

# EVOLUTION OF TERTIARY MAMMALS OF NORTH AMERICA

VOLUME 1:  
TERRESTRIAL CARNIVORES, UNGULATES,  
AND UNGULATELIKE MAMMALS

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# Part I: Overview of context for the evolution of North American Tertiary mammals

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## 1 The chronological, climatic, and paleogeographic background to North American mammalian evolution

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### INTRODUCTION

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This chapter summarizes changes in North American climate and landscape through the Tertiary. Since Webb's (1977) account, there have been significant changes in the interpretation of North American paleogeography and paleoclimate, largely due to more detailed stratigraphic work and an increased understanding of global climatic history.

Most significantly, our understanding of mammalian evolution is based on a chronological framework that has been radically changed. Although we still use the relative sequence of North American faunas first codified over fifty-five years ago by the Wood Committee (Wood et al., 1941), its geochronology and correlation with the global time scale has been fundamentally altered by breakthroughs in  $^{40}\text{Ar}/^{39}\text{Ar}$  dating and magnetic stratigraphy. Some faunas that were once thought to be Eocene are now considered Paleocene; others long considered early Oligocene are now late Eocene; much of the North American "early Miocene" is actually late Oligocene, and most of the original "Pliocene" of the Wood Committee is now late Miocene. Some of these changes are reflected in the most recent summary of North American biochronology (Woodburne, 1987), but many others have occurred in the last few years. Because much of this new geochronology is still in progress, I cannot give a complete and detailed description here. However, this chapter summarizes the major changes that have occurred and describes how they affect the correlation of North American fauna and climatic changes with the rest of the world.

### PRINCIPLES OF NORTH AMERICAN LAND CHRONOLOGY

In North America, the only practical method of correlating and dating most Cenozoic terrestrial deposits has been with land mammals. At the turn of the century, William Diller Matthew and Henry Fairfield Osborn (Matthew, 1899; Osborn and Matthew, 1909; Osborn, 1907, 1910, 1929) attempted to create biostratigraphic zonations of the North American terrestrial sequence based on

principles followed by European stratigraphers at the time (Tedford, 1970). But the good beginning established by Osborn and Matthew was essentially lost because the next generation of vertebrate paleontologists ignored their pioneering work.

Unlike most marine invertebrates, fossil mammals typically occur in localized fossiliferous horizons or in isolated pockets or quarries without apparent stratigraphic superposition. As a result, traditional biostratigraphic methods developed by European invertebrate paleontologists (based on detailed stratigraphic ranges of fossils in measured sections) were not widely followed by North American vertebrate paleontologists. In 1937, the Vertebrate Paleontology Section of the Paleontological Society appointed a committee to clarify the confusion over correlation and to adopt a terminology that could be widely used. Known as the Wood Committee (Wood et al., 1941), it was chaired by Horace E. Wood II, and included four other vertebrate paleontologists (Edwin H. Colbert, John Clark, Glenn L. Jepsen, and Chester Stock), plus paleobotanist Ralph Chaney and invertebrate paleontologist J. B. Reeside.

The Wood Committee's "Provincial Ages" were a complex hybrid of local rock units and time units delineated by first and last occurrences of mammalian genera. As Tedford (1970) and Woodburne (1977, 1987) have pointed out, these units were not true time-stratigraphic ages, which must be based on biostratigraphic zones and stages (according to Western stratigraphic codes such as the 1983 North American Code of Stratigraphic Nomenclature). Because they were not true stratigraphic ages, the North American land mammal "ages" should properly be put in quotes in all publications. Instead, the Wood Committee's methods resemble a system called "biochronology," first proposed by H. S. Williams (1901), which attempts to reconstruct the sequence of occurrences of taxa without necessarily documenting every event in a stratigraphic section. Demonstration of stratigraphic superposition was noted when available, but detailed work in the European biostratigraphic tradition was not considered essential to their "provisional" system.

In spite of their loose characterization, the land mammal "ages" of the Wood Committee worked quite well for over forty years, mostly because mammals do evolve and disperse very rapidly

(Savage, 1977). However, problems eventually arose. For example, the Chadronian was originally defined both on the co-occurrence of *Mesohippus* and brontotheres, and also on the limits of the Chadron Formation. At the time, the last occurrence of brontotheres was thought to coincide with the top of the Chadron Formation, so there was no conflict. When Morris Skinner documented brontothere specimens in rocks correlative with the overlying Orella Member of the Brule Formation, however, the difficulty with defining the Chadronian both biochronologically and lithostratigraphically became apparent (Prothero, 1982; Emry, Bjork, and Russell, 1987; Evanoff, Prothero, and Lander, 1992). Yet many of Skinner's contemporaries could not accept this evidence because the misconception that rock units could be treated as if they were time units was widespread among paleontologists of that generation.

Since the 1950s, another generation of paleontologists (see Savage, 1955, 1962, 1977; Tedford, 1970; Woodburne, 1977, 1987) has tried to bring vertebrate paleontology back to classical biostratigraphic methods. Trained in modern stratigraphic thinking, vertebrate biostratigraphers now appreciate the possibility that rock units can be time transgressive over distance (Shaw, 1964; Prothero, 1990), and so they rarely confuse rock units with time units. In addition, recent generations have come to realize that detailed stratigraphic zonations of mammal fossils provide much higher resolution of time than do collections whose only stratigraphic information is the formation from which they came. As discussed by Woodburne (1977), such detailed zonation could potentially subdivide the Cenozoic into increments of time of 300,000 years or less. In recent years, much of the North American continental Tertiary has been formally subdivided by proper biostratigraphic methods. However, the zonation of the entire Cenozoic is still in progress, and biochronological methods are still widely used.

Although mammalian paleontologists are increasingly moving toward classical biostratigraphic procedures, there are still problems. For example, biostratigraphic "zonations" proposed by Gingerich (1980, 1983) and Gunnell (1989) do not meet all the criteria established by the North American Code of Stratigraphic Nomenclature. Typically, these "zones" do not have type sections (required in Article 54e), and in some cases, the actual local stratigraphic ranges of key taxa are not clearly indicated. In this sense, they are "biochrons" based on the abstract first and last occurrences of taxa, not true biostratigraphic zones and stages, which must be based on local ranges of fossils in a particular section.

In addition to higher resolution biostratigraphy, other techniques have come along to improve terrestrial correlations. The original application of K-Ar dating by Evernden, Savage, Curtis, and James (1964) to the North American terrestrial record provided an independent test of the Wood Committee sequence, and showed it to be substantially correct. For over twenty-five years, K-Ar methods continued to refine the chronology and provided numerical estimates of the age for most of the sequence (Savage, 1977; Savage and Russell, 1983; Woodburne, 1987). However, in the last ten years, the development of  $^{40}\text{Ar}/^{39}\text{Ar}$  dating (McDougall and Harrison, 1988) has shown that many of the classical K-Ar dates must be recalibrated. In some cases, they have radically changed the chronology that was accepted for decades (Swisher and Prothero, 1990; Prothero and Swisher, 1992).

The most recent breakthrough has been due to the application of magnetic stratigraphy to terrestrial sections. Unlike any other method, magnetic stratigraphy can supply many globally synchronous, numerically dated time horizons to terrestrial sections (Lindsay et al., 1987; Prothero 1988, 1990; Opdyke, 1990). The combination of both magnetic stratigraphy and  $^{40}\text{Ar}/^{39}\text{Ar}$  dating has provided much higher temporal resolution and precision than was thought possible just twenty years ago. More importantly, magnetic stratigraphy is the only technique that allows direct correlation with the global polarity record, and thus with the marine time scale. This in turn allows us to make direct comparison between global climatic changes, diversity fluctuations, and mass extinctions and the North American terrestrial record for the first time.

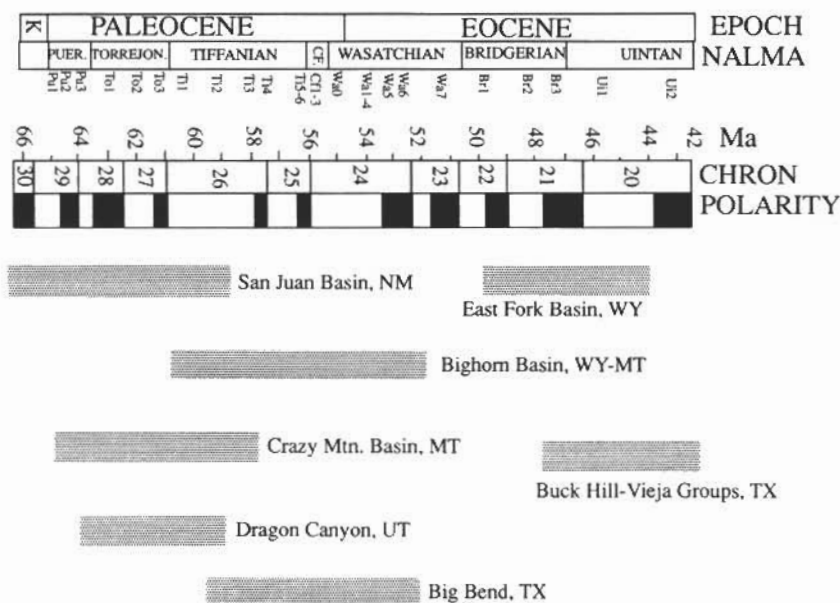
## PALEOCENE

### CHRONOSTRATIGRAPHY AND BIOSTRATIGRAPHY

The term "Paleocene" was originally proposed in 1874 by the paleobotanist W. P. Schimper for strata in Europe with a distinctive assemblage of fossil plants (Schorn, 1971). Carved out of a series of rock units that were part of the original Lyellian Eocene sequence in France, the Paleocene was difficult to recognize because its characteristic fossil plants did not occur widely outside the type area, and many of the rock units in Schimper's original list were clearly Eocene (Berggren, 1971; Savage and Russell, 1983). As a result, early twentieth-century vertebrate paleontologists (e.g., Osborn, 1929) used "lower Eocene" for beds we now consider Paleocene, and older literature must be read with this in mind. The U.S. Geological Survey did not adopt the term until 1939 (Russell, 1964).

Two years later, the Wood Committee (1941) created five land mammal "ages" for an interval they considered Paleocene (Figure 1.1). The first two, the Puercan and Torrejonian, were based on faunas from the San Juan Basin in New Mexico. The controversial Dragonian "age" was based on the limited Dragon Canyon local fauna from the North Horn Formation of central Utah. Although correlative faunas were found in New Mexico (Tomida, 1981), most paleontologists now consider the Dragonian to be the beginning of the Torrejonian (Archibald et al., 1987). The Tiffanian was originally based on faunas from the San Juan Basin in Colorado, but since then the Tiffanian has become much better known from faunas in Montana and Wyoming. The Clarkforkian Land Mammal age was originally based on faunas from the Clark's Fork Basin on the Montana Wyoming border. Originally considered latest Paleocene by the Wood Committee (1941), in recent years it was thought to straddle the Paleocene/Eocene boundary based on correlations of plesiadapids from Europe and North America (Gingerich, 1976; Gingerich and Rose, 1977; Rose, 1981). As we see in the next section, the Paleocene/Eocene boundary is now thought to occur in the earliest Wasatchian, so not only the Clarkforkian but also the earliest Wasatchian are now considered Paleocene.

North American Paleocene chronology was most recently summarized by Archibald et al. (1987). Most of the Paleocene has now



**Figure 1.1.** Paleocene-early Eocene time scale for North America. Global time scale and magnetic polarity time scale after Berggren et al. (1995). North American land-mammal "ages" (NALMA) and "subages" after Gunnell (1989). Temporal span of important magnetostratigraphically calibrated sections are shown; sources are given in text. CF = Clarkforkian; Puer. = Puercan; Torrejon. = Torrejonian.

been subdivided into a series of biostratigraphic "zones," abbreviated "Pu1, Pu2, Pu3, To1" ("Puercan 1, 2, 3, Torrejonian 1") and so on, under a scheme originally introduced by Gingerich (1976, 1980, 1983) based on plesiadapids and primates (Figure 1.1). As already noted, these "zones" do not meet the criteria of the North American Code of Stratigraphic Nomenclature because they lack type sections. Although there are some difficulties with this zonation (Schankler, 1980, 1981), the scheme has been modified with additional mammalian groups and has been widely adopted (Archibald et al., 1987). There are three Puercan "zones," three Torrejonian "zones," six Tiffanian "zones," and three Clarkforkian "zones" (plus Wasatchian 0, which is also latest Paleocene).

Relatively few radiometric dates are available for the Paleocene. The most recent dates on the Cretaceous/Tertiary boundary place its age at 65 Ma (Berggren et al., 1995). Revisions to the dating of the magnetic polarity time scale and new dates on the Eocene (Berggren et al., 1992, 1995; Cande and Kent, 1995) place the Paleocene/Eocene boundary at about 55 Ma. Magnetic polarity stratigraphy (Figure 1.1) has been studied in the key sections of the Bighorn Basin of Wyoming and Montana (Butler, Gingerich, and Lindsay, 1980, 1984), the Crazy Mountains in Montana (Butler, Krause, and Gingerich, 1987), the San Juan Basin in New Mexico (Butler et al., 1977; Lindsay, Jacobs, and Butler, 1978; Taylor and Butler, 1980; Lindsay, Butler, and Johnson, 1981; Butler and Lindsay, 1985), Dragon Canyon in Utah (Tomida and Butler, 1980), and the Big Bend region in Texas (Rapp, MacFadden, and Schiebout, 1983).

