Interplay of oceanographic and paleoclimate events with tectonism during middle to late Miocene sedimentation across the southwestern USA

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ABSTRACT

Continental sedimentation reflects a complex interplay of tectonics and climate. A 2000-km transect from coastal California to the western Great Plains documents a major increase in sedimentation (ca. 16–6 Ma) coeval with deposition of the hemipelagic Monterey Formation along the California coast. Basin and Range-style regional extension following elongation of the Pacific–North American transform boundary at ca. 17.5 Ma provided fault-bounded basins for accommodation space, but sedimentation also occurred on unextended erosional surfaces of the Great Plains and Colorado Plateau. Two global climate transitions bracket this sedimentary interval. The middle Miocene transition (ca. 17–12 Ma) records the global change from equatorial to meridional circulation caused by: (1) closing of the eastern Tethys Seaway (ca. 18 Ma); (2) opening of the Arctic–North Atlantic connection (ca. 17.5 Ma); (3) growth of the East Antarctic Ice Sheet (ca. 14 Ma); and (4) closing of the Indonesian Seaway (ca. 12 Ma). Upwelling of cold waters along the California coast, abetted by domination of La Niña phases of El Niño–Southern Oscillation (ENSO), progressively aridified the Southwest as reflected in sedimentary and biologic records. The second climate transition occurred as opening of the Gulf of California (ca. 6 Ma) intensified the North American monsoon, resulting in integration of drainages, incision of uplifts, and exhumation of basin fills. The Miocene ended with the driest climate of the Tertiary (both regional and global) accompanied by conversion of savanna to steppe or scrub desert, spread of C4 grasses, and the greatest mammal extinction of the Neogene.

INTRODUCTION

The southwestern United States hosts extensive continental sedimentary deposits of middle and late Miocene age between the western Great Plains and the coast of California (Fig. 1). The deposits occur within several physiographic provinces and at variable elevations and tectonic settings. Deposition occurred mainly between 16 and 6 Ma (Fig. 2), coeval with deposition of the hemipelagic Monterey Formation (Ingle, 1981; Behl, 1999) along the California coast (Figs. 1 and 2). In several areas, the deposits accumulated following long intervals of erosion and/or nondeposition (Fig. 2). The middle to late Miocene interval of continental sedimentation ended between ca. 6 and 5 Ma, as integration of drainages brought about widespread incision, exhumation, and deposition of contrasting fluvial deposits of exterior drainages (Eberly and Stanley, 1978; Scarborough, 1989; Spencer et al., 2001a; Connell, 2004; Mack, 2004; Smith, 2004, Polya et al., 2008).

Tectonic events affecting middle and late Miocene sedimentation and erosion across the Southwest were mainly of two types: (1) those that resulted in accommodation space for aggradation of sedimentary deposits, and (2) those that changed oceanic and atmospheric circulation, which affected both climate and sedimentation. A correlation chart that summarizes tectonic events, oceanic and atmospheric changes, and various sedimentation and climatic effects in a temporal framework is presented as Figure 3. The timing of events is based on compilation of published dates of various types from the literature. Ages and descriptions of stratigraphic units are summarized in Appendix 1, with the key references.

Figures 2 and 3 show two periods of exhumation that affected the Southern Rocky Mountains. The first occurred in Oligocene to middle Miocene (ca. 27–16 Ma) and is reflected in the unconformities and missing time-stratigraphic intervals (lacunas) beneath the Ogallala Formation on the east flank of the Southern Rocky Mountains and the Fence Lake and Bidahochi formations on the west flank. Cather et al. (2007) document ~1230 m of exhumation of the southeastern Colorado Plateau between eolian accumulation of the Chuska erg (ca. 33.5–27 Ma) and deposition of the Bidahochi Formation (16–6 Ma). Similar exhumation (~1500 m) and timing (ca. 28–16 Ma) were obtained by Flowers et al. (2008) from (U-Th)/He thermochronometry. This exhumation began during the peak of middle Tertiary ignimbrite volcanism and supports the interpretations of Roy et al. (2004) and Eaton (2008) of magmatically driven middle Tertiary uplift. The second period of exhumation began in late Miocene (ca. 7–6 Ma) and is interpreted herein as resulting from intensification of the North American monsoon and integration of drainage systems that largely ended accumulation of closed-basin continental deposits. Thus, the two periods of exhumation illustrated in Figure 2 document both middle Tertiary tectono-magmatic–driven uplift and late Miocene–Pliocene climatically driven exhumation.

