Early Eocene biotic and climatic change in interior western North America

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ABSTRACT

Imprecise correlation of the marine and terrestrial fossil records has been a major obstacle to understanding migration and extinction of continental biotas and early Cenozoic climate change. New $^{40}Ar/^{39}Ar$ data from the Willwood Formation in the Bighorn Basin of Wyoming establish an age of 52.8 ±0.3 Ma for earliest Lostcabinian (late Wasatchian) faunas and coeval early Eocene floras. Strata just beneath earliest Wasatchian faunas can be correlated with the NP9/NP10 boundary in marine sedimentary units, which has an interpolated age of ~55.7 Ma. This new information allows us to estimate the durations of the Wasatchian (~5 m.y.) and the Lostcabinian (~2 m.y.) and shows that the continental biotas are coeval with the acme of Cenozoic warmth inferred from $\delta^{18}O$ measurements of foraminifera. From 58 to 50 Ma, paleoclimate in the continental interior at about 45°N was warm and equable, but patterns of temperature change inferred from continental floras do not track precisely the marine $\delta^{18}O$ record.

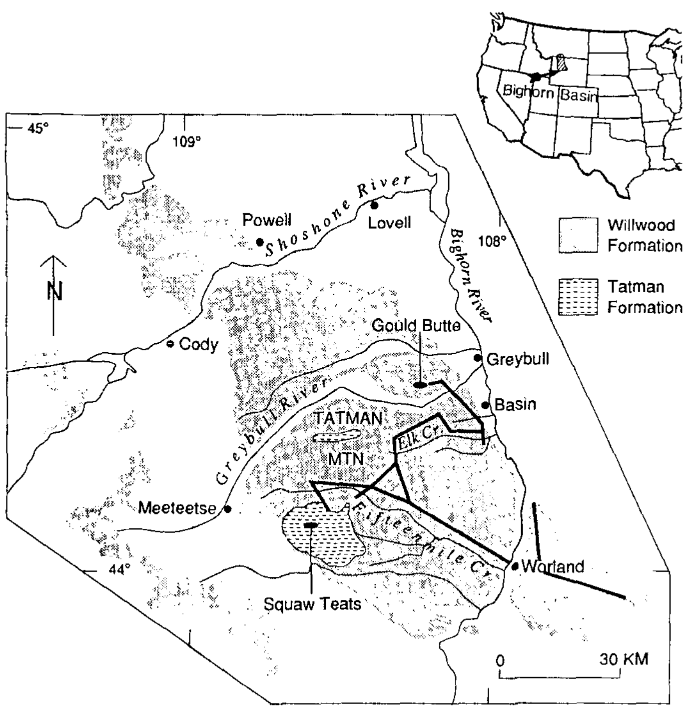

Figure 1. Location of Bighorn Basin and outcrop area of Willwood and Tatman Formations. Heavy black lines are transects of major sections.

INTRODUCTION

The Bighorn Basin of Wyoming has a thick and abundantly fossiliferous sequence of upper Paleocene–lower Eocene continental rocks (Fig. 1). Strata of the upper Fort Union and lower Willwood Formations contain faunal assemblages from which the Clarkforkian North American Land Mammal Age (NALMA) and the Greybullian subage of the Wasatchian NALMA (Wood et al., 1941; Rose, 1980) were recognized. All Paleocene and early Eocene NALMAs and subages are represented in the Bighorn Basin (Gingerich et al., 1980; Schankler, 1980). The Fort Union and Willwood Formations also have produced a stratigraphically dense paleobotanical record (Hickey, 1980; Wing, 1980, 1984; Farley, 1989). The isotopic age and palynological correlation presented here are important for establishing the chronology of these strata and their correlation to records in the deep sea and on other continents.

STRATIGRAPHIC AND BIOSTRATIGRAPHIC FRAMEWORK

The upper Fort Union and Willwood Formations in the central and eastern Bighorn Basin are about 1 km thick and represent mostly aggradation on floodplains, with shorter intervals of fluvial channel, back-swamp, and small-scale lacustrine deposition. The relative stratigraphic positions of more than 1400 vertebrate localities and 100 plant localities have been established through bed tracing and section measuring (summarized by Bown et al., 1991). As with most fluvial sequences, hiatuses representing 10^2 to 10^4 yr are abundant, and they are usually represented by paleosol development or relatively minor erosional surfaces (Bown and Kraus, 1981). Erosional surfaces with tens of metres of relief are known within the Willwood Formation, but most of them are in the 430–530-m interval of our composite section (Fig. 2; Bown, 1984; Bown et al., 1991). The general homogeneity of upper Fort Union and Willwood strata is consistent with fairly uniform long-term rates of deposition and linear interpolation of time between tie points. Consequently we use stratigraphic thickness as a general approximation of time over stratigraphic intervals on the order of 100 m.