Based on magnetic stratigraphy from the San Juan and Crazy Mountain Basins, Butler and Lindsay (1985) and Butler, Krause, and Gingerich (1987) squeezed the Puercan into Chron C29n and latest C29r (64–65 Ma), so Puercan "zones" Pu0–Pu3 are each about 250,000 years in duration. In the San Juan and Crazy Mountain Basins, Torrejonian "zone" To1 occurs in Chron C28n (62.5–63.5 Ma), To2 in C27r (61.3–62.5 Ma), and To3 in Chron C27n (61.0–

61.3 Ma), so the three Torrejonian "zones" To1–To3 range from 0.3–1.2 million years in duration. The Torrejonian/Tiffanian boundary occurs early in Chron C26r, about 61 Ma. Based on magnetic stratigraphy from the northern Bighorn Basin, Butler, Gingerich, and Lindsay (1980, 1984) showed that Ti2 occurs early in Chron C26r (about 59–60 Ma), and Ti3 in late Chron C26r and C26n (57.5–58.5 Ma). Ti4 occurs in earliest Chron C25r (57.0–57.3 Ma), and Ti5 at the base of Chron C25n (56.1–56.3 Ma). The Tiffanian/Clarkforkian boundary occurs in Chron C25n, about 56.0 Ma, so the five Tiffanian "zones" range from 0.2–1.0 million years in duration. The Clarkforkian/Wasatchian boundary occurs in the middle of Chron C24r, about 55.5 Ma, so the three Clarkforkian "zones" are each about 150,000 years in duration.

## PALEOCLIMATE

Paleoclimatic evidence from the Paleocene of North America suggests that the warm, subtropical "greenhouse" conditions that dominated in the Cretaceous persisted into the Paleocene (Wing, this volume, Chapter 2). This is consistent with other data from around the globe that argue for a much warmer, "greenhouse" planet with a reduced pole-to-equator temperature gradient during the Paleocene. However, there was a slight cooling compared to the Late Cretaceous, especially in the Tiffanian and Clarkforkian. Oceanic temperatures were quite high, although estimates vary widely. For example, Shackleton and Kennett (1975) calculated average sea surface temperatures of 18°C in the sub-Antarctic, and bottom water temperatures about 16°C at 1,000 m depth. However, Miller, Fairbanks, and Mountain (1987) obtain estimates closer to 10°C for the deep waters of the Atlantic in the Paleocene. Because both South America and Australia were still connected to Antarctica, there was no circum-Antarctic circulation. Today, this current traps polar waters in a clockwise flow around the Antarctic in the Southern Ocean, allowing cold conditions to build up and accumulate a permanent

Antarctic ice sheet. In the Paleocene and Eocene, polar waters mixed with equatorial waters, ameliorating temperatures and climate and promoting a low temperature gradient between poles and equator. In addition, a warm tropical Tethys seaway still stretched from the Mediterranean to Southeast Asia, contributing to the homogeneity and warmth of the low latitudes.

## PALEOGEOGRAPHY

In many ways, the paleogeographic characteristics of the Paleocene were a continuation of those found in the Cretaceous. The Laramide Orogeny, which began in the latest Maastrichtian, uplifted high basement-cored mountain ranges in the Rocky Mountain region; these same areas had been under marine seaways as recently as the early Maastrichtian (Dickinson et al., 1988). The last of these great epeiric seaways, the Cannonball Sea, covered much of Montana and North Dakota in the early Paleocene; its final retreat marked the end of the repeated transgressions of epicontinental seas that had lasted throughout the Cretaceous and much of the Paleozoic. As the Laramide ranges rose, deep basins formed throughout the Rocky Mountains, which filled up with Paleocene and Eocene fluvial and lacustrine sediments (Dickinson et al., 1988). Some of these basins still contain Paleocene-Eocene deposits as thick as 10,000 m (33,000 feet), producing our primary record of the early Cenozoic in western North America.

Although there were regional differences, most of these Laramide basins were accumulating thick sequences of swampy floodplain deposits during the Paleocene (Dickinson et al., 1988). The Fort Union Group fills many of these intermontane basins in Wyoming, Montana, and North Dakota. In the Powder River Basin of Wyoming and Montana, Fort Union sediments were formed in broad swampy floodplains that deposited coal seams over 100 m (330 feet) thick in places (Flores, 1981, 1983; Flores and Ethridge, 1985). The Bighorn Basin Fort Union sediments were formed in a similar swampy floodplain environment. In the San Juan Basin of New Mexico and Colorado, the Paleocene Nacimiento Formation (from which the "type" Puercan, Torrejonian, and Tiffanian faunas came) was deposited in a swampy floodplain setting, with abundant subtropical plants, fish, crocodiles, and aquatic turtles in addition to the mammals (Smith, Lucas, and Elston, 1985). On the edges of rapidly uplifting Laramide basins were active alluvial fan complexes, which produced thick sequences of Paleocene conglomerates in Utah, Idaho, and Montana (Ryder, Fouch, and Elison, 1976; Dickinson et al., 1988).

Outside the Laramide tract, North America was much less active tectonically. The great Sierra Nevada volcanic arc ceased to erupt in the Maastrichtian, probably as a result of shallow subduction, which may have caused the Laramide Orogeny (Dickinson and Snyder, 1978; Keith, 1978; Dickinson, 1979). However, the Sierran volcanic chain was still a high drainage divide that caused rivers to flow east across Nevada and Idaho into the Laramide lacustrine basins (Nilsen and McKee, 1979). Most of Oregon and Washington were part of a marine basin with an actively subducting trench, and large portions of coastal Oregon and Washington were exotic terranes that had still not accreted to North America

by the Paleocene. In California, the onset of the proto San Andreas fault and its associated crustal stresses produced a variety of early Cenozoic basins, most filled with deep marine turbidites and shales of Paleocene age (Nilsen, 1977, 1987). Major portions of coastal California were still hundreds of kilometers south of their present position and acted as emergent islands, shedding submarine fan complexes in all directions. The Salinian block, which now underlies most of the California Coast Ranges, has apparently traveled about 2,600 km northwesterly since the late Cretaceous (Kanter and Debiche, 1985).

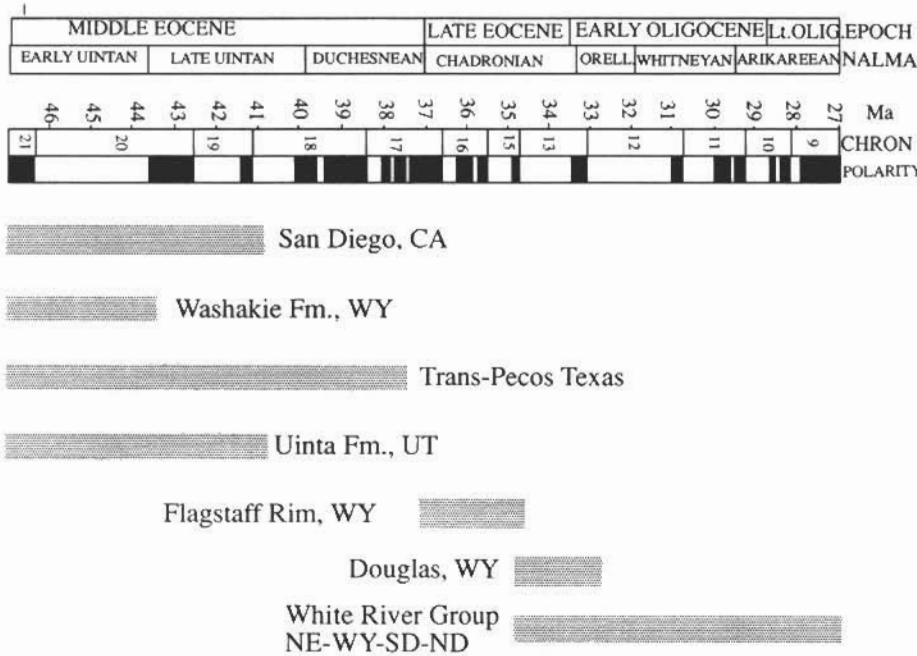
East of the Rockies, most of the Midwest and Appalachian region was emergent and accumulated little or no Paleocene record. Fossil mammals are rare in this region, but do occur (Schoch, 1985). The Atlantic and Gulf Coast passive margins continued to sink and accumulate thick floodplain and marine shelf sequences, as they had been doing since the North Atlantic began to open in the Jurassic. The Atlantic margin subsided relatively slowly, accumulating only about 2 km of Cenozoic sediment (covering over 10 km of Cretaceous deposits). The Gulf margin, by contrast, had a much more abundant supply of clastics from the drainages coming off the Laramide Orogeny, producing over 12 km of Cenozoic sediments alone (Frazier and Schwimmer, 1987; Galloway et al., 1991). Florida and the Bahamian Platform, isolated from clastic sediments, were carbonate banks and reef complexes, as they had been since the Cretaceous.

## EOCENE

### CHRONOSTRATIGRAPHY

The Eocene was one of the three original Cenozoic epochs established by Lyell (1831–1833, p. 55). Lyell's conception of the Eocene was based on a "clock" model of turnover of molluscan faunas and not well suited to traditional stratigraphic procedures (Rudwick, 1978; Berry, 1987). Although Lyell gave some idea of what he considered "typical" deposits of Eocene age, his list led to much confusion. Indeed, some of the units were clearly not Eocene in the modern sense (Berggren, 1971; Savage and Russell, 1983; Berggren, Kent, and Flynn, 1985). Based on studies of marine microfossils in the various type areas of the marine Eocene stages in Europe, a consensus has gradually emerged as to what constitutes "Eocene" and how it correlates with global marine planktonic record and magnetic polarity time scale (Berggren, 1971; Hardenbol and Berggren, 1978; Aubry, 1985; Berggren, Kent, and Flynn, 1985; Aubry et al., 1988). Recently, however, new radiometric dates have forced a significant recalibration of the Eocene time scale (Berggren et al., 1992; Cande and Kent, 1992, 1995; Berggren et al., 1995). Although the relative sequence of biostratigraphic stages and their magnetic polarity correlations are still valid, the numerical age estimates of most Paleogene events are 1–2 million years too old in most publications.

Fortunately, the type areas of many of the European marine stages interfinger with mammal-bearing beds in both the Paris and London Basins (Savage and Russell, 1983). This allows direct correlation of



**Figure 1.2.** Middle Eocene-Oligocene time scale for North America. Global time scale and magnetic polarity time scale after Berggren et al. (1995). Temporal span of important magnetostratigraphically calibrated sections are shown; sources are given in text. Lt. Olig. = Late Oligocene; Orell. = Orellan.

the European Eocene mammalian chronology with the global time scale. During intervals of faunal interchange with North America (such as in the early Eocene, but not the middle or late Eocene), we can correlate the North American sequence with the Lyellian epochs based in Europe. We also have magnetic stratigraphy and a large number of radiometric dates for many areas in the North American Eocene. New radiometric dates, however, have greatly changed our notions of the middle and late Eocene (Swisher and Prothero, 1990; Prothero and Swisher, 1992).

When the Wood Committee subdivided what they considered the Eocene beds of North America, they recognized four land mammal "ages" (Figures 1.1, 1.2). The Wasatchian was named for the Wasatch Formation in basins of southern and eastern Wyoming. The Bridgerian got its name from the faunas of the Bridger Basin in southwest Wyoming, and the Uintan and Duchesnean were named for the Uinta and Duchesne River Formations of northeast Utah. The Wasatchian was considered early Eocene, the Bridgerian middle Eocene, and the Uintan and Duchesnean were thought to be late Eocene. The next "age," the Chadronian (named after the Chadron Formation in the High Plains, especially Nebraska and South Dakota) was considered early Oligocene.

However, a number of recent lines of evidence suggest that the Paleocene/Eocene boundary actually falls within the earliest Wasatchian. The original correlations of Gingerich and Rose were based primarily on the interpretation of adapid primates and pliadapids. Other taxa, such as pantodonts (Lucas, 1984, 1989, 1993; Rea et al., 1990) and omomyid primates (Beard and Tabrum, 1991) suggested that the boundary occurs in the earliest Wasatchian. Another datum is the first appearance of *Platycarya* pollen in the earliest Wasatchian of the northern Bighorn Basin (Wing, Bown, and Obradovich, 1991). This palynological datum occurs at the NP9/NP10 nannofossil zone boundary in the Gulf Coast (Frederiksen 1980), which is 300,000 years older than the Pale-

ocene/Eocene boundary (Berggren, 1993). These correlations were borne out by the detection of the striking carbon isotopic event near the Paleocene/Eocene boundary (Rea et al., 1990; Kennett and Stott, 1991) in earliest Wasatchian terrestrial carbon isotopes extracted from paleosols and mammalian teeth and bones (Koch, Zachos, and Gingerich, 1992).

Yet some mammalian paleontologists (Gingerich, 1989; Gunnell, Bartels, and Gingerich, 1993) continue to place the Paleocene/Eocene boundary at the Clarkforkian/Wasatchian boundary. The key to the entire controversy lies in the interpretation of the Sparnacian mammalian faunas of Europe. In the past, European mammalian paleontologists had considered the Sparnacian the beginning of the Eocene because its mammals were more similar to later Eocene faunas and differed radically from the underlying upper Paleocene Thanetian faunas (Savage and Russell, 1983). The discovery of a new earliest Wasatchian fauna (Wa0) in North America with strong similarities to the Sparnacian in Europe supported that correlation (Gingerich, 1989, pp. 83–87). Gunnell, Bartels, and Gingerich (1993) argued that the beginning of the Clarkforkian/Wasatchian transition was the time of greatest faunal turnover, and therefore it seemed to be the best place to mark the boundary.

However, the Paleocene/Eocene boundary is defined on the basis of marine taxa in European type sections, not on the basis of North American or European mammalian faunas. Where the Clarkforkian/Wasatchian boundary in North America (or the Thanetian/Sparnacian boundary in Europe) falls in relation to marine stratotypes must be demonstrated, not asserted. One cannot assume the coincidence of mammalian faunal turnover and European epoch boundaries. For instance, the "Grande Coupure" in Europe was long thought to represent the Eocene/Oligocene boundary because of the great faunal turnover, but more recent work (Hooker, 1992) has demonstrated that it actually occurred in the early Oligocene, about 2 My after the Eocene/Oligocene boundary.

Marine stratigraphers have long ago shown that the European Sparnacian correlates with late Paleocene nannoplankton zone NP9 (Costa and Downie, 1976; Costa, Denison, and Downie, 1978; Berggren, Kent, and Flynn, 1985; Berggren et al., 1995; Aubry et al., 1988). It is at least one or two sequences lower (and about 1 million years older) than the Paleocene/Eocene boundary as denoted by the base of the London Clay or the Argile d'Ypres (Berggren, 1993), long recognized as the base of the Ypresian and therefore the base of the Eocene (Berggren, Kent, and Flynn, 1985; Aubry et al., 1988).