The Miocene record of continental sedimentation and erosion has traditionally been interpreted in terms of tectonic or epeirogenic uplift (for example, Trimble, 1980; Steven et al., 1997; Eaton, 1987, 2008) with little regard to possible climatic effects (Molnar and England, 1990; Molnar, 2004). Since climate is determined mainly by coupling of oceanic and atmospheric circulation, I burrowed into the oceanographic literature seeking tectonic, oceanographic, and paleoclimatic events that had the appropriate timing, scale, and location to have influenced the Miocene sedimentation and erosion history of the southwestern USA. Two climate transitions...
Figure 2. Correlation chart of Miocene continental and coastal sedimentary sequences along a 2000-km transect from coastal California on the west (left) to the western Great Plains on the east (right). Formation names or areas, in the case of multiple formations, are listed across the top. Missing chronostratigraphic intervals (lacunas) are shown in diagonally ruled pattern. Dashed lines above and below labeled stratigraphic units indicate uncertainty in age control. The age ranges of detachment terranes that preceded and partly overlapped with Basin and Range extension are shown for southern Arizona and the Mojave Desert. Ages of formations of wide lateral extent underlying regional unconformities are capitalized and listed below the columns.
**Figure 3.** Chart showing chronology of middle Miocene and younger tectonic events and their oceanographic, climatic, and sedimentologic effects. Antarctic and Arctic glaciations are included because of their significance to the hemispheric thermal gradient and global oceanic and atmospheric circulation. The age ranges for events and effects are based on a compilation of published data cited in the text and Appendix 1. Boundaries of the North American land mammal “ages” (NALMA) are after Tedford et al. (2004) and Woodburne (2004). Abbreviations: AZ—Arizona; NM—New Mexico; ENSO—El Niño–Southern Oscillation; PLST.—Pleistocene; IRV—Irvingtonian; RA—Rancholabrean.
stand out: (1) the middle Miocene global climate transition (ca. 16.8–12 Ma) as defined by Shevenell and Kennett (2004), to which I add effective closing of the Indonesian Seaway to surface and thermocline waters by 12 Ma (van Andel et al., 1975; Leinen, 1979; Keller and Barron, 1983; Kennett et al., 1985; Romine and Lombardi, 1985; Kuhnt et al., 2004; Cloos et al., 2005) with its effect on ocean circulation and initiation of the ENSO climate system; and (2) intensification of the North American monsoon by opening of the Gulf of California at ca. 6.4 Ma (Oskin and Stock, 2003).

MIDDLE TO LATE MIOCENE CONTINENTAL SEDIMENTATION

The age ranges of nine middle to upper Miocene continental sedimentary formations, or groups of geographically related formations, are shown diagrammatically on Figure 2 and summarized in Appendix 1. The timing of the beginning and end of deposition across a west-to-east transect of ~2000 km from California to the western Great Plains is remarkably similar. For several of the sedimentary units, deposition began in middle Miocene after many millions of years of erosion and exhumation that removed older Cenozoic and Mesozoic formations. For example, the Ogallala Formation on the southern Great Plains of Texas and New Mexico overlies a gentle, southeast-sloping surface carved across rocks of Late Cretaceous to Permian age (Hawley, 1993; Gustavson, 1996). Variations of volcanic ash that must have been deposited on the southern plains during the Eocene-Oligocene ignimbrite flare-up (Chapin et al., 2004) of New Mexico, Arizona, and northern Mexico were also eroded from the southern plains. The time interval represented by the basal unconformity of the Ogallala Formation is labeled a lacuna on Figure 2. It represents a specific chronostratigraphic interval between the age of the underlying formations and the ca. 12 Ma age of the basal Ogallala Formation (the lacuna diminishes to the north). Major unconformities are also present beneath the Bidahochi and Fence Lake formations on the west side of the Rio Grande rift–Southern Rocky Mountains (Fig. 1). These lacunas reflect uplift of the Southern Rocky Mountains during the Eocene-Oligocene peak of ignimbrite volcanism, as suggested by Roy et al. (2004) and Cather et al. (2007). Beginning in the middle Miocene, sedimentation began across the eroded flanks of the Southern Rocky Mountains approximately concurrent with deposition of other middle Miocene continental deposits elsewhere in the southwestern USA.

In contrast to the southern Great Plains, middle Miocene continental deposits on the northern Great Plains and in the basins of Wyoming and northern Colorado generally overlie eolian and fluvial deposits of Oligocene-lower Miocene age with variable, relatively minor, unconformities between them (Swinehart et al., 1985; Tedford et al., 1987, 2004; Larson and Evanoff, 1998; MacFadden and Hunt, 1998). The reasons for this contrast with more southerly deposits are the enormous quantities of volcanic ash blown northeastward from volcanic centers of Eocene to early Miocene age in the Great Basin and western Snake River Plains (Larson and Evanoff, 1998) and greater erosion to the south discussed above.