All fossil mammals currently known from our composite section (Fig. 2) are characteristic of the Wasatchian NALMA. The stratigraphically lowest faunas come from the ~0 m level of the composite section in the upper part of the Fort Union Formation near Gould Butte (Fig. 1), but these assemblages contain *Hyracotherium* and other early Wasatchian indicators (Wing and Bown, 1985). Mammalian fossils from the lowest part of the Willwood Formation near the town of Worland are similar to basal Wasatchian faunas from the Clarks Fork region in the northern Bighorn Basin (Bown, 1979; Wa0 of Gingerich, 1989); this similarity suggests that the Clarkforkian/Wasatchian boundary is at or not far below the contact of the Fort Union and Willwood Formations in the central and southeastern Bighorn Basin.

Faunas containing the Lostcabinian index fossil *Lambdotherium* are known from the 580–750 m level of the composite section, which includes the bottom 100 m of the Tatman Formation (Fig. 2; Schankler, 1980; Bown et al., 1991). The transition to Bridgerian faunas has not been documented in our composite section, possibly because the upper Tatman Formation is poorly fossiliferous. Biostratigraphically important first and

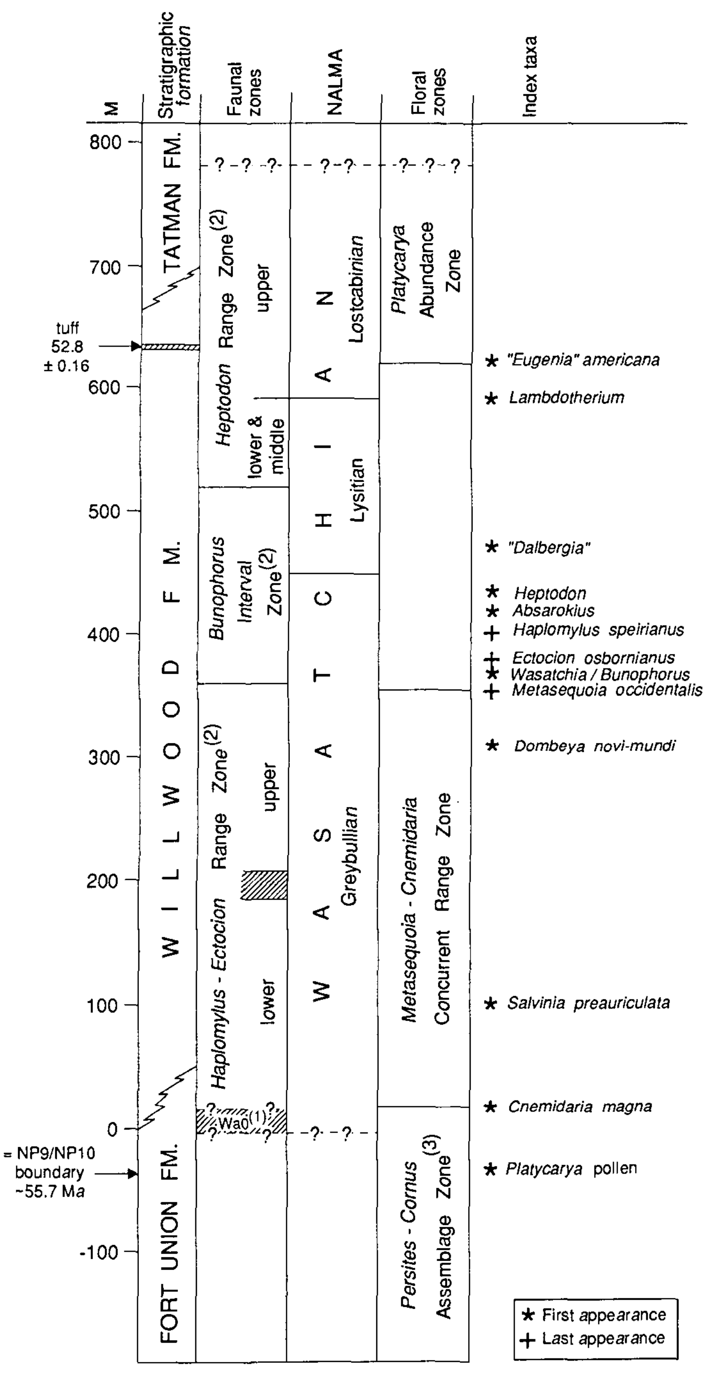

Figure 2. Composite section of upper Fort Union and Willwood Formations, showing mammalian and plant biostratigraphic zonation, dated tuff, point of correlation to standard marine zonation, and significant FADs and LADs. (1) Basal Wasatchian (Wa0) of Gingerich (1989). (2) Mammalian zonation modified from Schankler (1980). (3) Floral zone of Hickey (1980). Stratigraphy and ranges of index taxa are from Bown et al. (1991).

last appearance datums (FADs and LADs) of mammals and plants, biostratigraphic zones, a generalized stratigraphic section, and the position of the tuff dated at 52.8 ±0.3 Ma are shown in Figure 2. The new data, in conjunction with previously reported radiometric ages for terminal Wasatchian faunas (Krishtalka et al., 1987) and an interpolated age of ~55.7 Ma for the base of the Wasatchian (see below), allow us to calculate durations of about 5 m.y. for the Wasatchian and about 2 m.y. for the Lostcabinian.

We recognize three megafloral zones in our composite section. Assemblages from −100 to 10 m in the section, contain *Persites*, *Cornus*, "*Carya*" *antiquorum*, *Porosia verrucosa*, and other indicators of the *Persites-Cornus* Assemblage Zone, which extends downward to at least the base of the Clarkforkian in the northern Bighorn Basin (Hickey, 1980). The FAD for the tree fern *Cnemidaria* at the 10 