After the Wasatchian, North America was separated from direct interchange with Europe, and the correlations become more indirect. As a result, the most effective technique has been radiometrically dated magnetic stratigraphy (Figures 1.1, 1.2). The magnetic stratigraphy of the type Bridgerian in the Bridger Basin of Wyoming has been studied, but not published (Jerskey, 1981). The magnetic stratigraphy of the classic sections of the Uintan and Duchesnean in the Uinta Basin of Utah was summarized by Prothero and Swisher (1992) and Prothero (1996a). Bridgerian and Uintan beds have been studied in Wyoming by Flynn (1986) and in Texas by Walton (1992) and Prothero (1996b). Uintan-Duchesnean beds in the Sespe Formation of Ventura County, California (Prothero, Howard, and Dozier, 1996), the Poway Group in San Diego County, California (Flynn, 1986; Walsh, Prothero, and Lundquist, 1996), the Galisteo Formation in central New Mexico (Prothero and Lucas, 1996), and several units in western Montana (Tabrum, Prothero, and Garcia, 1996) have also been sampled. These studies show that the Wasatchian/Bridgerian boundary occurs in Chron C22r, about 50.4 Ma. The Bridgerian-Uintan transition occurs late in Chron C20n, about 46.5 Ma. The Uintan-Duchesnean transition occurs within Chron C18n, about 40 Ma. The latest stratigraphic correlations of the middle Eocene Bartonian and late Eocene Priabonian stages in Europe (Berggren et al., 1995) place the Bartonian/Priabonian boundary in Chron C17n1, so that the middle Eocene includes not only the Bridgerian, but also the Uintan and Duchesnean.

The most surprising conclusion of recent dating, however, concerns the Chadronian Land Mammal age. Considered early Oligocene by the Wood Committee, it had been K-Ar dated by Evernden et al. (1964) at between 32 and 36 Ma. Because the Eocene/Oligocene boundary was generally placed around 36.5 Ma (Berggren, Kent, and Flynn, 1985), the correlation of the Chadronian with the early Oligocene seemed secure. However, the Eocene/Oligocene boundary has been recently redated at 33.5 Ma (Berggren et al., 1995). At the same time, redating of the K-Ar-dated ashes by  $^{40}\text{Ar}/^{39}\text{Ar}$  methods have shown that the Chadronian spans an interval from 34–37 Ma (Swisher and Prothero, 1990; Prothero and Swisher, 1992). This places the Chadronian at the end of the Eocene, and the Eocene/Oligocene boundary appears to fall near the Chadronian/Orellan boundary, not at the Duchesnean/Chadronian boundary, as long thought.

## BIOSTRATIGRAPHY

The most recent summary of most of the Eocene North American land mammal "ages" (Krishtalka et al., 1987) did not attempt to divide formally the entire interval into biostratigraphic

stages (Figures 1.1, 1.2). However, biostratigraphic zonation for the Wasatchian and Bridgerian have been proposed by Savage (1977), Gingerich (1976, 1980, 1983), Gingerich and Simons (1977), Schankler (1980), Stucky (1984), and Gunnell (1989). Some of these biostratigraphic schemes (especially those of Savage, 1977; Schankler, 1980; and Stucky, 1984) follow the 1983 North American Stratigraphic Code in specifying type sections, but others do not. Currently, the Wasatchian is divided into eight "zones," labeled Wa0 to Wa7 (Gingerich, 1983, 1989). Clyde, Stamatakos, and Gingerich (1994) found that Wa0 to lower Wa5 occurred in Chron C24r (53.5–55 Ma) in the northern Bighorn Basin, and that upper Wa5 and Wa6 correlated with C24n (52.3–53.4 Ma). Wa7 occurred in C23r (51.6–52.3 Ma). This is good agreement with a new  $^{40}\text{Ar}/^{39}\text{Ar}$  date on Wa6–7 of 52.8  $\pm$  0.3 Ma.

Gunnell (1989) also named two "zones" (Ui1, the *Epihippus* assemblage "zone," and Ui2, the camelid-canid appearance "zone") for the early and late Uintan. However, unlike earlier Eocene zones, these are not based on recent detailed biostratigraphic work, but simply formalize the distinction between the faunas of Uinta Formation Member "B" (Wagonhound Member of Wood, 1934) and Member "C" (Myton Member of Wood, 1934). In addition, both of Gunnell's (1989) names are inappropriate because camelids appear in the early Uintan in the Washakie Basin (McCarroll, Flynn, and Turnbull, 1996), and *Epihippus* ranges throughout the Uintan. More detailed biostratigraphic work to subdivide the 6 million years of the Uintan in the Uinta Basin is presented in Prothero (1996a). However, the high degree of endemism of Uintan faunas makes true first and last occurrences difficult to distinguish from local geographic effects (Lillegraven, 1979; Flynn, 1986; Krishtalka et al., 1987), so a complete zonation for the Uintan is not yet available.

The Duchesnean has always been the most controversial land mammal "age" of the Eocene. Originally considered latest Eocene by Wood et al. (1941), opinions have fluctuated from placing it in the early Oligocene (Scott, 1945) or back in the late Eocene (Simpson, 1946), and back and forth several times since then. Gazin (1955, 1956, 1959) assigned the faunas from the lower Duchesne River Formation (Randlett and Halfway faunas) to the late Uintan. The poor quality of the overlying Lapoint fauna (the "classic" Duchesnean), and its similarity to the Chadronian fauna, led several workers to either reduce the Duchesnean to a subage of the Chadronian or drop it altogether (Wilson, 1978, 1984, 1986; Emry, 1981). In recent years, opinion has swung back toward recognizing a distinct Duchesnean, with important correlatives in Saskatchewan, South Dakota, Wyoming, Montana, Texas, New Mexico, Oregon, and California (Krishtalka et al., 1987; Kelly, 1990; Lucas, 1992).

Even though the Duchesnean spans nearly 3 million years, efforts to subdivide it into biostratigraphic zones have been controversial because the faunas are so sparse and often endemic to the many scattered localities (Lucas, 1992). Several authors have attempted to recognize an early and late Duchesnean in Texas (Wilson, 1984, 1986) and California (Kelly, 1990). Further radiometric dating and magnetic stratigraphy will probably provide the best test of the age of these faunas and determine whether a zonation can be constructed for the Duchesnean.

At the time of the Wood Committee report, the biostratigraphy of the Chadronian was very poorly understood. Many of the

taxa (especially brontotheres and oreodonts) were badly oversplit, and the stratigraphic data on the fossils from the classic collections in the Chadron Formation were inadequate (Emry, Bjork, and Russell, 1987). Attempts to divide the Chadronian based on lithostratigraphy of the Chadron Formation in Nebraska (Schultz and Stout, 1955) or South Dakota (Clark, 1937, 1954; Clark, Beerbower, and Kietzke, 1967) were unsuccessful (Emry, 1973; Emry, Bjork, and Russell, 1987). Since that time, however, Emry (1973, 1992) has carefully documented the mammalian biostratigraphy in the thickest and most fossiliferous Chadronian sequence at Flagstaff Rim, Wyoming. Based on this work, Emry (1992; Emry, Bjork, and Russell, 1987) suggested criteria for recognizing early, middle, and late Chadronian, but no formal biostratigraphic zonation was proposed. Suggestions of a zonation for the Chadronian are presented by Prothero and Emry (1996b).

## PALEOCLIMATE

From the slight cooling trend of the Paleocene, the globe underwent a dramatic warming at the beginning of the Eocene. Global temperatures may have risen by about 5°C (Miller, Fairbanks, and Mountain, 1987), reaching the warmest state since the mid-Cretaceous. Changes in the carbon isotopes indicate a dramatic alteration of deep-water circulation, and mass extinction of benthic foraminiferans is thought to have been caused by a sudden warming of deep marine waters, so that the water column was almost isothermic from the shallow surface waters to the bottom (Rea et al., 1990; Kennett and Stott, 1991). The causes of this dramatic change are controversial, although an equatorial warm saline deep water (Kennett and Stott, 1990, 1991) or a sudden increase in greenhouse gases, such as CO<sub>2</sub> from increased seafloor spreading (Rea et al., 1990) or methane from Eocene swamps (Cirbus Sloan et al., 1992), have been suggested. This dramatic warming was reflected in early Eocene landscapes and biotas all over the world, including North America (Wing, this volume, Chapter 2).

After the early Eocene peak in warmth, global climate deteriorated through the rest of the Cenozoic. The effects are apparent in North America as early as the middle Eocene. Wolfe (1978) and Wing (this volume, Chapter 2) describe two episodes of cooling in the Eocene of North America, one at about 50 million years ago (Wasatchian-Bridgerian transition) and another at about 38–40 million years ago (approximately the Uintan-Duchesnean transition). As discussed by Wing (this volume, Chapter 2), the floral record showed changes at the beginning and the end of the middle Eocene that seem to correspond to the oceanic cooling shown in the oxygen isotopic record. The early-middle Eocene transition is represented by only slight evidences of cooling and drying. The middle-late Eocene transition, on the other hand, shows significant evidence of cooling of about 7–11°C (11–15°F) in northwestern North America and a loss of tropical elements in both Europe and North America. Clearly, the tropical rain forests of the early Eocene were on the decline around the world.

Yet warm conditions were not gone forever. In some marine isotopic records, there is evidence of a slightly warming trend in the late Eocene, just before the great cooling event in the early Oligocene. According to Wolfe (1978), the late Eocene warming

trend was much more dramatic. In areas like the Big Badlands, ancient soil horizons give a vivid picture of the landscape roamed by the land mammals. According to Greg Retallack (1981, 1983a, 1983b, 1990, 1992), early late Eocene forests received about 1,000 mm of annual precipitation. By the latest Eocene, the forests had changed to a dry woodland with open patches of grasses and herbs. This allowed archaic browsers, like the brontotheres, to persist. The more varied, open habitat gave room for a variety of mixed-feeding herbivores, some of which could eat tougher grasses and herbs.

What caused this dramatic cooling around the globe? Although there are still major arguments about this “doubt house” world between the early Eocene “greenhouse” and the Oligocene “ice house,” evidence is accumulating that there were significant glaciers in Antarctica as early as the middle Eocene, and major pulses in glaciation in the late Eocene (Miller, Wright, and Fairbanks, 1991; Wise et al., 1992). Evidence of middle Eocene mountain glaciers on the Antarctic Peninsula (Birkenmajer, 1987) and in the Pacific sector of the Southern Ocean (Margolis and Kennett, 1971; Wei, 1989) suggests a short-term ice buildup during the middle Eocene. The ultimate cause of the cooling that led to this glaciation is still very controversial. McGowran (1989) attributes some of it to a “reverse greenhouse,” with the excess greenhouse gases in the atmosphere returning to the lithosphere in the form of lignites. Bartek et al. (1992) point out that increased seafloor spreading between Australia and Antarctica produced a broad area of ocean that could have supplied moisture for snow and ice to the Antarctic for the first time since the early Mesozoic. If, as they argue, the Antarctic was already cold enough for ice caps, the triggering factor for middle-late Eocene glaciation may have been the supply of moisture in the proper location.

## PALEOGEOGRAPHY

The Laramide Orogeny, which had changed the North American landscape since the latest Cretaceous, continued into the early Eocene. Many of the Rocky Mountain basins that began to develop and fill in the Paleocene contain even thicker Eocene sequences (Dickinson et al., 1988). Most of these units, such as the Willwood Formation of the Bighorn Basin of Wyoming and Montana, the Wasatch and Wind River Formations in other basins in Wyoming, and the San Jose Formation in the San Juan Basin of New Mexico, were deposited on broad swampy floodplains (Smith, Lucas, and Elston, 1985; Wing and Bown, 1985). In the middle Eocene, thick lacustrine sequences of the Green River Formation accumulated in the Green River and Washakie Basins of Wyoming, the Piceance Basin of Colorado, and the Uinta Basin of Utah. The overlying fluvial Uinta, Duchesne River, and Washakie Formations show that these lakes had receded by the late middle Eocene and late Eocene (Dane, 1955; Bradley, 1964; Ryder, Fouch, and Elison, 1976; Johnson, 1985).

Continued tectonism along the active margin of the Pacific produced a wide variety of Eocene basins in California, Oregon, and Washington (Nilsen, 1987). High sea levels during the Eocene, however, meant that most of these basins were filled with marine sediments, and only a few have nonmarine deposits containing fossil



mammals. Notable among these were the interfingering of marine/nonmarine units in the San Diego region (Lohmar and Warne, 1979; Flynn, 1986; Walsh, Prothero, and Lundquist, 1996) and the Sespe Formation in Ventura County, California (Dibblee, 1966; Taylor, 1983; Kelly, 1990; Prothero, Howard, and Dozier, 1996). Basins in Washington and Oregon were also filled with marine deposits, including major deltaic complexes in central Oregon. During the Eocene, a number of seamounts and other exotic terranes apparently collided with the Pacific Northwest, dramatically increasing the land area (Cady, 1975). Eventually, these seamounts and other exotic fragments clogged up the ancient subduction zone in central Washington and Oregon, and forced it to jump westward to near its present position. At the same time, the entire Coast Range province rotated clockwise by about 50° (Simpson and Cox, 1977; Hammond, 1979). By the later Eocene, the Laramide magmatic null had ended, and arc volcanism returned all over the Cordillera (Lipman, Prostka, and Christiansen, 1972; Snyder, Dickinson, and Silberman, 1976), indicated by the eruption of the Clarno volcanics in central Oregon, the Challis volcanics in Idaho, and the Absaroka volcanics in Wyoming and Montana.

High eustatic sea levels meant that much of the Atlantic and Gulf Coastal Plains were inundated during most of the Eocene (Olsson et al., 1988; Galloway et al., 1991). Although most of the Eocene units found in the Coastal Plain were deposited under shallow to deep marine conditions, a few were nearshore enough to contain fossil mammals washed out to sea (summarized in Westgate, 1986).