The beginning of regional Basin and Range-style extension in early middle Miocene (ca. 17–15 Ma, Dickinson, 2002; McQuarrie and Wernicke, 2005) provided both fault-bounded basins for depocenters and uplifted basin margins for source areas. Earlier detachment-style deformation generally formed shallower, more complex supradetachment basins in southern Arizona and the Mojave Desert during late Oligocene and early Miocene (see for example, Spencer and Reynolds, 1991). Some overlap in both time and space exists between these two styles of extension (Fig. 2). The major half-graben basins of the Rio Grande rift overlie poorly defined basins in some areas that are generally dominated by volcaniclastic deposits (Chapin and Cather, 1994; Smith, 2004; Connell, 2004; Mack, 2004). These early-rift basins formed (ca. 36–16 Ma) while erosion was stripping the eastern and western flanks of the rift (Fig. 2); thus, they may represent localized stretching during the late Eocene-Oligocene magmatically induced uplift proposed by Roy et al. (2004) and Eaton (2008). An alternative view is presented by Ingersoll (2001).

While subsidence of Basin and Range-style half-graben basins was a major factor in the roughly synchronous onset of continental sedimentation, it was not the only factor. The deposits on unextended terrain of the Great Plains (Fig. 1) and the Colorado Plateau blanket gently sloping surfaces that lack subsidence features except in local areas of evaporite dissolution (Hawley, 1993; Gustavson, 1996). Could a major change in climate be the other factor? Middle to late Miocene aridity in the southwestern USA is evidenced by several sedimentologic and biologic indicators. Caliche horizons are common in sedimentary deposits throughout the region (Appendix 1; Holliday, 1987; Hawley, 1993) and became increasingly abundant in late Miocene. Internally drained basins containing alkaline and saline lakes and playas left evaporite deposits from Wyoming to California (Eberly and Stanley, 1978; Flanagan and Montagne, 1993). Several basins in southern Arizona accumulated thick sequences of anhydrite, halite, and gypsum (Eberly and Stanley, 1978; Scarborough and Peirce, 1978). Eolian sands and loess are common components of the sedimentary record. The western Great Plains and southeastern Colorado Plateau experienced major erosional stripping that cut deeply into underlying formations (Fig. 2) with removal of enormous volumes of sediment from the region. The abrupt change from erosion to aggradation suggests that the changing climate had become so arid by 12 Ma that there was no longer adequate stream flow to carry erosional detritus from the region. Thus, the Ogallala Formation on the western Great Plains is essentially a runoff deposit consisting of superposed paleovalley cut-and-fill deposits (Fig. 4) that, together with eolian contributions, formed a widespread, relatively thin, blanket proximal to the Southern Rocky Mountains.

The Gulf of Mexico is the terminal depocenter for sediment eroded from the Great Plains and east flank of the Southern Rocky Mountains; it should reflect the Neogene tectonic and climatic history. However, multiple fans, growth faults, salt tectonics, sediment bypass, and sea level fluctuations make paleogeographic analyses difficult. Nevertheless, a few observations from the massive compilation of Galloway et al. (2000) are pertinent here. The Oligocene is recorded as an immense, long-lived influx of recycled sedimentary and volcanic rocks, and reworked ash into the Frío and Vicksburg depocenters in the northwest and western Gulf Basin. The early Miocene interval (25–18 Ma) records the maximum extent of the sandy basin floor apron and the first major sediment influx to the central Gulf, perhaps reflecting the pre-Ogallala erosion from the flanks of the Southern Rocky Mountains. The middle Miocene interval (15.6–12 Ma) shows a marked reduction in sediment influx to the western Gulf, which may reflect both a lull in volcanism and sediment trapping in subsiding half-graben basins of the Rio Grande rift and southern Basin and Range. The middle Miocene interval also records the onset of an energetic deep Gulf current by the first deposition of contourite drift deposits in the western Gulf and erosional truncation along the Florida carbonate ramp (Mullins et al., 1987). The late Miocene interval (12–6.4 Ma) records the final decline of the western and northwestern sediment dispersal systems and the shift of fluvial influx to the Mississippi delta system of the central Gulf. The latest Miocene–early Pliocene interval (6.4–4.2 Ma) is dominated by the Mississippi delta system as reflected in a broad bulge of the central Gulf margin. The Rio Grande had not yet reached the Gulf, and much of the drainage from southwestern USA flowed southwestward to the Gulf of California.
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For about ten million years following initial contact between the Pacific and North American plates opposite southern California or northern Baja at ca. 28 Ma, the southwestern margin of North America underwent a very complex structural evolution (Atwater, 1989; Nicholson et al., 1994). The Farallon plate fragmented into several partially subducted microplates, and the Pacific–North American transform boundary grew amidst a welter of rotating blocks, numerous strike-slip faults, slab windows, volcanism, and large-magnitude regional extension (Crouch and Suppe, 1993; Nicholson et al., 1994; Atwater and Stock, 1998; Dickinson, 2002). Beginning ca. 24–22 Ma, detachment faulting and tectonic exhumation affected the inner California borderlands, Los Angeles Basin, Colorado River extensional corridor, southwestern Arizona, and the Mojave Desert, coeval with rotation of the western Transverse Ranges (Crouch and Suppe, 1993; Nicholson et al., 1994; Miller, 2002; Dickinson, 2002). These episodes of large-magnitude, “ductile” extension are represented on Figure 2 as detachment terranes that precede and partially overlap with the widespread middle Miocene Basin and Range extension.