m level of the composite section marks the bottom of the *Metasequoia-Cnemidaria* Concurrent Range Zone, which terminates with the local extinction of the conifer *Metasequoia* at the 353 m level. Megafloras between 353 and 621 m in the composite section are not zoned, although "*Dalbergia*," an important form in later early Eocene floras, has its FAD at 468 m. Megafloras from the 621 m level and higher are dominated by *Platycarya castaneopsis* and also have taxa (*Aleurites*, "*Eugenia*," and "*Populus*" *wyomingiana*) that elsewhere are associated with Lostcabinian and Bridgerian faunas (MacGinitie, 1969, 1974). These floras are referred to the *Platycarya* Abundance Zone, which continues to at least the 740 m level of the composite section. Floras above this level are not well known in our section. These megafloral zones probably apply regionally, because in North Dakota and other parts of Wyoming, *Metasequoia-Cnemidaria* Concurrent Range Zone floras are associated with early Wasatchian mammals and *Platycarya* Abundance Zone floras with late Wasatchian mammals.

No formal palynostratigraphic zonation exists for our composite section, but most upper Fort Union Formation samples yield abundant *Momipites* and *Caryapollenites* similar to the P6 palynofloras described by Nichols and Ott (1978) in the Wind River Basin (M. B. Farley, 1989, personal commun.). The FAD for *Platycarya* pollen is at −35 m in the composite section (M. B. Farley, 1989, personal commun.); this is 135 m below the position given by Wing (1984). The FAD for *Platycarya* coincides almost precisely with the NP9/NP10 boundary across much of North America (Frederiksen, 1979, 1980, 1983; Frederiksen et al., 1982), and therefore we consider its FAD in Wyoming to be approximately isochronous with the NP9/NP10 boundary, which recently has been assigned a latest Paleocene age (Aubry et al., 1988).

Although Aubry et al. (1988, Fig. 7) presented an interpolated age of ~57.6 Ma for the NP9/NP10 boundary, this figure may be substantially too old. $^{40}Ar/^{39}Ar$ data from sanidine in the "−17 ash" of the Fur Formation (Mo Clay) in Denmark, which is high in the *Apectodinium hyperacanthum* dinoflagellate zone and correlates with mid-NP10, indicate an age of 55.1 ±0.3 Ma (Obradovich, 1988, and unpublished data). Interpolation based on this age (Obradovich, 1988, Fig. 5) yields an age of about 55.7 Ma for the NP9/NP10 boundary. Using the FAD of *Platycarya* pollen for correlation from the marine to continental sections, we infer the −35 m level of our Bighorn Basin composite section to be ~55.7 Ma.