## OLIGOCENE

### CHRONOSTRATIGRAPHY

The Oligocene was proposed by H. E. von Beyrich in 1854 for strata in Germany and Belgium that had previously been assigned to the upper Eocene or lower Miocene. Von Beyrich's list of "Oligocene" rocks was very heterogeneous, including units that are now clearly assignable to the Eocene or Miocene, and one unit (the bone sand of Eppelsheim) which produced a late Miocene *Hipparion* fauna. Another problem is that the "type" strata of von Beyrich's Oligocene in Belgium and Germany do not overlie the "type" strata of the Eocene in the Paris Basin or Italy, so the Eocene/Oligocene boundary cannot be recognized in either area. As in the case of the rest of the European Cenozoic, the type sections of the stages within the Oligocene represent only a small portion of the total time (Hardenbol and Berggren, 1978). Although these problems made the Oligocene hard to correlate, it has now been tied to the global marine plankton stratigraphy and magnetic polarity time scale (Berggren, 1971; Berggren, Kent, and Flynn, 1985; Aubry et al., 1988; Berggren et al., 1995).

The biggest dispute has been over recognition of the Eocene/Oligocene boundary and its chronostratigraphy. This argument has received much attention because its date is important to understanding the major extinction events that occurred in the later Eocene (Berggren and Prothero, 1992; Prothero, 1994a). Because the

original "type" sections were inadequate, marine stratigraphers have long sought better sections that preserve the transition without unconformities. Scientists who participated in International Geological Correlation Project 174 met in the 1980s to settle the question, and their results have been published in several places (Pomerol and Premoli-Silva, 1986; Premoli-Silva, Coccioni, and Montanari, 1988). From these conferences, a consensus emerged that the Eocene/Oligocene boundary should be defined at the last appearance of the spiny planktonic foraminiferan *Hantkenina*.

At the 24th International Geological Congress in Washington, D.C., in July 1989, the International Subcommission on Paleogene Stratigraphy met and considered the recommendation of Isabella Premoli-Silva and her colleagues. After some debate, they voted to establish the Massignano section, near Ancona on the Adriatic coast of Italy, as the type section for the Eocene/Oligocene boundary. By vote, the "golden spike" that had been driven at the 19-meter mark on the Massignano quarry face was recognized as "the boundary," and further disputes seemed to be over. Even where planktonic foraminiferans were unavailable, the magnetic stratigraphy at Massignano and Gubbio showed that *Hantkenina* last appeared in the upper third of magnetic Chron C13r. Using this criterion, the Eocene/Oligocene boundary could be recognized in nonmarine rocks as well.

Just as the issue seemed settled, problems emerged. In type section of the uppermost Eocene Priabonian Stage near Priabona, northern Italy, Brinkhuis (1992) reported that the upper part of the type Priabonian is early Oligocene, using the *Hantkenina* datum. The latest research on the changes in oxygen isotopes and Antarctic ice volume increases (Miller, 1992), global oceanic circulation (McGowan et al., 1992), and land plant extinctions (Wolfe, 1992) shows that the most natural event on which to draw the boundary is about a million years later than the *Hantkenina* datum, in what the Paleogene Subcommission designated "early Oligocene" or within Chron C13n (Berggren and Prothero, 1992). Future stratigraphic commissions may decide to change the boundary stratotype, but for now, the boundary remains the *Hantkenina* datum in the upper third of Chron C13r.

Most European mammalian paleontologists placed the Eocene/Oligocene boundary at the dramatic change in mammals known as the "Grande Coupure" (Stehlin, 1909; Brunet, 1977; Savage and Russell, 1983; Brunet et al., 1987). The Grande Coupure marked the extinction of many archaic, endemic Eocene mammalian groups and their replacement by a number of Asian immigrants, especially rhinocerotoids (amynodonts, hyracodonts, and rhinocerotids), anthracotheres, entelodonts, peccaries, several families of advanced rodents (including beavers, pocket gophers, theridomyids and eomyids), hedgehogs, and a variety of carnivorans. It was long assumed that this event correlated with the *Hantkenina* datum (Pomerol and Premoli-Silva, 1986). However, the most recent research (Hooker, 1992) has shown that the Grande Coupure is earliest Oligocene (using the *Hantkenina* criterion), but probably coincides with the global events (circulation changes, Antarctic glaciation, oxygen isotope event, land floral crisis) found to occur in early Oligocene Chron C13n. This evidence might also weigh in favor of changing the Eocene/Oligocene boundary definition.

The geochronology of this boundary has also been very controversial. For decades, the dates produced by K-Ar methods on volcanics and by various isotopes in marine glauconites produced highly contradictory estimates ranging from 32 to 38 Ma (Hardenbol and Berggren, 1978; Odin, 1978, 1982; Curry and Odin, 1982; Glass and Crosbie, 1982; Berggren, Kent, and Flynn, 1985; Odin and Curry, 1985; Berggren, 1986; Glass, Hall, and York, 1986; Aubry et al., 1988; Obradovich, 1988; summarized in Berggren et al., 1992 and Prothero, 1994a). However, recent  $^{40}\text{Ar}/^{39}\text{Ar}$  dating of volcanic ashes in the deep marine sections in Gubbio and Massignano, Italy, directly tied to marine microfossils and magnetic stratigraphy (Montanari et al., 1985, 1988; Montanari, 1988, 1990; Odin et al., 1988, 1991) and of terrestrial volcanic ashes in North America (Swisher and Prothero, 1990; Prothero and Swisher, 1992) have resolved the controversy by eliminating many erroneous age estimates. The emerging consensus places the Eocene/Oligocene boundary at about 33.5 Ma (Berggren et al., 1995).

In addition to redating the Eocene/Oligocene boundary, new  $^{40}\text{Ar}/^{39}\text{Ar}$  methods provided radiometric dates on the rest of the North American Oligocene land mammal sequence for the first time (Figure 1.2). The Wood Committee named three land mammal "ages" that they thought were Oligocene: the Chadronian, the Orellan (based on the Orella Member of the Brule Formation in Nebraska), and the Whitneyan (based on the Whitney Member of the Brule Formation in Nebraska). They considered the Chadronian Land Mammal age to be early Oligocene, the Orellan to be middle Oligocene, and the Whitneyan to be late Oligocene. The Arikareean (based on the Arikaree Group in Nebraska) was thought to be early Miocene. As we have already seen, the Chadronian is now latest Eocene. The Orellan/Whitneyan boundary occurs at about 32 Ma, within Chron C12r, and the Whitneyan/Arikareean boundary within Chron C11n, about 30 Ma (Swisher and Prothero, 1990; Prothero and Swisher, 1992; Tedford et al., 1996). In this calibration, the Orellan, Whitneyan, and earliest Arikareean are early Oligocene, and most of the Arikareean is late Oligocene. (Because the European type Oligocene has only two stages, the Rupelian and Chattian, there is no formal "middle" Oligocene.) Magnetic stratigraphy (Figure 1.2) has now been completed on virtually all the important Orellan, Whitneyan, and early Arikareean outcrops from the White River and Arikaree Groups (Prothero, 1982, 1985a, 1985b, 1996c; Prothero, Denham, and Farmer, 1983; Evanoff, Prothero, and Lander, 1992; Prothero and Swisher, 1992; Tabrum, Prothero, and Garcia, 1996), as well as important localities in Montana (Prothero, 1984; Tabrum, Prothero, and Garcia, 1996) and California (Prothero, 1991; Prothero, Howard, and Dozier, 1996).

The Oligocene/Miocene boundary is less well constrained. When the Wood Committee correlated the Arikareean with the early Miocene, their concept of the Arikareean was based on the Agate Springs fauna, which is latest Arikareean and does have taxa in common with the early Miocene in Europe. However, all of the underlying units in the Arikaree Group have a very different fauna, and the entire Arikareean appears to span the interval from 30 Ma to 21 Ma, almost 9 million years – by far the longest land mammal "age" (Tedford et al., 1987, 1996). In addition, the definition of the Oligocene/Miocene boundary in Europe has fluctuated over

the years, but now seems stabilized around 23.5 Ma (Berggren et al., 1995). Recent  $^{40}\text{Ar}/^{39}\text{Ar}$  dating now places the early Arikareean Gering Fauna between 28 and 30 Ma (Tedford et al., 1996), and most of the Monroe Creek Fauna also appears to be late Oligocene (Tedford et al., 1987, 1996). The Harrison Fauna and the overlying Agate Springs Quarry (in the Upper Harrison Formation of Peterson, 1909) may be earliest Miocene.

## BIOSTRATIGRAPHY

Unlike the detailed biostratigraphic "zonation" now in place for most of the Paleocene and Eocene (Gingerich, 1983; Archibald et al., 1987; Krishtalka et al., 1987; Gunnell, 1989), a detailed Oligocene biostratigraphy is still not fully published. For decades, it was impossible to subdivide the Orellan or Whitneyan land mammal "ages" because most of the early collections had very poor stratigraphic data. The stratigraphic scheme of Schultz and Stout (1955, 1961) was actually based on lithostratigraphic units and did not have biostratigraphic resolution within the units. Outlines of the biostratigraphic potential of the Oligocene were presented by Emry, Bjork, and Russell (1987), but no detailed zonation was proposed. Thanks to decades of work by Morris Skinner, Bob Emry, and collectors of the Frick Laboratory, however, there are now large collections of White River mammals with stratigraphic data zoned to the nearest foot from volcanic ashes. These collections allowed Prothero (1982) to propose a preliminary biostratigraphy that divided the Orellan into four zones and recognized one zone for the early Whitneyan. Korth (1989) also proposed a biostratigraphic zonation for the Orellan, but it is based on University of Nebraska collections that do not have the resolution to subdivide the lithostratigraphic units. Hence Korth's (1989) "faunal zones" are really based on lithostratigraphic boundaries.

A finely resolved biostratigraphy for the Orellan and Whitneyan has long been in preparation, but its publication has been delayed by the lack of systematic revisions of many of the key taxa, especially oreodonts, leptomyricids, and ischyromyid rodents. Such revisions are now published (various papers in Prothero and Emry, 1996a), and Prothero and Emry (1996b) suggested criteria for a zonation of the Chadronian through early Arikareean. Prothero and Whittlesey (1998) proposed a formal biostratigraphic zonation for the Orellan and Whitneyan in the White River Group.

Rensberger (1971, 1973, 1983; Fisher and Rensberger, 1972) proposed a biostratigraphic zonation of rodents from the late Oligocene-early Miocene John Day Formation of central Oregon. The magnetostratigraphy of these beds has also been published (Prothero and Rensberger, 1985), although it will need revision with the new  $^{40}\text{Ar}/^{39}\text{Ar}$  dates recently analyzed by Carl Swisher. Unfortunately, some of the key rodent taxa used in Rensberger's zonation do not occur in the classic areas of the Arikareean in the High Plains, so his biostratigraphic zones have proven useful only in Oregon, Montana, and South Dakota (Tedford et al., 1987). When the magnetic stratigraphy of the type Arikareean in Nebraska is published (Hunt and MacFadden, 1998), it will be possible to overcome these difficulties and establish biostratigraphic zones for the Arikareean throughout the western United States.

## PALEOCLIMATE

Ever since the term "Terminal Eocene Event" (TEE) was proposed (Wolfe, 1978), scientists have focused on events at the end of the Eocene or at the boundary itself (e.g., Pomeroy and Premoli-Silva, 1986). Actually, recent evidence shows that the TEE was insignificant (Berggren and Prothero, 1992; Prothero, 1994a), and the important climatic shift occurred a million years later (about 33 Ma), in what is now defined as early Oligocene (see earlier discussion). In oxygen isotope records, for example, both benthic and planktonic foraminiferans show an increase of about 1.3 per mil in the earliest Oligocene (Miller, Fairbanks, and Mountain, 1987; Miller, 1992; Zachos, Breza, and Wise, 1992). Assuming some Oligocene ice, Miller (1992) calculated that 0.3–0.4 per mil of the change is due to an ice volume increase, which lowered global sea level by 30 m (Haq, Hardenbol, and Vail, 1987). The remaining 0.9–1.0 per mil is explained by about a 5–6°C (9–11°F) cooling, lowering global mean temperature to about 5°C (41°F). It had been as high as 13°C (55°F) in the early Eocene, and 7°C (45°C) in the latest Eocene (Miller, Fairbanks, and Mountain, 1987). This was the most dramatic temperature shift during the entire Cenozoic, until a similar increase in the middle Miocene signaled the onset of the modern Antarctic ice cap.

In 1987, Ocean Drilling Project (ODP) (successor to the Deep Sea Drilling Project) Leg 113 drilled the East Antarctic margin, Maud Rise, and the Weddell Sea in the South Atlantic sector of Antarctica and the Southern Ocean (Kennett and Barker, 1990). These drill cores produced evidence of ice-rafted detritus beginning in the middle Oligocene. Dramatic confirmation of early Oligocene ice in Antarctica came in 1988, when ODP Leg 119 drilled cores on the submerged Kerguelen Plateau and in Prydz Bay, both on the Indian Ocean side of Antarctica (Barron et al., 1989). Thick glacial deposits found in Prydz Bay and ice-rafted sediments on the southern Kerguelen Plateau showed that a major ice sheet was grounded over part of Antarctica by at least the earliest Oligocene. In addition, glacial sediments were found in even older strata, which suggested a middle or late Eocene glaciation near Prydz Bay, although the age of these deposits was not well constrained. The next leg of the Ocean Drilling Project (Leg 120) drilled the central Kerguelen Plateau and provided further evidence of early Oligocene glaciers (Breza et al., 1989; Schlich et al., 1989; Zachos, Breza, and Wise, 1992). The CIROS-1 drill hole in the Ross Sea also produced lowermost Oligocene glacial sediments (Barrett et al., 1989). Although the size and duration of these early Oligocene glaciers is still controversial (Kennett and Barker, 1990), it is clear that there were significant glaciers on some parts of the Antarctic continent about 33 million years ago and short-term glaciation events since the middle Eocene.