In late early Miocene (ca. 17.5 Ma), the partially subducted Monterey and Arguello microplates were captured by the Pacific plate, greatly elongating the Pacific–North American transform boundary as the Rivera triple junction jumped south past the California borderlands block (Nicholson et al., 1994; Atwater and Stock, 1998; Miller, 2002; Dickinson, 2002). An 800-km segment of the North American plate margin was then free to expand westward behind the northwest-moving Pacific plate (Dickinson, 2002; McQuarrie and Wernicke, 2005). Space for plate-margin expansion led to rapid extension of the central Basin and Range province (Fig. 1) between the Colorado Plateau and the Sierra Nevada block (Dickinson, 2002; McQuarrie and Wernicke, 2005) and accelerated extension in the Rio Grande rift east of the Colorado Plateau (Chapin and Cather, 1994). Widespread half-graben development provided both accommodation space for aggradation of continental sediments and uplifted basin margins for source areas. Numerous continental-margin basins (Figs. 1 and 2) formed along the new Pacific–North American plate boundary, including the basin-and-ridge submarine topography of the southern California borderlands (Ingle, 1981; Dickinson, 2002).

Late early Miocene (ca. 19–17 Ma) also saw the opening and closing of seaways between ocean basins that resulted in major effects on oceanic circulation, global climate, and sedimentary environments. Research by Vogt (1972), Schnitker (1980), and Wright

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**Figure 4.** Diagrammatic north-south cross section of Ogallala paleovalleys between Scotts Bluff, Nebraska, on the North Platte River and Agate on the Niobrara River (modified from Skinner et al., 1977, Fig. 12). The section is based on detailed mapping and paleontological work by the above authors in a 52 km² (20 mi²) area and wider reconnaissance. See Figure 3 for age ranges of North American land mammal “ages” (NALMA). The equivalent numerical ages in Ma were measured on Figure 6.2 of Tedford et al. (2004) and are only approximations. Note that the basal Ogallala Runningwater paleovalley cut almost as deep as the present Niobrara River at Agate. The Snake Creek beds have since been correlated with the widespread Ash Hollow Formation that comprises the upper Ogallala Group in Nebraska (Tedford et al., 1987). More than 7 m.y. of geologic time is represented in ~170 m of net thickness Ogallala Group deposited during several cut-and-fill cycles. Fm.—Formation.
and Miller (1996) indicates that subsidence of the Greenland-Scotland ridge permitted overflow from the Arctic Ocean and Norwegian-Greenland Sea into the North Atlantic, producing a major flux of Northern Component Water (proto–North Atlantic Deep Water) between ca. 20 and 15 Ma (Wright and Miller, 1996). Coring of the Lomonosov Ridge in the central Arctic Ocean by the Integrated Ocean Drilling Program in 2004 (Jakobsson et al., 2007) provided sedimentologic evidence and detailed age control for the initiation of bidirectional flow between the Arctic and North Atlantic oceans through Fram Strait between Greenland and Svalbard. Jakobsson et al. (2007) report that the transition began at 18.2 Ma, was completed by 17.5 Ma, with exchange strengthening by 13.7 Ma when Fram Strait began to open at great depths (present sill depth is >2000 m with a width of 7400 km, Jakobsson et al., 2007). At about the same time (ca. 18 Ma), the eastern Tethys Seaway between Africa and Eurasia was closed as evidenced by exchange of a diverse fauna across a land bridge (Keller and Barron, 1983; Boehme, 2003; Prothero, 2006). Closing of the eastern Tethys Seaway diverted part of the latitudinal, semi-equatorial global ocean circulation, active since the Cretaceous (Haq, 1981), and partially isolated the Mediterranean Sea. Saline outflow waters of the Mediterranean flowed northward in the Atlantic and became an important factor in generation of Northern Component Water (Reid, 1979; Price et al., 1993).

