table 2), where fish were trapped and devoured by predators.

Contrary to Shotwell (1963), Bernor et al. (1988) concluded that coeval late Miocene (11 Ma) hipparionine horses of Austria were forest browsers, because forest leaves were found in associated lake beds. This view is not supported by the abundance of hipparionines in Tutukan paleosols and rarity in Skaw paleosols of Juntura and Unity (Fig. 11). Later studies of hipparionine tooth wear (Hayek et al., 1992) and associated mammalian fauna (Webb, 1983) also support Shotwell’s (1963) conclusion that hipparionines were grassland and grassy woodland mixed feeders. Open grassy terrane was also required by an extinct species of Old World vulture from Juntura (Table 5).

Proboscidians are now associated with mosaics of grassy and woody vegetation, which they maintain by destructive feeding (Owen-Smith, 1988). A diversity of proboscidean habitats may be expected from the high diversity of Clarendonian proboscideans, including eight genera in North America and three in the Unity and Juntura faunas (Tables 4–5). Shovel-tusker (Platybelodon) are commonly portrayed as aquatic shovelers, although tusk wear patterns indicate scraping against bark, twigs, pebbles, or other hard debris (Lambert, 1992). Four-tuskers (Gomphotherium) also show tusk wear from bark stripping and digging, and two-tusk mastodonts (Mammut) are known from Pleistocene stomach contents to have eaten conifer needles, cones, and grass (Lambert and Shoshani, 1998). Associated paleosols are Abiaxi for Gomphotherium, Tutukan for Mammut, and probably Skaw for cf. Platybelodon (uncertainty comes from poor exposure and inexact location). The inferred paleoenvironments of these paleosols (Table 3) cannot be claimed to represent habitat preferences based on only one or two specimens of each species.

Paleosols also support the idea that late Miocene rangelands of the Pacific Northwest were distinct from those of the Great Plains (Leopold and Denton, 1987). Late Miocene summer-dry siliceous alfisol and aridisol paleosols of Oregon (Retallack et al., 2002a) can be contrasted with summer-wet Mollisols of Nebraska and Montana (Retallack, 1997b, 2001b). The hipparionine province of the Pacific Northwest was distinct from a Pliohippus province of southern California and the Great Plains during the Clarendonian (Shotwell, 1963). Late Miocene (7 Ma) grazing Pliohippus in Oregon appeared in calcareous Mollisol paleosols reflecting an incursion of summer-wet Gulf Coast air masses into eastern Oregon at that time (7.3 Ma; Retallack et al., 2002a). The tridactyl mixed-feeding hipparionines may reflect rigors of summer-dry rangeland vegetation botanically and structurally distinct from summer-wet rangeland of monodactyl grazing Pliohippus. Comparable differences in vegetation and seasonality may also explain the numerical dominance of foregut-fermenting ruminant camels over hindgut-fermenting horses at Juntura (Fig. 11) and other western Clarendonian faunas (Tedford and Barghoorn, 1993), whereas horses dominated Clarendonian assemblages of Nebraska and Texas (Webb, 1983). Muzzle morphology of the camels Procamelus and Megatylopus has been taken to indicate that they were browsers, perhaps mixed feeders, but not grazers (Dompierre and Churcher, 1996). Similar dominance of ruminants over perissodactyls is found in Kenyan (Retallack et al., 2002b) and Greek (Solounias, 1981) Miocene faunas. By reprocessing their cud, ruminants are sustained by less-nutritious forage than perissodactyls of comparable body sizes (Janis et al., 1994). Summer-dry seasons in Kenya, Greece, and western...
North America selected for ruminant-dominated faunas able to cope with seasonal shortages of forage, whereas equid-dominated faunas of the Great Plains and Gulf Coast were supported by year-round forage.

7. Implications for global change

7.1. Neogene paleoclimate in eastern Oregon

A detailed Neogene paleoclimatic time series (Fig. 12A–C) can be constructed by extending the results from paleosols presented here to other paleosols, including paleosols from the early Miocene upper John Day Formation near Kimberly, Oregon (Retallack, 2004) dated by extrapolation from $^{40}$Ar/$^{39}$Ar radiometry (Fremd et al., 1994), the middle Miocene Mascall Formation and Columbia River Basalts near Dayville, Oregon (Bestland and Krull, 1997; Sheldon, 2003) dated by magnetostratigraphy and radiometry (Draus and Prothero, 2002), the latest Miocene Rattlesnake Formation near Dayville, Oregon (Retallack et al., 2002a) dated by magnetostratigraphy (Hoffman and Prothero, 2002), the Mio–Pliocene Ringold Formation near Pasco and Taunton, Washington (Gustafson, 1978; Smith et al., 2000) dated by magnetostratigraphy (Gustafson, 1985; Morgan and Morgan, 1995), and the Palouse Loess near Helix, Oregon, and Washtucna, Washington dated by magnetostratigraphy, tephrochronology, and thermoluminescence dating (Busacca, 1989, 1991; Tate, 1998; Blennikov et al., 2002).