One of the most dramatic responses to the early Oligocene cooling was demonstrated by North American land plants (see Wing, this volume, Chapter 2). In the middle Eocene, continental climate exaggerated the global temperature change, producing much more extreme chilling on land. The Oligocene deterioration, however, was the most extreme paleobotanical change in all of the Cenozoic. Even in regions where organic material (including pollen or

megascopic plant fossils) are not well preserved, there is evidence of vegetational change. In the Big Badlands of South Dakota (which produces many of the critical mammal fossils), the only plant fossils are hackberry seeds (which are virtually fossilized with calcite when they are alive) and fossil wood of walnut trees, so the fossil flora is too small for paleoclimatic reconstruction. However, Retallack (1983a, 1983b, 1990, 1992) suggested that paleosols preserved in the late Eocene Chadron Formation were formed under moist closed forest canopies, with between 500 and 900 mm (20–35 inches) of rainfall per year. In the overlying early Oligocene (Orellan) Scenic Member of the Brule Formation, paleosols indicated that rainfall was less than 500 mm (20 inches) per year, and the soil structure was typical of a more open, dry woodland.

In eastern Wyoming, Evanoff (in Evanoff, Prothero, and Lander, 1992) found that moist Chadronian floodplain deposits abruptly shifted to drier, windblown deposits in the early Orellan. These same beds preserve an excellent record of climate-sensitive land snails. According to Evanoff, Chadronian land snails are large-shelled taxa similar to snails now found in subtropical climates with seasonal precipitation, such as in the southern Rocky Mountains and the central Mexican Plateau. Based on modern analogues, these snails indicate a mean annual temperature of 16.5°C (62°F) and a mean annual precipitation of about 450 mm (18 inches) during the late Chadronian in eastern Wyoming. By the Orellan, the large-shelled snails had been replaced by drought-tolerant small-shelled taxa indicative of a warm-temperate open woodland habitat with a pronounced dry season. Such snail faunas are today found in regions like southern California and northern Baja California.

The amphibian and reptile fauna shows a similar trend toward cooling and drying. According to Hutchison (1982, 1992), the aquatic forms (especially salamanders, freshwater turtles, and crocodylians) steadily declined in the late Eocene, and by the Oligocene only terrestrial tortoises were common. This indicates a pronounced drying trend during the late Eocene. Crocodiles were gone by the Chadronian, but more cool-tolerant alligators persisted until the early Orellan.

Although it triggered massive vegetational changes and significant extinctions and migrations in much of the world, the early Oligocene glacial event was not very long lived. According to Miller, Wright, and Fairbanks (1991), the early Oligocene pulse of glaciation may have lasted only a million years, concentrated in western Antarctica and the Indian Ocean sector of the Southern Ocean. It apparently had limited effect in East Antarctica and the South Atlantic sector of the Southern Ocean (Kennett and Barker, 1990). However, there is evidence for a larger and more protracted glacial event or events starting in the late early Oligocene (sometimes called "middle" Oligocene, although the Oligocene has been formally divided only into early and late stages), in magnetic Chrons C11, about 30 million years ago. Benthic foraminiferan  $\delta^{18}\text{O}$  values again exceeded 1.6 per mil, which approximates the threshold of renewed glaciation (Miller, Wright, and Fairbanks, 1991; Miller, 1992). These glacial values of  $\delta^{18}\text{O}$  persisted for about 4 million years, inflicting Antarctica with the longest glaciation (or multiple episodes of glaciation) since the Triassic.

The effects of this great ice sheet are apparent all over the south polar region. The CIROS-1 drill hole in the Ross Sea region cored almost 200 m (650 feet) of mid-Oligocene glacial sediments (Barrett et al., 1989). Mid- or late Oligocene glacial sediments were also reported on Marie Byrd Land (LeMasurier and Rex, 1982), 200 km away in the MSST-1 core in Victoria Land (Barrett et al., 1987), and in King George's Island on the Antarctic Peninsula near South America (Birkenmajer, 1987). Even more striking is the evidence from voyages of the research vessel *R/V Polar Duke* in the Ross Sea region. Seismic reflections off the sediments beneath the Ross ice cap revealed a major mid-Oligocene unconformity, overlain by hundreds of meters of late Oligocene glacial sediments (Bartek and Anderson, 1990; Bartek et al., 1992). Bartek and colleagues were able to trace this unconformity up onto the Antarctic continent, and over at least 100,000 square kilometers beneath the Ross ice shelf. Such a large feature could only have been produced by the grounding of a large ice sheet, comparable to the glacial activity in the Pleistocene.

The most striking global effect of this massive ice sheet was a worldwide drop in sea level. As sea level retreated, the continental shelf was exposed to erosion, forming a deeply incised unconformity. A major mid-Oligocene unconformity is found in many marine sequences around the world, including the Gulf Coast (Keller, 1985a), the Atlantic Coast (Olsson, Miller, and Ungry, 1980; Poag and Schlee, 1984; Poag and Ward, 1987), in Europe (Aubry, 1985), in Australia (Quilty, 1977; Loutit and Kennett, 1981; McGowran, Moss, and Beecroft, 1992), and in southern Africa (Siesser and Dingle, 1981). Deep-sea cores typically have a major gap representing the mid-Oligocene; some were so deeply eroded that the early Oligocene record was wiped out as well (Kennett et al., 1972; Keller et al., 1987). Seismic reflections from beneath the continental shelf off New Jersey and Virginia revealed deeply incised submarine canyons, cut by rivers when the retreating ocean left the shelf exposed in the mid-Oligocene (Miller, Fairbanks, and Mountain, 1987).

Even though this mid-Oligocene glaciation was much bigger than the early Oligocene ice buildup, it had much less of an effect on global climate. Wolfe (1978) indicated no significant mid-Oligocene effect on North American land plants after the earliest Oligocene refrigeration, possibly because late early Oligocene floras were already cold adapted, and tropical taxa had been driven extinct or to lower latitudes. In the Badlands paleosols, Retallack (1983a, 1983b, 1990, 1992) documented a transition from early Oligocene (Orellan) wooded grasslands and gallery woodland to late Oligocene (Whitneyan and early Arikarean) open grasslands with trees only along watercourses. There was also a drying trend, from average annual rainfall of 500–900 mm (20–35 inches) in the early Oligocene to 350–450 mm (14–18 inches) in the late Oligocene. Whitneyan and early Arikarean deposits in the Great Plains were predominantly windblown silts and volcanic dust, indicating very dry conditions.

What caused these episodes of glaciation in the Oligocene? For three decades, oceanic drilling and isotopic studies have suggested that the isolation of the Southern Ocean and the development of circum-Antarctic circulation was the critical factor (Devereux, 1967;

Kennett et al., 1972, 1975; Kennett, 1977). When evidence for the beginning of circum-Antarctic circulation first emerged, attention focused on the timing of the separation of Antarctica and Australia. The two continents began to separate in the late Cretaceous and Paleocene (Weissel, Hayes, and Herron, 1977; Mutter et al., 1985; Veevers, 1986), with rifts opening from west to east (McGowran, 1973; Kennett et al., 1975). By the middle and early late Eocene there was oceanic spreading between the continents. The microfossils suggest that there was still only a shallow marine gulf between the continents, apparently blocked from the South Pacific by the South Tasman Rise (Kennett et al., 1975; Kennett, 1977, 1978, 1980; Murphy and Kennett, 1986; Kamp, Waghorn, and Nelson, 1990). In 1973, Deep Sea Drilling Project Leg 29 drilled in the South Pacific and discovered the first evidence of shallow marine circulation over the Tasman Rise; it apparently occurred in the latest Eocene (Kennett et al., 1975; Shackleton and Kennett, 1975). This may explain why Antarctica began to cool and deep bottom waters were formed shortly thereafter in the earliest Oligocene.

Continued spreading between Antarctica and Australia enlarged the seaway between them. Evidence of the sudden expansion of glacial ice and the widespread "middle" Oligocene unconformity suggests that a new phase of circumpolar circulation had begun at the end of the early Oligocene. Based on the isotopic and micropaleontological evidence, several authors (Kennett et al., 1975; Kennett, 1977, 1978, 1980; Murphy and Kennett, 1986; Kamp, Waghorn, and Nelson, 1990) have attributed this massive cooling event to the beginning of deep-water circulation through the gap between the South Tasman Rise and Antarctica. This is particularly apparent in DSDP Site 277, which lies on the submarine Campbell Plateau (due south of New Zealand); it is right in the path of currents flowing through the gap. In the early Oligocene, Site 277 was still bathed in warm waters flowing down the coast of eastern Australia, but as time passed, the differences in isotopic values between shallow and deep-dwelling foraminiferans increased. Just before the time of the major mid-Oligocene unconformity, there was a sharp increase in the difference in isotopic values between Site 277 and sites northwest of New Zealand (still bathed in warm currents). According to Murphy and Kennett (1986) and Kamp, Waghorn, and Nelson (1990), this is evidence that a blast of deep cold water was passing south of Tasmania, separating the Antarctic circumpolar current from currents flowing further north. Once these currents were decoupled, the beginning of the modern Antarctic polar front was established, separating the circum-Antarctic current from the more northerly currents.

Given the scale of the mid-Oligocene glaciation event and sea-level drop, it seems that circum-Antarctic circulation should have also occurred between Antarctica and South America through the Drake Passage. Studies in this region (Barker and Burrell, 1977, 1982; Sclater et al., 1986) suggest that the Drake Passage did not open until the late Oligocene. However, the timing on this is not as well constrained in terms of biostratigraphy or magnetics. Thus the opening of the Drake Passage might explain either the long duration of the mid-Oligocene cooling event (possibly event Oi2a of Miller, Wright, and Fairbanks, 1991) or the renewed cooling and glaciation at the Oligocene/Miocene boundary (event Mil of

Miller, Wright, and Fairbanks, 1991). If Barker and Burrell (1982) are correct, circulation through the Drake Passage did not really commence until the Oligocene/Miocene boundary.

Although the Southern Ocean seems to have been the major "cold spigot" providing deep, cold bottom waters, it was not the only source. A major early Oligocene unconformity in the North Atlantic produces reflection horizons in the seismic records of the Atlantic margin (Miller and Tucholke, 1983; Mountain and Tucholke, 1985). In addition, studies of benthic foraminiferans showed that in the early Oligocene, the North Atlantic bottom waters were richer in  $^{13}\text{C}$  than those of the Pacific (Miller and Fairbanks, 1983; Miller and Tucholke, 1983; Miller and Thomas, 1985; Miller, 1992). This suggests that the source of some of the deep waters of the Atlantic must have come from the north. The Arctic Ocean had been isolated from the rest of the world's oceans since the Mesozoic, developing its own cold waters with high  $^{13}\text{C}$  ratios. When the Arctic was reconnected with the North Atlantic, such cold waters would be analogous to the modern North Atlantic Deep Water (NADW). The best candidate for this marine passageway was the Norwegian-Greenland Sea, which apparently opened in the early Oligocene (Talwani and Eldholm, 1977; Berggren, 1982). Other possibilities include the Faeroe-Shetland channel, or possibly through the Denmark Straits (Miller and Curry, 1982).

There is no shortage of potential oceanographic sources for the cooling and climatic changes in the Oligocene. The well-established shallow-water circum-Antarctic circulation, plus the cold Arctic source in the North Atlantic, probably triggered the earliest Oligocene cooling event that produced the first significant Antarctic glaciers. The passage of deep water south of Tasmania in the middle Oligocene further accentuated the cold trend, producing a major Antarctic ice cap and the largest sea-level drop in the last 100 million years. The completion of circum-Antarctic circulation through the Drake Passage completed the refrigeration, whether it happened in the late Oligocene or at the Oligocene-Miocene boundary.

## PALEOGEOGRAPHY

The Laramide Orogeny, which had dominated western North America through most of the Paleocene and Eocene, ceased by the late Eocene. In its place was a return of normal subduction and arc volcanism (Snyder, Dickinson, and Silberman, 1976; Dickinson, 1979), especially in the Buck Hill-Vieja volcanics of Trans-Pecos Texas, the Mogollon-Datil volcanic field of New Mexico and Arizona, the San Juan volcanics of southwest Colorado, numerous volcanic fields in Nevada and Utah, and the ancestral Cascades of central Oregon and Washington (Armstrong et al., 1969; Lipman, Prostka, and Christiansen, 1972; Snyder, Dickinson, and Silberman, 1976). Some of these episodes of explosive volcanism were truly immense, dwarfing events such as the 1980 eruption of Mt. St. Helens. For example, the Needles Range volcanics of eastern Nevada and Utah consist of five flows traceable over 35,000 square kilometers, and in southern Nevada, there are individual ash flows containing over 500 cubic miles of volcanic debris (Cook, 1965)!

With the end of Laramide uplift, the Rocky Mountains began to erode away and their surrounding basins began to fill with debris,

especially with silicic volcanic ash from the explosive arc volcanoes to the west. Eventually the Laramide basins had filled to the point where only a low, rolling topography with isolated remnant uplifts was left of the mighty Laramide orogeny (Lillegraven and Ostresh, 1988). This is shown by the development of drainages across the ancient ranges (Clark, 1975; Seeland, 1985), an Eocene-Oligocene surface incised into the tops of many of the Rocky Mountain ranges (Epis and Chapin, 1975; Scott, 1975; Bradley, 1987), and isolated sedimentary remnants of Oligocene and early Miocene age perched on the top of many ranges throughout the Cordillera, such as the Gravelly Range in Montana and the Absarokas and Bighorns in Wyoming (McKenna and Love, 1972; McKenna, 1980). However, Evanoff (1990) has shown that the relief on this surface was not as gentle as once thought. Some paleovalleys filled with White River sediments have Oligocene relief on the order of 500–1,000 m, and many of the supposed "Eocene" subsummit surfaces are actually Miocene in age. Thus the beveling did not develop to its full extent in all parts of the Rockies at the end of the Eocene, but was completely developed by the Miocene.