Opening of the Arctic–North Atlantic connection and closing of the eastern Tethys had a dramatic effect on global climate, as reflected in the middle Miocene climatic optimum. This relatively brief interval (ca. 17–14 Ma) in the middle Miocene climatic optimum strongly suggests a major increase in thermohaline circulation. Schnitker (1980) proposed that early North Atlantic Deep Water resulted from Norwegian Sea Overflow Water, which then traveled down the length of the Atlantic Ocean and upwelled as an intermediate water mass into the circum-Antarctic system. The large volume of relatively warm and saline North Atlantic Deep Water injected into the cold circum-Antarctic environment contained heat that could be converted to latent heat by evaporation (Schnitker, 1980). The resultant high evaporation rates supplied moisture to the Antarctic Continent that had been precooled since its isolation from the warmth of low-latitude oceans by formation of the Antarctic Circumpolar Current at ca. 25–23 Ma (Lyle et al., 2007).

Another plate-tectonic event that contributed to Monterey deposition and the middle Miocene climate transition was progressive restriction of the Indonesian Seaway by northward movement of the Australian Continent and growth of the Indonesian archipelago. The best evidence for timing is development of the equatorial undercurrent system (ca. 12 Ma) that returns warm water to the eastern Pacific from the pileup of surface waters against the Indonesian blockage (van Andel et al., 1975; Leinen, 1979; Keller and Barron, 1983; Romine and Lombardi, 1985; Kennett et al., 1985). Also tectonic studies by Cloos et al. (2005) indicate that the Central Range orogeny created the mountainous spine of New Guinea by ca. 12 Ma. Equatorial Pacific waters were diverted into the North and South Pacific gyres, thus strengthening the California and Peru boundary currents that bring cold, nutrient-rich waters toward the equator. The Coriolis effect, which deflects winds and surface waters to the right in the Northern Hemisphere and to the left in the Southern Hemisphere, produced offshore flow via the Ekman effect and intensified upwelling of cold subthermocline waters.
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along the coasts of California and Peru (Ingle, 1981; Soutar et al., 1981).

In addition to strengthening gyral circulation in the Pacific, closing of the Indonesian Seaway provided the basis for the ENSO system. ENSO results from an anomalously warm body of surface water that moves back and forth along the equatorial belt in the Pacific, depending on the strength of the easterly trade winds and modifying effects of ephemeral bursts of westerly winds and eastward-propagating Kelvin waves (see reviews by Cane [1986] and Wang and Picaut [2004]). Restriction of surface and thermocline waters by the rise of the Indonesian archipelago (ca. 12 Ma) caused the buildup of a vast pool of warm surface waters in the western Pacific that previously would have been driven into the Indian Ocean by the easterly trade winds. The resultant Western Pacific Warm Pool (WPWP) is the largest single expanse of anomalously warm open-ocean water on Earth, covering an area larger than the continental United States and containing waters 2 to 5 °C warmer than other equatorial waters (Yan et al., 1992). When the trade winds weaken, waters of the WPWP flow back to the central and eastern Pacific producing the El Niño, or warm phase of ENSO, which recools on a 2- to 7-yr time frame (Wang and Picaut, 2004).

ENSO is an important factor in the regional climate of the Southwest for several reasons. Most importantly, the relatively cold eastern Pacific promotes atmospheric high pressure that deflects the subtropical jet stream toward the Pacific Northwest during the cold (La Niña) and near-neutral phases (~80% of the 1948–1993 period; U.S. National Oceanic and Atmospheric Administration (NOAA) National Weather Service Climate Prediction Center (CPC) web site, 2005), leaving the Southwest relatively dry. In contrast, during El Niño phases, the subtropical jet is steered directly across the southern tier of states, resulting in heavy snow packs in the higher terrain and widespread winter rains, both of which affect the continental sedimentary record. The bimodal southwestern winter, arid most of the time, but with episodic extremes of moisture and runoff, was an important contributing factor in aggradation of middle Miocene continental deposits.