The resulting curve shows great variation in both precipitation (Fig. 12A–B) and temperature (Fig. 12C) through time and is comparable with variation in carbon isotopic composition of benthic marine foraminifera (Fig. 12F). Also comparable are time series of North American mammalian diversity, which was
higher during warmer and wetter times than during cooler and drier times (Alroy et al., 2000; Janis, 2002).

### 7.2. Mismatch of Neogene terrestrial and marine oxygen isotopic records

Neither paleoclimatic fluctuations from paleosols (Fig. 12A–C) nor oxygen isotopic fluctuations from mammal teeth (Fig. 12E) correspond with fluctuations in oxygen isotopic composition of benthic marine foraminifera (Fig. 12F). Marine and terrestrial carbonate oxygen isotopic records do not correspond during the late Paleocene (Koch et al., 2003) or late Permian (MacLeod et al., 2000) either. Although marine oxygen isotopic records have been regarded as a global paleoclimatic standard (Zachos et al., 2001), their mismatch with continental records is a fundamental problem for global change studies.

The mismatch between North American mammalian evolutionary dynamics and marine oxygen isotopic records has been discussed at length by Alroy et al. (2000), who attribute the mismatch to biotic insensitivity to climatic forcing. This is contrary to the observation of iterative evolution of mammal ecomorphs within North American land mammal “ages” which average 2.3 Myr in duration (Martin, 1994; Meehan and Martin, 1994; Meehan, 1996, 1999). The sediments of each mammal age in the Great Plains are bounded at the base by erosional downcutting, interpreted as evidence of a wet climatic phase, and terminated by shallow calcic paleosols or a caprock caliche, interpreted as evidence of a dry climatic phase (Schultz and Stout, 1980). I have made comparable observations in Montana and Oregon (Retallack, 1997b, 2001a; Retallack et al., 2002a).

Late Oligocene mammalian communities of Oregon also show alternation between sagebrush steppe with Hypertragulus and wooded grassland communities with Nanotragulus on Milankovitch time scales (41–100 ka) within paleosols with differing depth to calcic horizons reflecting alternating semiarid and subhumid paleoclimates (Retallack, 2004; Retallack et al., 2004). Finally, there is annual variation in oxygen isotopic composition of growth bands within Miocene mammalian tooth enamel (Fox, 2001; Kohn et al., 2002). North American mammals show clear sensitivity to climate on time scales ranging from evolutionary to ecological contrary to Alroy et al. (2000).

A possible reason for mismatch of marine oxygen and continental paleoclimatic records is the Cenozoic evolution of C₄ photosynthetic pathways in land plants. The C₄ (Hatch–Slack) pathways of tropical grasses result in isotopically heavier plant carbon than C₃ (Calvin–Benson) pathways (Cerling et al., 1997), and this pathway increases oxygen isotopic values of plant as well because operating on CO₂ (Farquhar et al., 1993). The antiquity of the C₄ pathway has been estimated from phylogenetic analyses at 25 Ma (Kellogg, 1999), from isotopic analyses of paleosols at 15 Ma (Kingston et al., 1994; Morgan et al., 1994), and from isotopic and anatomical studies of permineralized fossil grasses at 12.5 Ma (Nambudiri et al., 1978; age revised by Whistler and Burbank, 1992). Isotopic studies of paleosol carbonate and mammalian teeth show that C₄ grasses became widespread in the southern Great Plains after 6.6 Ma but were never common in Oregon or the northern Great Plains (Cerling et al., 1997; Passy et al., 2002; Fox and Koch, 2003). C₄ grasses avoid cool and summer-dry climates today (Sage et al., 1999). The spread of C₄ grasses after 6.6 Ma is too late to explain the earlier divergences of marine carbon and oxygen isotopic records (Fig. 12F), mismatch of marine and terrestrial oxygen isotopic records (Fig. 12A–C,F), and the generally upward trend of marine oxygen isotopic values (Fig. 12F).