GEOCHRONOMETRY

Samples for $^{40}Ar/^{39}Ar$ dating were taken of a 1-m-thick, crystal-rich bentonite at 634 m in our section on the north side of the Squaw Teats Divide. (Even though this is a continental unit, we use the term "bentonite" because all of the glass in the tuff has been converted to clay.) Sanidine was recovered by wet sieving, use of heavy liquids, and magnetic separation. Examination of the sanidine concentrate employing focal masking techniques revealed about one grain of microcline per thousand of sanidine. The final sample concentrate of 100 mg was achieved by removing

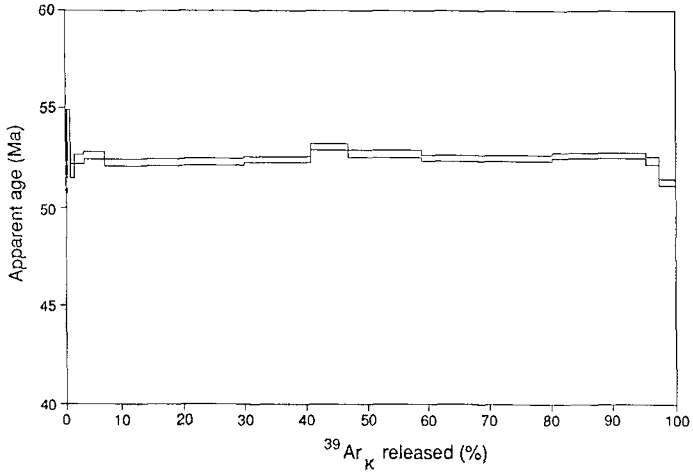

Figure 3. Age spectrum for $^{40}Ar/^{39}Ar$ analysis of sanidine from bentonitic bed at 634 m in composite section. $^{39}Ar_K$ is argon derived from potassium.

cloudy or weathered crystals by hand, under a stereomicroscope using both transmitted and reflected light. This sample of optically clear grains was washed in 6% HF for 5 min, rinsed in distilled water, and then irradiated for ~35 h in the TRIGA reactor at the U.S. Geological Survey facilities in Denver. Argon was extracted with a double-walled vacuum furnace (modified from the design of Staudacher et al., 1978), and the purified argon was analyzed by using a static rare-gas mass spectrometer. Corrections were made for the interfering argon isotopes resulting from neutron reactions with potassium and calcium (Dalrymple et al., 1981). The Minnesota hornblende standard (MMhb-1; Sampson and Alexander, 1987), which has an age of 520.4 Ma, was used to monitor the neutron flux. Although the weighted mean plateau age of $52.8 \pm 0.3_2$ Ma (± 2 standard error of the mean; Fig. 3) is based on just 51% of the total potassium-derived ^{39}Ar that was extracted, the total-gas age of 52.6 Ma is analytically indistinguishable.