LATE MIOCENE TECTONIC AND OCEANOGRAPHIC EVENTS

A change in direction of the Pacific plate to a more northerly trend between ca. 8 and 6 Ma (Cox and Engelbreton, 1985; Atwater and Stock, 1998) resulted in capture of the Baja Peninsula by the Pacific plate and opening of the Gulf of California by ca. 6.4 Ma (Oskin and Stock, 2003). The resultant marine incursion provided an 1100-km-long, NWW-trending conduit from the tropical Pacific to southwestern North America (Hales, 1972). Such a long fetch over very warm water greatly increased advection of water vapor by southerly winds with resultant intensification of the North American monsoon (Adams and Comrie, 1997; Hunt and Elders, 2001). Today, the North American monsoon provides 25%–50% of the annual precipitation in the region from the western Great Plains to western Arizona, as shown by the contours on Figure 1 (Higgins et al., 1999). There is now general agreement that the bulk of monsoon moisture is advected at low levels from the eastern tropical Pacific and the Gulf of California (Douglas et al., 1993; Higgins et al., 1999; Hunt and Elders, 2001; Mitchell et al., 2002; Bordoni et al., 2004). These studies have shown that monsoon rainfall does not occur prior to the onset of Gulf of California sea-surface temperatures (SSTs) exceeding 26 °C and that the incremental advance of SSTs >26 °C up the mainland coast of Mexico appears necessary for northward advance of the monsoon. Additional moisture is carried overland at higher levels in the troposphere from the Gulf of Mexico (Meehl, 1992) by clockwise circulation around a high-pressure system centered over Texas. Monsoonal flow is frequently interrupted or enhanced during the summer by passage of high- and low-pressure systems across the southwestern United States. Additional late Neogene uplift of the Southern Rocky Mountains and Colorado Plateau (Trimble, 1980; McMillan et al., 2002; Steven et al., 1997; Eaton, 2008) would also have increased monsoonal circulation by providing a more elevated heat source. The consequent increase in the regional land–sea temperature contrast would facilitate convection and drawing in of low-level moisture from the Gulf of California (Meehl, 1992; Bordoni et al., 2004).

LATE MIOCENE EXHUMATION AND DRAINAGE INTEGRATION

Thunderstorms generated by monsoonal moisture (Mann and Meltzer, 2007; Tucker et al., 2006) dramatically increased erosion of the semiarid Southern Rocky Mountains–western Great Plains area in latest Miocene. Incision of strike valleys along the east flank of the Southern Rocky Mountains beheaded the Ogallala depositional systems from sources of sediment in the mountains. The western edge of the remaining Ogallala outcrop belt retreated rapidly 10–140 km to the east (Fig. 1), where it now caps local drainage divides as high as 250 m above the bottoms of strike valleys. In northeastern New Mexico, the upper Canadian River parallels the Rocky Mountains–High Plains boundary in a broad strike valley that contains basalt flows from 7.4 Ma to Quaternary in age (Stroud, 1997; Olmsted and McIntosh, 2004) that directly overlie Cretaceous and older rocks. The Ogallala Formation was largely eroded from the Raton area (Fig. 1) before 7 Ma. In the Ocate volcanic field, 80 km to the south (Fig. 1) basalt flows dated by 40Ar/39Ar at 6.6–6.1 Ma (Olmsted and McIntosh, 2004) decline in elevation from 2800 m to 1950 m over a distance of 60 km on a southeast-facing slope. This relationship was previously interpreted as evidence of uplift of the Rocky Mountains relative to the High Plains (O’Neill and Mehnert, 1988). However, field relationships indicate that the basalts flowed downslope into the strike valley of the Canadian River after erosion removed the Ogallala Formation. Drainage integration in northeastern New Mexico was similar in timing to other areas of the Southwest, as shown on Figure 2; incision in the Raton and Ocate areas of as much as 750–1000 m occurred between 8 and 6 Ma (Chapin, 2002; Olmsted and McIntosh, 2004). Similar strike valleys eroded in Paleogene and Cretaceous formations parallel the east flank of the Southern Rocky Mountains in Colorado and contain the major cities of the Front Range urban corridor.

Aggradation of most middle to upper Miocene sedimentary formations ended in latest Miocene time (Fig. 2), as formerly closed basins were integrated into regional drainages. The oldest known record of the Rio Grande is an axial river gravel in the Santa Domingo basin between Albuquerque and Santa Fe, which contains rheotonic pumice dated by 40Ar/39Ar at 6.9–6.8 Ma (Smith et al., 2001). By ca. 5 Ma, the Rio Grande had integrated several basins to the south by basin filling and overtopping (Comell, 2004) and was emptying into paleolake Cabeza de Baca in the Hueco Basin near El Paso, ~500 km to the south (Mack, 2004). The lower Colorado River is younger than 6 Ma at the mouth of the Grand Canyon, as evidenced by the 5.97 ± 0.07 Ma 40Ar/39Ar age (biotite) of a tuff bed near the top of the Hualapai Limestone (Spencer et al., 2001a). The Colorado River arrived at the head of the early Pliocene Gulf of California by ca. 5 Ma, as evidenced by 40Ar/39Ar ages of 5.51 ± 0.13 Ma (sandine) of a tuff bed underlying the Bouse Formation midway to the Gulf (Houze et al., 2005) and 5.01 ± 0.09 Ma (glass, minimum age) on a tuff bed in the Bouse Formation near the Gulf (Spencer et al., 2001a). The Bouse Formation consists of lacustrine and fluvial sedimentary deposits of the Colorado River (Spencer et al., 2001a; Spencer and Pearthree, 2001). How the Colorado River integrated the 500-km stretch from the Grand Canyon to the Gulf of California has been a subject of controversy. Headward erosion and