A more likely explanation for mismatch of the oceanic oxygen isotopic and continental paleoclimatic proxies is isotopic depletion as rain shadows spread and intensified with uplift of the Cascade volcanic and other western mountain ranges (Kohn et al., 2002). The idea of Cenozoic drying by means of an orographic rain shadow has been a traditional explanation for pronounced Neogene aridity evident from fossil plants (Chaney, 1948; Ashwill, 1983). The growth of the Oregon Cascades can be inferred from numerous radiometrically dated eruptions (Fig. 12D; McBirney et al., 1974; Priest, 1990), but this does not match either marine or continental records (Fig. 12A–F). The Cascade and Klamath Mountain rain shadow has been important in creating generally dry long-term climate in eastern Oregon since the late Oligocene (30 Ma; Retallack et al., 2000). Rocky Mountain barriers isolated eastern Oregon from Gulf Coast cyclonic circulation since at least early Miocene (19 Ma), as indicated by the
appearance of sodic–silicic rather than calcic paleosols (Retallack, 2004) and very different oxygen isotopic composition of fossil horse teeth in Oregon and Nebraska (Fig. 12E). Fossil floras from the northern Great Basin indicate high altitudes since at least the early Miocene, for example, 2100 m for the middle Miocene 49 Camp flora of northwestern Nevada (Wolfe et al., 1997). Eastern Oregon was elevated during Miocene initial eruptions of the Yellowstone hot spot (Humphreys et al., 2000).

The most important reason for a mismatch of oceanic and continental records is isotopic enrichment of oceanic oxygen by growth of isotopically depleted polar ice caps, which has long been recognized to compromise direct paleotemperature interpretation of marine oxygen isotopic values (Zachos et al., 2001). This factor is not unrelated to the growth of mountains with their expanding rain shadows, because montane ice caps enrich the isotopic composition of atmospheric oxygen as well. Ice cap fluctuation in volume introduces a highly variable bias into paleotemperature interpreted from the oxygen isotopic record of marine benthic foraminifera. This bias can be corrected using independent estimates of paleotemperature from Mg/Ca ratios of foraminifera (Lear et al., 2000), but this is compromised by dissolution and other forms of diagenesis (de Villiers, 2003), and by crustal recycling and other long-term influences on ocean chemistry (Veizer et al., 2000).

7.3. Neogene greenhouse–icehouse fluctuation and its causes

Greenhouse mechanisms for Neogene paleoclimatic variation are suggested by correspondence of the Oregon paleosol sequence (Fig. 12A–C) with the marine signal of carbon sequestration inferred from benthic foraminiferal δ13C decrease and carbon oxidation inferred from benthic foraminiferal δ13C increase (Fig. 12F). High carbon isotopic values in the middle Miocene ocean correspond to warm-wet conditions in Oregon, whereas low carbon isotopic values in the Plio–Pleistocene ocean corresponded to cool-dry conditions in Oregon. High carbon isotopic values in marine foraminifera reflect diminished burial of chemically reduced carbon and higher atmospheric CO2 levels due to some combination of oceanic respiration, volcanic emission, or impact destruction of biomass. A middle Miocene high in atmospheric CO2 has been confirmed by stomatal index estimates of atmospheric CO2 (Kürschner et al., 1996; Retallack, 2001c, 2002). Oceanic proxies of atmospheric CO2 do not show such variation (Pagani et al., 1999; Pearson and Palmer, 2000) but may be compromised by volatility of isotopic composition and runoff (Retallack, 2002). Despite these problems, a useful working hypothesis is control of paleoclimate by a varying greenhouse effect from changing atmospheric concentrations of CO2.