The burial of these ranges and the filling of their surrounding basins meant that the bulk of sedimentation in the late Eocene and Oligocene shifted to the High Plains further east. The best record of this is the White River Group, famous for the spectacular spires and pinnacles of Badlands National Park in South Dakota (Emry, Bjork, and Russell, 1987). White River sediments were deposited in forested floodplains in the Chadronian and Orellan (Retallack, 1983a, 1983b), with the volcanoclastic component increasing toward the sources in the west (Emry, Bjork, and Russell, 1987). By the Whitneyan and Arikarean, eolian volcanoclastics dominated the section, indicating a semiarid landscape with abundant dunes of volcanic ash and streams with flash flooding (Swinehart et al., 1985; Emry, Bjork, and Russell, 1987).

While the Rocky Mountain region quieted down, the West Coast became more tectonically active in the Oligocene. Sometime around 30 Ma, the spreading ridge and associated transforms between the Farallon and Pacific Plates collided with the subducting margin of North America (Atwater, 1970). When this occurred, the margin switched from conventional subduction of the Farallon Plate to a transform margin. This occurred because the motion of the Pacific Plate relative to the North American Plate was sliding along a northwest-trending transform fault, rather than colliding in a subduction zone. This transform, the San Andreas Fault, dramatically changed the tectonics of the California coast. Exotic terranes were transported north from Mexico and southern California at a rapid rate. Numerous pull-apart basins were developed in the Oligocene and Miocene between sliding fault blocks, and crustal blocks in the Transverse Ranges rotated in a clockwise direction by about 90° (Luyendyk, Kamerling, and Terres, 1980; Luyendyk, 1991). In the Pacific Northwest, conventional subduction continued to generate forearc basins filled with marine sediments.

Although basins were subsiding rapidly all along the Pacific Coast, low eustatic sea levels (especially the 250 m drop in the mid-Oligocene – Haq, Hardenbol, and Vail, 1987) meant that most of the region was exposed to erosion during the Oligocene. In a few places, basins accumulated nonmarine deposits bearing fossil

mammals. For example, in southern California, the late Eocene and Oligocene is represented by sandy floodplain-alluvial fan deposits of the Sespe Formation (Dibblee, 1966; Taylor, 1983; Kelly, 1990). The Sespe produces Uintan-Duchesnean mammals and, after an 8-million-year hiatus possibly related to the mid-Oligocene sea-level drop, early Arikareean assemblages (Prothero, Howard, and Dozier, 1996). Similar Oligocene continental deposits are known elsewhere along the California coast (Bartow, 1978; Nilsen, 1987), although they rarely produce identifiable fossil mammals.

In addition to fault-bounded basins along the coast, the abundance of arc volcanics also preserved a record of the Oligocene landscape in the interior. The widespread, nearly uniform blankets of ash all over eastern Oregon, Nevada, and western Utah suggest there was very little relief between the Rockies and California. In Nevada, volcanic ash frequently filled major lakes that are now preserved by the Elko and Sheep Pass Formations (Nilsen and McKee, 1979). Central Oregon was blanketed by hundreds of meters of volcanoclastics of the late Oligocene-early Miocene John Day Formation (Fisher and Rensberger, 1972). The landscape was apparently forested and semiarid but not as humid or densely forested as the late Eocene environments represented by the Clarno Formation.

As in the earlier Cenozoic, the Midwest and Appalachian region continued to erode during the Oligocene, and we have little or no record for that epoch. However, low eustatic sea levels (Haq, Hardenbol, and Vail, 1987) mean the Atlantic and Gulf Coast were also emergent and eroding during much of the Oligocene as well (Olsson, Miller, and Ungry, 1980; Galloway et al., 1991). In most places, there is a widespread Oligocene unconformity; in others, there are thin nearshore marine units and fluvial-deltaic units, such as the Catahoula Formation of Texas. These deposits show that the Gulf Coast was mostly a flat, swampy floodplain in the Oligocene, with major rivers and deltas building out into the Gulf (Galloway, Hobday, and Magara, 1982; Galloway et al., 1991).

## MIOCENE

### CHRONOSTRATIGRAPHY

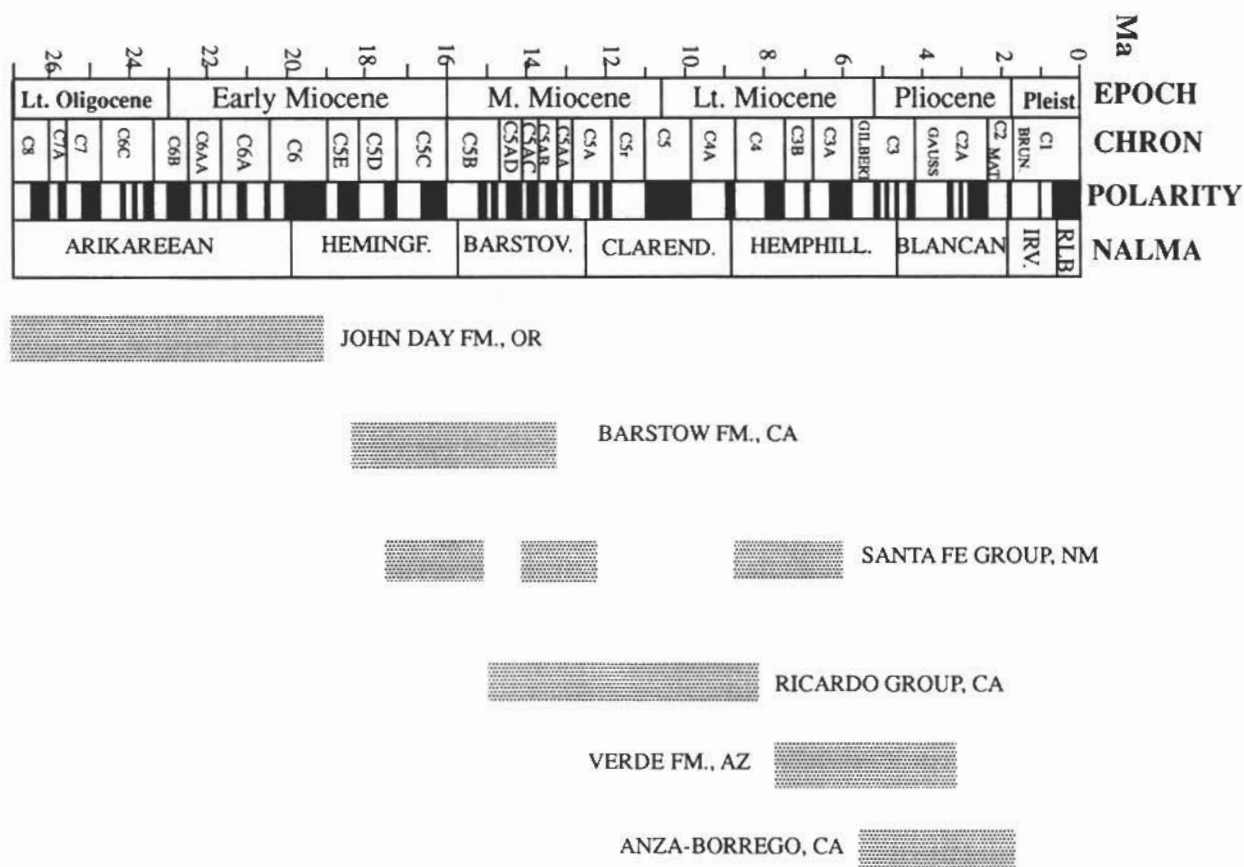
Like the Eocene, the Miocene was another of the three original epochs in Lyell's (1831–1833) subdivision of the Cenozoic. It also had a controversial history because it was based on the "clock" model of molluscan turnover (Rudwick, 1978) and included a confusing combination of "typical" rock units over a wide area of Europe (Berggren, 1971; Berggren and Van Couvering, 1974; Savage and Russell, 1983). The limits of the Miocene have been even more debated (Berggren, 1971; Berggren and Van Couvering, 1974; Berggren, Kent, and Van Couvering, 1985; Berggren et al., 1995), but the present consensus places the base of the Miocene at the base of the Aquitanian Stage in Europe, now dated around 23.5 Ma. Six marine stages—the Aquitanian, Burdigalian, Langhian, Serravallian, Tortonian, and Messinian—are recognized in Europe, and their chronostratigraphy and magnetic stratigraphy have been much more stable than the Paleogene chronostratigraphy reviewed earlier (Berggren, Kent, and Van Couvering, 1985).

Because Miocene marine and nonmarine beds frequently interfinger in Europe, it is possible to establish correlation of European mammal zones with the marine stratotypes (Fahlbusch, 1976; Mein, 1979; Savage and Russell, 1983). However, such a correlation has been more difficult with the North American mammalian record. In parts of the coastal regions of New Jersey, Florida, Texas, and California, marine beds with planktonic foraminiferans, calcareous nannofossils, or diatoms interfinger with mammal-bearing beds, allowing direct correlation with the global time scale (Tedford and Hunter, 1984; Tedford et al., 1987). But most of the classic Miocene faunas come from the western interior, with no possibility of marine interfingering. In these regions, land mammal biostratigraphy is fundamental, greatly aided by radiometric dating and magnetic stratigraphy (Tedford et al., 1987).

When the Wood Committee subdivided their North American "Miocene" in 1941, they recognized only three land mammal "ages"—the Arikareean, Hemingfordian, and Barstovian (thought to be early, middle, and late Miocene, respectively) (Figure 1.3). The Hemingfordian was named after the Hemingford Group in western Nebraska, and the Barstovian received its name from the Barstow Formation in the Mojave Desert of southeastern California. Tedford et al. (1987) reviewed the status of the North American Miocene and pointed out that most of the Arikareean is actually late Oligocene (discussed earlier). The Hemingfordian is early, not middle Miocene, and the Barstovian is middle, not late Miocene. More importantly, revisions of the Miocene/Pliocene boundary in Europe (Berggren and Van Couvering, 1974) produced ages that were much younger than the estimates of the Wood Committee in 1941. This pushed the Clarendonian and Hemphillian Land Mammal ages (both named after towns in the Texas Panhandle region, and considered early and middle Pliocene by the Wood Committee) into the late Miocene.

Tedford et al. (1987) reviewed the large number of K-Ar dates on North American Miocene rocks. Some of these age estimates will have to be revised when more  $^{40}\text{Ar}/^{39}\text{Ar}$  dates become available. Unfortunately, many of these classic Miocene localities have limited stratigraphic continuity or faunal superposition, so the Miocene has not been as amenable to magnetic stratigraphy as have the Paleogene rocks just discussed (Figure 1.3). The late Arikareean has been magnetically sampled (Prothero and Rensberger, 1985; Hunt and MacFadden, 1998), but most of the classic early and middle Hemingfordian localities have not. Tedford et al. (1987) interpolated the age of the Arikareean/Hemingfordian boundary at about 20 Ma. On the other hand, the Hemingfordian/Barstovian boundary has been precisely dated at 15.9 Ma in Chron C5Br in the Barstow area of California (MacFadden et al., 1990; Woodburne, Tedford, and Swisher, 1990). The Barstovian/Clarendonian boundary has been dated at about 12.5 Ma in Chron C5Ar in the Dove Spring Formation (Ricardo Group) in California (Burbank and Whistler, 1987; Loomis and Burbank, 1988; Whistler, 1991; Whistler and Burbank, 1992). Barghoorn (1981) reported the magnetic stratigraphy for Barstovian and Clarendonian faunas in the Tesuque Formation of the Santa Fe Group of New Mexico, which corroborated these correlations.

The Clarendonian/Hemphillian boundary was dated in the Dove Spring Formation (Ricardo Group) of California at 8.9 Ma, early in



**Figure 1.3.** Miocene-Holocene time scale for North America. Global time scale and magnetic polarity time scale after Berggren et al. (1995). Temporal span of some of the important magnetostratigraphically calibrated sections are shown; sources are given in text. Miocene land-mammal "ages" in order as follows: Arikareean, Hemingfordian, Barstovian, Clarendonian, Hemphillian. Pleistocene land-mammal "ages" are the Irvingtonian and Ranchoabrean.

Chron 10 (Whistler and Burbank, 1992). Magnetic stratigraphies have been published for the late Hemphillian Chamita Formation of the Santa Fe Group in New Mexico (MacFadden, 1977) and the Verde Formation of Arizona (Bressler and Butler, 1978). Based on the biostratigraphy, radiometric dating, and magnetic polarity of a large number of late Hemphillian and early Blancan localities all over the western United States, the Hemphillian/Blancan boundary is estimated at about 4.5 Ma, midway through the Gilbert magnetic chron (Lindsay, Johnson, and Opdyke, 1975; Lindsay, Opdyke, and Johnson, 1984). Because the Miocene/Pliocene boundary is placed at the base of the Gilbert chron (about 5.3 Ma), the latest Hemphillian and all of the Blancan are Pliocene (Lindsay, Opdyke, and Johnson, 1984; Tedford et al., 1987).

### BIOSTRATIGRAPHY

Range-zone biostratigraphies have been proposed for the late Arikareean and early Hemingfordian of Oregon (Fisher and Rensberger, 1972; Rensberger, 1971, 1973), the late Hemingfordian and "type" Barstovian in California (Lindsay, 1972; Woodburne, Tedford, and Swisher, 1990), the late Barstovian-Clarendonian-early Hemphillian sequence in the Dove Spring Formation of the Ricardo Group in California (Whistler and Burbank, 1992), and

other Clarendonian localities in California (Savage, 1955). However, many critical Miocene mammal localities in North America must still be correlated by biochronology because they often occur in thin, isolated river channel deposits or quarries with no stratigraphic continuity or superposition (Tedford et al., 1987). Nevertheless, detailed biochronology of the North American Miocene, emphasizing the first and last appearance of key mammalian taxa, yielded a series of biochrons typically 1–2 million years in duration (Tedford et al., 1987). This is comparable to the resolution produced by range-zone biostratigraphy in the earlier Cenozoic (as discussed earlier). For consistency, however, it would be advantageous eventually to establish a true biostratigraphic zonation for the entire Miocene, as has already been achieved for most of the Paleogene.