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Sedimentation and erosion history can indicate climate change, but proof rests with changes in the biologic response to changing environmental conditions. The middle Miocene climate transition was superposed on a long-term cooling and drying trend that began at ca. 50 Ma, following the early Eocene climatic optimum (Wing, 1998; Zachos et al., 2001). Between ca. 20 and 15 Ma, forested habitats gave way to mixed woodland-grassland mosaics (Fig. 3), and ungulate (hooved) herbivores underwent an impressive diversification, during which many species evolved high-crowned (hypsodont) teeth adapted for grazing and/or gritty vegetation (Webb, 1983; MacFadden, 1992, 2005; Webb and Opdyke, 1995; Janis et al., 2002). In addition to hypsodont teeth, several ungulate lineages (equids, rhinoceroses, oreodonts, and camels) also developed long legs and large size, which traditionally has been interpreted as evidence that savanna or grassland vegetation spread during this time (Webb and Opdyke, 1995; Jacobs et al., 1999; Stromberg, 2006). However, the assumption that tooth crown height in equids was an evolutionary response to open, grass-dominated habitats has been questioned by Strombelius et al. (2007) based on studies of fossilized plantopal (phytoliths) that indicate a 4-m.y. lag between spread of grass-dominated habitats in early Miocene (ca. 22 Ma) and development of full hypsodonty. Research by Fortelius et al. (2006) suggests a strong relationship between local mean hypsodonty and local mean annual precipitation in modern mammal communities. Hypsodonty is not restricted to grass eaters today, and not all modern grazers are hypsodont (Fortelius and Solounias, 2000). Many factors favor hypsodonty, and virtually all increase in effect with increasing aridity and openness of the landscape (Fortelius et al., 2006).

The late Miocene climate change occurred between ca. 7 and 5 Ma, as opening of the Gulf of California intensified the North American monsoon. This latest Miocene climate change was part of a global climate change, often referred to as the terminal Miocene event, or the Messinian, after the European stage during which dramatic changes occurred in the Mediterranean area (see summaries by Cita and McKenzie [1986] and Hodel et al., [1986]). The latest Miocene was the driest part of the Tertiary (Axelrod, 1981; Webb and Opdyke, 1995) and was marked by major changes in both flora and fauna. Grasses underwent a global transition from the C3 to C4 photosynthetic pathway that began earlier in tropical regions (ca. 8–7.5 Ma) and expanded into sub-tropical and temperate regions at higher latitudes between ca. 7 and 4 Ma (Cerling et al., 1997). C4 grasses have a competitive advantage under conditions of aridity, higher temperatures, low atmospheric CO2, and strong seasonality with a summer growing season (Cerling et al., 1997; Tipple and Pagani, 2007). Today, C4 grasses dominate the Great Plains from Mexico to the Canadian border, north of which C3 grasses predominate (Wang et al., 1994). Timing of the C3–C4 transition in southwestern USA has been documented by changes in carbon-isotope ratios measured in fossil horse teeth (Wang et al., 1994; Latorre et al., 1997; Cerling et al., 1997) and in pedogenic carbonate (Quade et al., 1989; Retallack, 1997, 2001).

The latest Miocene and early Pliocene also witnessed the greatest land-mammal extinction of the Neogene, during which 62 genera (35 of which were large mammals) disappeared from North America (Webb, 1984; Webb and Opdyke, 1995). The stable assemblage of the middle Miocene (Clarendonian) chronofauna evolved the greatest diversity of land mammals of the entire Cenozoic and was the peak of equid (horse) diversity, with as many as 13 contemporaneous genera and 15 species (MacFadden, 1992). This explosive radiation of browsing and grazing ungulates occurred as forests were converted to mixed woodlands and grasslands under increasingly cool and dry conditions (MacFadden, 1992). After ten million years of stability, the Clarendonian chronofauna was decimated at the end of the Miocene, as the higher productivity savanna mosaic was replaced by extensive, largely treeless, arid steppes, such as the southern Great Plains (MacFadden, 1992; Webb and Opdyke, 1995), and by scrub deserts to the southwest. The late Miocene aridification is also reflected in an increased terrigenous component in marine hemipelagic sediments beginning ca. 15 Ma and increasing sharply in the past 5 m.y. (Donnelly, 1982; Janecek and Rea, 1983; Hay, 1988; Molnar, 2004). Recent research on the late Miocene–early Pliocene (ca. 8–4 Ma) biogenic bloom (Hermoyian and Owen, 2001; Diester-Haass et al., 2004) in the World Ocean, also known as the late Miocene carbon shift (LMCS) (Tedford and Kelly, 2004), indicates a widespread increase in nutrient flux, perhaps due to the combined effects of aridification and increased monsoonal runoff.