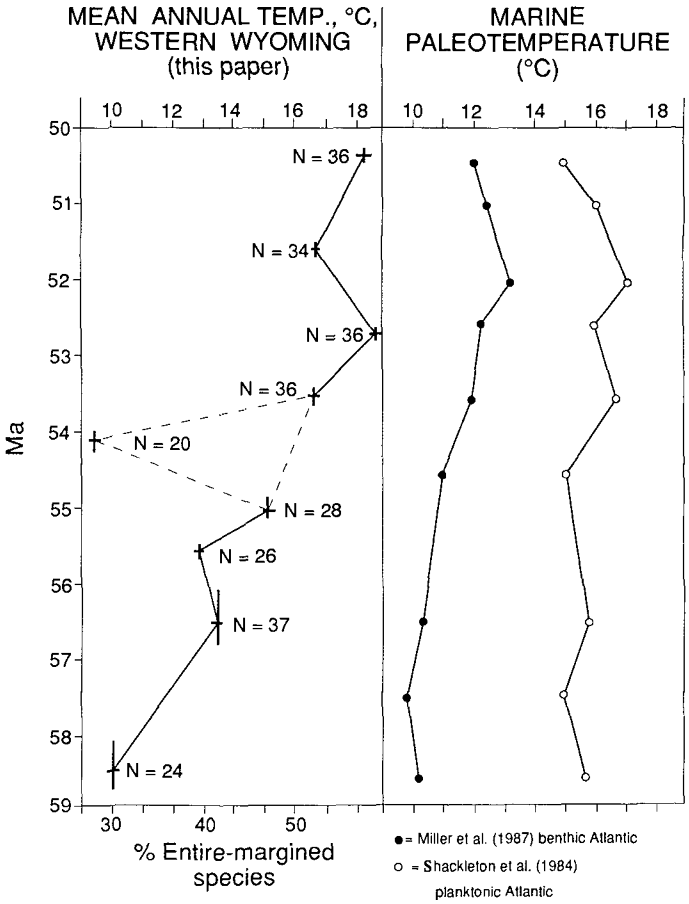

Figure 4. Temperature curves derived from leaf-margin analysis of Wyoming floras (left) and $\delta^{18}O$ measurements of benthic and planktonic foraminifera (right), plotted against time scale of Harland et al. (1989). N = number of dicotyledonous species used in each estimate; vertical length of crosses in floral curve is proportional to stratigraphic thickness from which collections were made.

COMPARISON OF TERRESTRIAL AND MARINE CLIMATE RECORDS

$^{18}O/^{16}O$ studies of foraminifera show that the early Eocene was globally the warmest part of the Cenozoic (e.g., Miller et al., 1987; Prentice and Matthews, 1988); this generally is confirmed by continental floras and faunas (e.g., Wolfe, 1978; Hutchison, 1982). Comparisons of the marine and terrestrial records (Wolfe and Poore, 1982), however, have been limited by low temporal resolution and uncertain correlations. The new data presented here permit a higher level of temporal resolution of temperature change through the 59–50 Ma period than has previously been possible for continental interiors.

Estimates of terrestrial paleotemperature (Fig. 4) are derived from leaf-margin analysis, which relies on the strong positive correlation in living vegetation between the proportion of species in a local flora that have entire-margined leaves and the mean annual temperature under which the flora grows (Wolfe, 1979). Paleotemperature estimates for seven time periods are available for the late Paleocene through early Eocene of the Bighorn Basin. The temperature estimates for ~58.5 and ~56.5 Ma were made by Hickey (1980); the ages proposed here rely on paleomagnetic correlations (Butler et al., 1980) and the time scale of Harland et al. (1989). The older temperature estimate is based on 20 localities that correlate with lower to middle Tiffanian faunas (~58–59 Ma), the younger on 31 localities that correlate with Clarkforkian faunas (~55.7–57 Ma). Both estimates are based on floral zone averages and may span 1 m.y. or more. The remaining five estimates from the Bighorn Basin are each based on floras collected from stratigraphic intervals of less than 50 m, each probably representing 0.2–0.3 m.y. Ages of these floras were estimated by linear interpolation based on their stratigraphic levels. The youngest two paleotemperature estimates come from radiometrically dated latest Wasatchian equivalent (~51 Ma) and Bridgerian equivalent (~50 Ma) floras in the western Wind River Basin (Wind River and Kisinger Lakes Floras of MacGinitie, 1974), about 160 km to the southwest of the central Bighorn Basin. The entire set of floral samples was derived from similar depositional environments, dominantly wet distal flood plains (Hickey, 1980; Wing, 1984), and from a confined geographic region, so most of the change in leaf-margin percentage should reflect change in paleotemperature rather than changes in taphonomic processes.

Paleobotanical estimates of mean annual temperature (Fig. 4) show an increase from 13 °C at 56.5 Ma to between 15 and 16 °C at about 55 Ma. The drop to 9 °C at about 54 Ma is probably an artifact of poor sampling; only 20 species are known from that interval. Successive samples above this level indicate a mean annual temperature of 16–18 °C between 54 and 50 Ma. Plants incapable of withstanding prolonged freezes, such as tree ferns, palms, and cycads, are found in many assemblages from 56.5 to 50 Ma. Although both terrestrial and marine temperature proxies attain maximum values during the early Eocene, most

$\delta^{18}O$ temperature records peak at about 52 Ma (time scale of Harland et al., 1989) and then decline (e.g., Shackleton et al., 1984; Miller et al., 1987; Prentice and Matthews, 1988). In contrast, the paleobotanical temperature estimates maintain peak levels until at least 50 Ma. The significance of this discrepancy is not clear at present. Additional paleoclimatic records from continental interiors are necessary to determine whether the decoupling of marine and terrestrial temperature curves from 52 to 50 Ma reflects sampling problems in the terrestrial flora, regional climatic evolution in western interior North America, or a global phenomenon.

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