### PALEOCLIMATE

After the Oligocene refrigeration, oxygen isotopic evidence indicates that climates warmed and recovered slightly in the early Miocene (Woodruff, Savin, and Douglas, 1981; Miller, Fairbanks, and Mountain, 1987; Miller, Wright, and Fairbanks, 1991), with the peak of warming occurring about 17 Ma (Woodruff, Savin, and Douglas, 1981). During most of the early and early middle Miocene (from about 22 to 15 Ma), Antarctica experienced only

short episodes of glaciation (Miller, Wright, and Fairbanks, 1991). Oceanic circulation was relatively sluggish in many parts of the world, with distinct provincialism in foraminiferal faunas and relatively little depth stratification (Keller and Barron, 1983; Keller, 1985b).

In the middle Miocene (about 15 Ma), major ice sheets returned permanently to the Antarctic, and the modern episode of polar glaciation began. Oceanic circulation became much more vigorous, and depth-stratified foraminiferans indicate the presence of shallow, intermediate, and deep waters in the tropics (Kennett et al., 1975; Keller and Barron, 1983; Keller, 1985b). Although the trigger for this pulse of glaciation is uncertain, the connection of the Arctic Ocean to the Atlantic over the sunken Iceland Ridge might have allowed the Arctic to become a heat sink for the first time (Tucholke et al., 1979). Its most striking effect is the increase in North Atlantic Deep Water, which pushed warm saline tropical Atlantic waters south of the Antarctic Convergence and supplied the moisture to build the Antarctic ice sheet (Schnitker, 1980; Miller and Fairbanks, 1985). Also, the closure of equatorial circulation through the Indonesian Archipelago had a profound effect on Pacific and Antarctic circulation, and possibly on global temperature balance at this time as well (Kennett et al., 1975).

Through the middle and late Miocene (15 to 6 Ma), oxygen isotopes indicate many short pulses of cooling and glaciation in the Southern Hemisphere (Woodruff, Savin, and Douglas, 1981). Finally, the Miocene was terminated by a major glacial advance and global cooling event that helped cut off flow into the Mediterranean through Gibraltar. This resulted in the great Messinian salinity crisis, with the entire Mediterranean Basin drying up and refilling several times between 6 and 5 Ma (Adams et al., 1977; Hsü et al., 1977).

These changes were reflected in the climate of North America. Marine Miocene records from the Pacific Coast show a general cooling trend after a peak of warmth in the early middle Miocene, with several cycles of warming and cooling throughout the Miocene (Addicott, 1970; Barron, 1973). In land floras (Wing, this volume, Chapter 2), early Miocene warming was followed by a mid-Miocene (about 13 Ma) cooling trend that corresponds to the growth of the Antarctic ice sheet. Temperatures in North America warmed again about 8 Ma, in the early late Miocene, followed by a final cooling at the end of the Miocene that corresponds to the great Messinian glaciation.

Along with general cooling in the Miocene was a pronounced drying trend, reflected in both flora and fauna. Although woodland savannas persisted well into the Miocene, the traditional interpretation is that the Miocene was characterized by a great increase in grasslands (Webb, 1977; Singh, 1988). The analogy with typical savanna grasslands, however, may not be completely apt because the isotopic signature of abundant C4 grasses (warm-season grasses typical of plains and savannas) in the soil carbonate does not begin until the late Miocene (around 7–8 Ma) in most parts of the world (Quade, Cerling, and Bowman, 1989; Cerling, 1992; Cerling, Wang, and Quade, 1993; Wang, Cerling, and MacFadden, 1994; Quade and Cerling, 1995). Yet a number of middle Miocene ungulates (especially horses – MacFadden, 1992, p. 240; Wang, Cerling, and MacFadden, 1994) had already developed high-crowned teeth,

apparently for grazing. Solounias and Quade (1993) argue that the traditional scenario of early middle Miocene savannas is a myth. Most middle Miocene mammals with hypsodont teeth apparently grazed within a mixed forest-C3 grass habitat, and this is concordant with the palynological evidence of mixed habitats in the Miocene (Leopold and Denton, 1987).

However, evidence from paleosols (Retallack, 1983a, 1990) and pollen (Leopold, Liu, and Clay-Poole, 1992) indicate that some kind of low-biomass vegetation formed open habitats in the late Oligocene and early Miocene, even if C4 grasses were not yet the dominant vegetation. A seasonally dry scrubby community, similar to the saltbush communities of the modern Great Basin, has been suggested (Retallack, 1990). Thomasson (1985) argues that the mid-Miocene silicified grass record is too diverse, and must have been preceded by an Oligocene radiation of unsilicified, unfossilizable grasses. According to this scenario, silicification then evolved in the mid-Miocene in response to heavy grazing by herbivorous mammals. By the late Miocene (7–8 Ma), however, all sources agree that true C4 grasslands were found throughout many of the continents.

Contrary to general misconceptions, North America was not covered by a uniform grassland during the middle and late Miocene. Instead, there was great biogeographic diversity, with habitats ranging from High Plains grasslands to a warm swampy Gulf Coastal Plain to cool temperate mesophytic forests in the Pacific Northwest to fairly arid conditions in parts of California and Nevada (Stenzel, Turner, and Hesse, 1944; Axelrod, 1958; Webb, 1977; Tedford et al., 1987).

## PALEOGEOGRAPHY

Relative to the simple history of North America in the Paleogene, the story from the Miocene onward is very complex (Hamilton and Myers, 1966; Atwater, 1970, 1989; Cole and Armentrout, 1979; Dickinson, 1979; Dickinson and Snyder, 1979; Christiansen and Yeats, 1989; Oldow et al., 1989). Intricate plate tectonic interactions along the Cordilleran margin caused the whole region to undergo major changes in paleogeography, leading to the complicated regional geology still found in the West. Before we review these complexities, however, we summarize the relatively "quiet" areas.

As we have seen throughout the Cenozoic, the Atlantic and Gulf margins continued to sink passively as the Atlantic widened. Unlike the lowlands of the Oligocene, however, the Miocene was characterized by higher sea levels (Haq, Hardenbol, and Vail, 1987), so that the coastal plains were repeatedly inundated, and shallow marine sequences (such as those exposed in the famous Calvert Cliffs of Chesapeake Bay in Maryland) were deposited over a wide area. Due to this complex transgressive-regressive history, fossil mammals are known from a variety of nonmarine and nearshore units in the coastal plains (Tedford and Hunter, 1984), which indicate that the coastal plain was again low and forested throughout the Miocene. This is particularly apparent in the Gulf Coastal Plain in the early and middle Miocene, which hosted a variety of endemic forest dwellers not found on the High Plains (Patton, 1969; Tedford et al., 1987; Prothero and Sereno, 1982).

The Great Plains and Rocky Mountains produce the major part of our North American Miocene record, yet the region was relatively



inactive. The Laramide orogeny, which had shut down in the late Eocene, continued to be eroded away and buried in its own debris. By the late Miocene, most of the Rocky Mountains were buried to the top, producing many of the flattened summit surfaces discussed previously and the erosional remnants left at their crests (Scott, 1975). With the filling of the basins and loss of relief, the sediment supply to the High Plains also began to decrease. Instead of the nearly continuous White River and Arikaree "blanket," which covered most of the northern High Plains during the Oligocene, Miocene sedimentation was much more local and episodic. In Nebraska, for example, much of the Miocene is represented by a series of stacked paleo-valley sequences, indicating decreased sediment supply and more rapid uplift (Stanley, 1976; Skinner, Skinner, and Gooris, 1977; Swinehart et al., 1985). The abundant volcanic debris and eolian deposition that were so important during the Oligocene had virtually vanished.

The most dramatic changes occurred in the Great Basin and Pacific Coast. A variety of events happened almost simultaneously, and most are probably interrelated. One of the most important events was the opening of the Basin and Range province to form a series of north-south-trending horsts and grabens (Nolan, 1943; Stewart, 1971; Proffett, 1977; Eaton, 1980, 1982). On a physiographic map, they form a long series of parallel ranges that C. P. Dutton compared to "an army of caterpillars marching north." Most of the ranges reached 2,000–3,000 m (6,000–9,000 feet) in elevation, and the basins are filled with as much as 3 km (about 2 miles) of sediments. Today, the Basin and Range province extends through Nevada all the way to southern Oregon. The crust of Nevada is now stretched to twice its Miocene width, resulting in thin crust only 20–30 km deep (as opposed to 50–120 km in most regions). This is some of the thinnest crust on any continent, and the nearby mantle supplies heat and magma to the fault zones all through the Basin and Range region (Eaton, 1980, 1982). In the middle Miocene, Basin and Range extension had just begun in the south, and a much narrower Nevada was intruded by arc volcanoes.

Another remarkable phenomenon of the middle Miocene were flood basalt eruptions of eastern Washington and Oregon. Known as the Columbia River basalts, these eruptions spewed out of crustal fissures from deep sources in the mantle (Waters, 1961). Like the Deccan traps and many of the other great flood basalt eruptions, they produced enormous volumes of very fluid basaltic lava in a short period of time. Some flows covered over 40,000 square kilometers in a matter of days, moving at about 5 km/hour in flows 30 m deep and 100 km wide at temperatures of 1100°C! In about 3.5 million years, flow after flow erupted from these fissures until they covered 300,000 square kilometers, filling former valleys with as much as 4,000 m of stacked lava flows. Between eruptions, the flows cooled and then weathered so deeply that forests grew on them. In central Washington, the famous Ginkgo Petrified Forest preserves hundreds of fossilized trees from these ancient forests.

The Columbia River eruptions cannot be explained by simple interactions of the plates, but were probably the result of a giant "hot spot" in the mantle, similar to those that produced Hawaii and the Deccan eruptions. In the late Miocene and Pliocene, similar eruptions occurred farther east in the Snake River plain of southern

Idaho. Today, a hot spot apparently lies under Yellowstone National Park in northwestern Wyoming. If these eruptions are all related, then they probably reflect the eastward apparent migration of the hot spot as the North American plate rode westward over it (Suppe, Powell, and Berry, 1975).

Through the late Miocene, the Basin and Range stretching and the shutoff of the Sierra Nevada volcanic arc continued as the boundary between the subduction zone and the San Andreas transform (known as the Mendocino triple junction) moved north. By the end of the Miocene (5 million years ago), the San Andreas reached north of San Francisco. Basin and Range extension had ripped Nevada apart, breaking it into its modern geometry of north-south-trending fault-block mountain ranges. The stretching of Nevada meant that the Sierras and Cascades had to move west to accommodate the expansion. Indeed, paleomagnetic studies of rocks from both the Sierras and Cascades show that they have moved westward by as much as 270 km (170 miles) since the middle Miocene (Magill and Cox, 1980). They swung like a door, with the hinge located in the Olympic Peninsula of Washington; the Cascades rotated slightly clockwise around it, and the southern Sierras rotated the most. It is analogous to opening a paper fan, with the opening frame represented by the Sierra-Cascade segment and the stretched paper folds symbolizing the stretched crust of the Basin and Range in Nevada and Arizona. The earliest and most rapid stretching occurs near the tip of the fan (Arizona), followed by stretching in the heart of the fan (Nevada), and areas near the hinge (in the Olympic Peninsula) stretched the latest and the least. Recent estimates of extension of about 200 percent in northern Nevada (Proffett, 1977) and up to 500 percent (170–250 km of stretching) across the Las Vegas portion of the Basin and Range are in good agreement with this model (Wernicke, Axen, and Snow, 1988).

By the late Miocene, the great Columbia River basaltic eruptions had long cooled, but the same mantle hot spot was under southern Idaho, erupting the Snake River basalt floodplain. The eastern edge of the Colorado Plateau in New Mexico was marked by another great rift valley that began opening in the middle Oligocene (Chapin, 1979). It is called the Rio Grande Rift because the Rio Grande River now flows through it. Miocene and Pliocene sediments accumulated and then broke into normal fault blocks, very similar to the Great Rift Valley of East Africa or the Triassic rift valleys formed when the Atlantic first opened. However, it is not clear whether the Rio Grande Rift is formed by a continent splitting apart or is part of more complex tectonics.

The northward expansion of the San Andreas transform meant that many pieces of Mexico slid northward into California (Crowell, 1974). The Salinian Block had almost reached the central Coast Ranges, where it now resides. Other blocks moved northward as well. In some places, crustal blocks could not slide smoothly northward along the faults parallel to the San Andreas. Instead, they got caught between fault zones and were sheared into a clockwise rotation. This is typical of the Transverse Ranges north of Los Angeles. As their name implies, the Transverse Ranges lie in an east-west trend that is almost perpendicular to the northwest-southeast trend of the San Andreas and the coastal ranges along it. Paleomagnetic

data shows that the Transverse Ranges have been rotated clockwise about 90° since the middle Miocene (Luyendyk, Kamerling, and Terres, 1980; Luyendyk, 1991). This movement can only be explained if these crustal blocks had pivoted like ball bearings between two sliding surfaces.

As rotation proceeded, deep structural chasms opened up between the pivoting crustal blocks (Crowell, 1974). These became very narrow, steep-walled, rapidly subsiding sedimentary basins. They were so deep that they rapidly filled with thousands of meters of marine shales and turbidites, accumulating thicknesses of as much as 10 km (30,000 feet) of sedimentary fill in the Los Angeles Basin in less than 5 million years! If you stand in downtown Los Angeles, the pre-Miocene bedrock lies more than 6 miles beneath your feet, yet all the sedimentary fill is less than 8 million years old. The Ventura Basin to the north is filled with almost 6 km (over 20,000 feet) of Miocene and Pliocene sediments, and there were many smaller basins with similar histories. The Ridge Basin, to the north of Los Angeles, contains 13.5 km (44,000 feet, or over 8 miles) of fill, yet is only 10–15 km wide! These basins are remarkable, not only because they are so narrow and yet very deep, but also because they were deformed as they grew, so they have steep folds and faults throughout them. Much of the oil wealth of southern California comes from these Mio-Pliocene basins. Oil migrated out of the organic-rich Miocene shales to the faults and folds along the basin edges, where it accumulated and formed great oil fields.

An even more remarkable phenomenon was the “great rip-off” of Baja California. The East Pacific Rise makes a triple junction with the southern San Andreas transform and the Central American trench. During the Miocene, this triple junction gradually moved south along the southern California coast as the San Andreas expanded along the former subduction zone (Atwater, 1970, 1989). About 5 million years ago, the triple junction reached the southern tip of Baja California, and the entire peninsula tore away from mainland Mexico (Crouch, 1979). The East Pacific Rise began to spread open the Gulf of California, placing Baja on the Pacific Plate. Baja California continues to move north today, bringing parts of Mexico with it.