Plausible mechanisms for transient climatic warming include impact, volcanism (Coutillot, 2002) and methane clathrate release (Kennett et al., 2002). There are numerous large (>10 km diameter) Neogene craters: Haughton, Canada (24 km, 23±1 Ma), Ries, Germany (24 km, 15±0.1 Ma), Karla, Russia (10 km, 5±1 Ma), El’gygytgyn, Russia (18 km, 3.5±0.5 Ma), Bosumtwi, Ghana (10.5 km, 1.03±0.02 Ma), and Zhamanshin, Kazakhstan (14 km, 0.9±0.1 Ma; Dressler and Reimold, 2001; website http://www.unb.ca/passe/impactdatabase accessed Jan. 10, 2003). Time series of volcanic eruptions are also strongly episodic (McBirney et al., 1974), with some exceptionally large eruptions, such as the middle Miocene Columbia River flood basalts of Oregon (Wignall, 2001). Columbia River Basalts and Ries Crater impact may have been involved in the middle Miocene peak of global warmth, but why was there not a comparable warm spike after the 23-Ma Haughton Crater or a better correlation between North American paleoclimate and Cordilleran volcanism (Fig. 12D). Because of methane’s distinctively low carbon isotopic values, such methane outbursts are suspected when carbon isotopic compositions fall by more than 4‰ (Jahren et al., 2001). There are no negative excursions of this magnitude in Neogene carbon isotopic records (Fig. 12F), thus methane is unlikely to be the whole explanation either.

A counterpoint to transient carbon-oxidizing events was progressive long-term Neogene global cooling, perhaps related to changing oceanic current configuration (Broecker, 1997), mountain uplift (Raymo and Ruddiman, 1992), or biotic intensification of weathering with sod–grassland coevolution (Retallack, 2001a). Both mountain uplift and increased oceanic thermohaline circulation create oligotrophic alpine and polar marine communities, with less carbon sequestra-
tion potential, but grasslands provided more organic and nutrient-rich runoff to oceanic phytoplankton (Retallack, 2001b). Grasslands in addition had greater amounts of soil carbon, higher albedo, and lower rates of transpiration than the woodlands and shrublands they replaced in subhumid to semiarid regions of the world (Nepstad et al., 1994; Jackson et al., 2002), and this also could have contributed to global drying and cooling (Retallack, 2001b). This mechanism of global change is biological, because it is driven by coevolution of grasses and grazers, with crumb-textured, organic, fertile soils nourishing hypsodont ungulates capable of withstanding abrasion from dusty, siliceous grasses (Retallack, 2001a). The progressive evolution and spread of crumb-textured soils, siliceous grasses, and hypsodont ungulates proceeded against a backdrop of volatile climatic change, and both short-sod and tall-sod grassland paleosols appearing at times of relatively high temperature and humidity (Fig. 12A). Grasslands did not merely fill in dry times and places (Kohn et al., 2002) but were a biological force for climatic cooling and drying in their own right (Retallack, 2001b).

Paleoclimatic warm spikes due to volcanism, bolide impact, and methane clathrate release may also have been exacerbated by newly coevolved Cenozoic grasslands, because ruminants liberate CO₂ and CH₄. The middle Miocene thermal optimum was a time of unusually high diversity of North American plants and mammals, in part due to evolution within high-productivity grassland–woodland mosaics in a warm-wet CO₂ greenhouse (Janis, 2002), and in part due to immigration from Asia through high-latitude land bridges (Wolfe and Tanai, 1980; Kohn, 1997; Lambert and Shoshani, 1998; Webb, 1998). There were more grazing mammal species then ever before, and also more browsers than before or after (Janis et al., 2000). The middle Miocene greenhouse peak was preceded and followed by several other episodes of greenhouse paleoclimate of lesser magnitude, including early Clarendonian communities described here. Warm-wet conditions of greenhouse transients, even if initiated by tectonic or impact forcing, would be enhanced by increased mammalian diversity and cropping of vegetation. Cold-dry conditions follow carbon and water sequestration by grassland soils, their high albedo and burial of carbon-rich grassland soil crumbs. Paleoecological cycles on million-year time scales could reflect alternating evolutionary advances by plants and animals. Such reciprocating coevolution is apparent from current estimates of the times for evolution of bunch grassland paleosols (33 Ma), short-sod grassland paleosols (19 Ma), and tall-sod grassland paleosols (7.3 Ma; Retallack, 1997b, 2001b, 2004; Retallack et al., 2002a), which alternate with evolution of horse tridactly (36 Ma), cursoriality (25 Ma), hypsodonty and springing gait (17 Ma), monodactyly, wide muzzles, and knee-locking (12 Ma), and large size and passive-stay shoulder (5 Ma; MacFadden, 1992; Janis and Wilheim, 1993; Hermanson and MacFadden, 1992, 1996). This schedule is compatible with a coevolutionary process long envisioned for grassland ecosystems (Jacobs et al., 1999). Global change involves more than just impact or volcanic forcing, but a complex interplay of factors for which detailed proxies are now becoming available (Fig. 12). These various records will have to become even more detailed before answers become clear.

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