The unusual plate geometry of a subducting margin turning into a transform margin makes Cordilleran geology very complex. For a long time, the Cordillera seemed too complicated for a plate tectonics explanation. In 1970, Atwater used the magnetic record of the Pacific seafloor to predict the behavior of Cordilleran crust. In the more than twenty-five years since then, the plate tectonic hypotheses just outlined have explained many parts of this complicated story.

Still, problems remain. We can see the coincidence in timing between the opening of the Basin and Range Province and the growth of the San Andreas, but what is the connection? A number of hypotheses have been offered. Some have argued that the East Pacific Rise must lie under Nevada, producing the mantle upwelling and crustal extension. However, plate tectonic geometries do not require that this midocean ridge continue indefinitely. It is merely the expression between two plate boundaries, and there need not be any spreading in the mantle where there is no overlying crustal

spreading. Because the Pacific Plate is spreading northwestward away from it, there is no problem with the East Pacific rise ending abruptly in a transform fault. Indeed, that geometry is characteristic of nearly every transform fault along the midocean ridges around the world. Some (e.g., Davis, 1980) have postulated that the Basin and Range is produced by back-arc spreading, as is now happening west of Japan. However, the Basin and Range is no longer behind an arc, but behind a transform margin, which is no longer subducting. Still others have suggested that the Pacific Plate is being subducted and melting beneath Nevada, but this does not produce Basin and Range extension.

Examining the plate geometries of the last 20 million years, it is striking that the beginning of the San Andreas transform means that no plate is going down in that region (Dickinson and Snyder, 1979; Severinghaus and Atwater, 1990). There are remnants of the Farallon Plate sinking underneath the Pacific Northwest and under Mexico, but the Pacific Plate is not subducting beneath the North American Plate. Instead, the two plates scrape sideways against each other, shearing off small fault blocks. Consequently, the North American Plate lies directly on the mantle in this region, whereas in the subduction zones to the north and south, Farallon Plate lies between the deeper mantle and the North American Plate. In other words, the region east of the San Andreas transform is a “slab gap” in the Farallon Plate.

The power of the slab-gap hypothesis is that it predicts so much of Cordilleran geology. Using known Pacific seafloor spreading geometries, and assuming a 30° dip of the Farallon Plate, Dickinson and Snyder (1979) projected the position of the gap in the Farallon Plate through the last 20 million years. As the gap expands, it successfully explains multiple events. For example, the Basin and Range opened progressively from south to north because the expansion of the gap would generate a northward exposure of the mantle. As the northern edge of the gap moves north, the Cascade arc is shut off in the south. The current northern edge of the gap would lie east of the Mendocino triple junction across the northern part of California, just south of the southernmost active Cascade volcanoes, Mt. Lassen and Mt. Shasta. The expansion of the slab gap under the Colorado Plateau occurred between 10 and 5 million years ago, when its uplift began. Geophysical data indicate that there is mantle upwelling underneath the Plateau.

## PLIOCENE

### CHRONOSTRATIGRAPHY

The Pliocene was the third part of Lyell's (1831–1833) original threefold division of the Tertiary. Lyell divided it into the “Older Pliocene,” which became the Pliocene in the modern sense, and the “Newer Pliocene,” which he later renamed the “Pleistocene” in 1839. Lyell based his “Older Pliocene” on the molluscan faunas from the “Subappennine formations” in northern Italy. Like the other epochs of the Tertiary, the stratigraphic arguments over the definition of the Pliocene were long and convoluted (Berggren and Van Couvering, 1974). Eventually, stratigraphers have come to

accept the Pliocene as the epoch following the great terminal Miocene Messinian event in the Mediterranean, whose effects can be recognized around the world. The Pliocene-Pleistocene boundary is currently placed at the top of the Olduvai magnetic subchron (Chron C2n) about 1.8 Ma (Berggren, Kent, and Van Couvering, 1985; Berggren et al., 1995).

The upward revision and redating of the Miocene/Pliocene boundary to the top of the Messinian stage in Europe (Berggren and Van Couvering, 1974) caused the reclassification of the "early Pliocene" Clarendonian and "middle Pliocene" Hemphillian Land Mammal ages of Wood et al. (1941) with the late Miocene (Tedford et al., 1987). Of the Wood Committee's original units, only the latest Hemphillian and the Blancan remain within the Pliocene as presently defined. Most authors (Berggren, Kent, and Van Couvering, 1985; Lundelius et al., 1987) place the Blancan/Irvingtonian boundary at the base of the Olduvai subchron, about 1.88 Ma, slightly older than the European Pliocene-Pleistocene boundary.

Blancan localities have been studied for their magnetic stratigraphy (Figure 1.3) and dated by numerous methods over a wide area of North America (Johnson, Opdyke, and Lindsay, 1975; Lindsay, Johnson, and Opdyke, 1975; Opdyke et al., 1977; MacFadden, Johnson, and Opdyke, 1979; Neville et al., 1979; Lindsay, Opdyke, and Johnson, 1984). According to the most recent summaries (Lindsay, Opdyke, and Johnson, 1984; Lindsay et al., 1987; Lundelius et al., 1987), the Hemphillian/Blancan boundary occurs within the upper Gilbert magnetic chron (approximately at the Cochiti event), about 4.0 Ma. Many events within the Blancan have also been precisely dated, using both magnetic stratigraphy and radiometric techniques.

## BIOSTRATIGRAPHY

Unlike some other intervals of the Cenozoic, the Pliocene-Pleistocene was characterized by rapid faunal turnover and frequent immigrations of mammals into North America from both Eurasia and South America. This has allowed fine-scale subdivision of the late Neogene by numerous faunal events. Lindsay, Opdyke, and Johnson (1984) and Lindsay et al. (1987) recognize at least five separate faunal horizons within the Blancan based on lowest and highest stratigraphic datum levels. These levels have been precisely dated in numerous sections with good stratigraphic and faunal data, especially those in the Anza-Borrego State Park in California (Opdyke et al., 1977) and the San Pedro Valley in Arizona (Johnson, Opdyke, and Lindsay, 1975).

Another scheme for subdividing the late Neogene was proposed by Repenning (1987). He used the biochronology of microtine rodents, a particularly common and rapidly evolving group found in nearly all late Cenozoic mammal localities. Microtines were also frequent migrants to and from the Old World, further enhancing their biochronological potential. Repenning (1987) marked the Hemphillian/Blancan boundary by the immigration of two subgenera of *Mimomys* at 4.8 Ma (older than the estimate discussed by Lundelius et al., 1987). He subdivided the Blancan into five intervals, labeled as Blancan I–V. Blancan I spans the interval from 4.8 to 4.25 Ma; Blancan II from 4.25 to 3.7 Ma; Blancan III is marked

by the immigration of the muskrat *Pliopotamys* at 3.7 Ma; Blancan IV spans an interval from 3.2 to 2.6 Ma; Blancan V, marked by the immigration of two subgenera of the lemming *Synaptomys*, runs from 2.6 to 1.9 Ma. It ends with the Blancan/Irvingtonian boundary, marked by the immigration of *Phenacomys*, *Proneofiber*, *Microtus*, and *Allophaiomys*.

## PALEOCLIMATE

After the great Messinian crisis, the Earth fluctuated in and out of glacial conditions through the Pliocene-Pleistocene. By the mid-Pliocene, closure of the Isthmus of Panama and the onset of Northern Hemisphere glaciation brought global climate into its present-day mode of orbitally controlled Ice Age fluctuations (Shackleton and Opdyke, 1973, 1977; Hays, Imbrie, and Shackleton, 1976; Keigwin, 1978, 1979; Berger, 1982; Berggren, 1982).

The effect on North America was seen in the dramatic change in vegetation, with increased cooling and drying to form a steppe environment (Gregory, 1971; Webb, 1977, 1983, 1984). This climatic change at the end of the Miocene (latest Hemphillian) resulted in dramatic faunal changes, including extinctions among many North American savanna denizens. These included nearly all the rhinocerotids (one tooth scrap is known from the Blancan of Texas), most of the horses, camels, antilocaprids, tapirs, peccaries, mastodons, and all of the protoceratids, moschids, dromomerycids, as well as a variety of rodents, including mylagaulids. In their place came immigrants from Eurasia, including several types of microtine and cricetid rodents, cervids, as well as edentates from South America. Native horses and camels that survived are noticeably more hypsodont in response to even grittier diets.

## PALEOGEOGRAPHY

Most of the events initiated in the Miocene still continue. The Yellowstone hot spot left the Snake River plain and now lies in its present location, generating geysers and hot springs in a huge caldera formed by a gigantic explosive eruption (Suppe, Powell, and Berry, 1975). The Basin and Range continues to rip apart, with the extension just now beginning in southeastern Oregon and Idaho (Dickinson and Snyder, 1979). The Cascades and Sierras have continued to rotate until they reached their present north-south orientation. However, the Cascade arc seems to be less active than it was in the Miocene (Lipman, Prostka, and Christiansen, 1972; Hammond, 1979). Only a few volcanoes, such as Mt. St. Helens, are still erupting in the Washington segment. The northern remnant of the Farallon Plate (known as the Juan de Fuca Plate) is still going down the subduction zone, although it is slowing down and subduction appears to be shallow. This may explain why there are fewer deep-focus earthquakes in the subduction zone beneath the Cascades. The trench off the Oregon-Washington coast is not very deep because it is very rapidly filled with sediments eroded from the humid, recently glaciated lands drained by a large river system.

Pliocene tectonism extended far beyond the Basin and Range and Cascades, however. About 5 million years ago, the "four corners" region in the corners of Arizona, Utah, Colorado, and New Mexico was uplifted as a single coherent block known as the

Colorado Plateau (McKee and McKee, 1972). Unlike the Basin and Range, the Colorado Plateau did not break into hundreds of parallel normal faults. Instead, it is only slightly deformed by small normal faults and gentle folds draped over them; most of the rocks remained nearly horizontal, even though they have been uplifted at least 1.5 km (about a mile). Some parts of the Colorado Plateau, such as the Kaibab Plateau on the north rim of the Grand Canyon, reach elevations over 2,700 m (9,000 feet). Geophysical studies show that the crust beneath the Colorado Plateau is not thick enough to explain its uplift by isostasy alone, so mantle upwelling must be occurring, as it is in the Basin and Range. Today, the Colorado Plateau is deeply eroded due to its gradual uplift, forming the spectacular "red rock" canyons and monuments of many scenic parks: Bryce, Zion, Capitol Reef, Canyonlands, Arches, Monument Valley, and especially the Grand Canyon. In the Grand Canyon and the Goosenecks of the San Juan River, the gently meandering course of the river was deeply incised into Paleozoic sandstone and limestone bedrock, producing meanders that lie in canyons almost a mile deep. This could only happen if the meanders had been established on a low-elevation, low-relief surface during the Miocene and then were forced to cut down through hard bedrock as the Colorado Plateau rose beneath them.

A similar phenomenon occurred in the Rocky Mountains and the Sierras. We have seen that the great Laramide uplifts were almost completely buried during the Miocene, and only small local basins accumulated sediments in Wyoming, Colorado, and Utah. At the end of the Miocene, however, long-buried Rockies were resurrected again after nearly disappearing under their own debris (Scott, 1975).

As their sedimentary cover was stripped away, the flat Eocene Rocky Mountain surface was exposed. The basins between the ancient Laramide uplifts were even easier to erode because they were filled with soft Cenozoic sediments, not hard granites and metamorphic basement. The rivers that once easily crossed the buried mountain ranges were now forced to cut down rapidly as they stripped away the basin fill on either side. Such rapid downcutting produced deep canyons cut through the hard cores of mountains, so that rivers appear to take the "hard" way through a mountain rather than the "easy" way around it. The Rocky Mountains are full of examples of water gaps caused by drainages superimposed on a buried topography. Spectacular canyons such as Royal Gorge in Colorado and Split Mountain and Flaming Gorge in Utah are well-known examples of superposed drainages.

Other regions, such as the Rio Grande Rift (Chapin, 1979) and Alaska, have long and complex Pliocene-Pleistocene histories, but space limitations prevent a full discussion here. Most of the Pliocene paleogeography of North America is a continuation of Miocene trends, and still can be seen today (although overprinted by the effects of Pleistocene glaciation).

## CONCLUSION

Our understanding of the Cenozoic paleogeography, climate, and chronostratigraphy of North America is constantly improving. From the generalized understanding of climatic and geographic patterns of Osborn (1910) and the chronology of Wood et al. (1941), we now can resolve many events to less than a million years and determine

paleotemperatures to a few degrees. More importantly, our improved methods of correlation and dating have allowed us to tie these North American events to the global history of climate and to determine what sort of causal relationships might exist between them. This connection is very important, for an improved understanding of global climatic effects on North American landscape and vegetation is critical to hypotheses of mammalian community evolution and interactions. For example, Osborn (1929) attributed the extinction of brontotheres to non-Darwinian internal forces such as "racial senescence." We now realize that their extinction at the end of the Chadronian coincides with a wide variety of climatic changes induced by Antarctic glaciation and global cooling, which had dramatic effects on the vegetation and habitat on which they depended (Retallack, 1983a, 1983b; Prothero, 1994a, 1994b; Berggren and Prothero, 1992; Evanoff, Prothero, and Lander, 1992).

The importance of this climatic-evolutionary nexus is now becoming widely appreciated as the recent workshops on ancient terrestrial ecosystems attest (e.g., Behrensmeyer et al., 1992). As further details emerge from the deep-sea floor, we are also reaching a much greater level of understanding of global climate than ever before. However, the critical key to all of these hypotheses is chronostratigraphy. Unless the precise temporal correspondence between events can be established, the cause-and-effect connection between climatic change and organismal response is fatally flawed. As the fundamental changes in Eocene-Oligocene chronology have demonstrated, evolutionary-climatic hypotheses built on a weak chronological foundation are as fragile as a house of cards (Prothero and Swisher, 1992).

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