DISCUSSION AND CONCLUSIONS

The middle Miocene global climate transition, defined by Shevenell and Kennett (2004) as the interval ca. 16.8–12 Ma (also reported as 11 Ma) was a multiphase revolution in Cenozoic climate. The transition was from warmer and wetter earlier to cooler and drier later. It started with the unusual warmth of the Miocene Climatic Optimum (MCO, ca. 17–14 Ma), followed immediately (Shevendell and Kennett, 2004) by atmospheric and deep-ocean cooling, culminating in peak growth of the East Antarctic Ice Sheet (EAIS) between ca. 14.0 and 13.8 Ma (Shevenell et al., 2004; Holbourn et al., 2005). The interval 13.8–12 Ma recorded continued growth of ice on Antarctica with deep-water circulation in the Southwest Pacific dominated by Southern Component Water and a strong influx of Pacific Deep Water (Hall et al., 2003; Shevendiell and Kennett, 2004). I suggest that effective closure of the Indonesian Seaway by ca. 12 Ma (Keller and Barron, 1983; Kennett et al., 1985; Cloos et al., 2005) is an important third phase of the middle Miocene climate transition.

Progressive restriction and effective closure of the Indonesian Seaway as Australia moved northward and a complex volcanic archipelago was constructed above subduction zones in the convergence zone with Asia produced major changes in ocean circulation. Most notable were strengthening of Pacific gyral circulation and associated upwelling, generation of the Western Pacific Warm Pool as easterly trade winds piled up warm surface waters against the Australian-Indonesian blockage, initiation of the Equatorial undercurrent system that returns thermocline waters to the eastern Pacific from the buildup in the West, and the El Niño–Southern Oscillation (ENSO) system of coupled oceanic and atmospheric effects as warm waters of the Western Pacific Warm Pool flow back to the central and eastern Pacific where trade winds weaken. Dating more precise than middle Miocene (Kennett et al., 1985) has been difficult, but paleontological and paleomagnetic dating of sharply increased biogenic sedimentation along the equatorial upwelling zone in the eastern Pacific (van Andel et al., 1975; Leinen, 1979; Keller and Barron,
1983), and recent tectonic studies of the Indonesian archipelago (Kuhnt et al., 2004; Cloos et al., 2005) indicate effective closure of the Indonesian Seaway had occurred by ca. 12 Ma. The middle Miocene climate transition (ca. 17–12 Ma) was a pivotal period, both globally and in the geologic evolution of the southwestern United States. Ocean circulation changed from largely equatorial to strongly meridional (Hay, 1988) with closing of the eastern Tethys (ca. 18 Ma, Hsi et al., 1977; Prothero, 2006) and Indonesian (ca. 12 Ma) seaways (Kennett et al., 1985; Romine and Lombardi, 1985; Cloos et al., 2005) and opening of bidirectional flow between the Arctic and North Atlantic oceans (ca. 17.5 Ma, Wright and Miller, 1996; Jakobsson et al., 2007). Strong thermohaline and gyral circulation enhanced by a steeper pole-to-equator thermal gradient, imposed by growth of the East Antarctic Ice Sheet (ca. 14.2–13.8 Ma, Shevenell and Kennett, 2004; Holbourn et al., 2005) brought ocean circulation near to the modern state. Upwelling of cold, nutrient-rich waters along boundary currents in middle latitudes (~20°–40°) produced major hemipelagic deposits of phosphorite and siliceous, organic-rich sediments like the Monterey Formation (Cook and McElhinny, 1979; Summerhayes, 1981). Regional expansion following major elongation of the Pacific—North American transform boundary at ca. 17.5 Ma (Atwater and Stock, 1998; Dickinson, 2002) resulted in formation of fault-bounded basins from the California coast to the Rio Grande rift. The sedimentary response to these tectonic, oceanographic, and climatic changes was deposition of the hemipelagic Monterey Formation along the California coast beginning ca. 18–16 Ma (Fig. 2), and concurrent deposition of a suite of continental sediments in extensional basins and on unextended areas of the western Great Plains and Colorado Plateau. Progressive aridification of the Southwest caused by the combined effects of strong coastal upwelling and La Niña domination of the El Niño–Southern Oscillation system transformed forested habitats into mixed woodland-grassland mosaics, or savannas, within which the ungulate population experienced dramatic growth and diversification (Janis et al., 2002; Prothero, 2006; Stromberg, 2006).

The second major Miocene climate transition to affect the southwestern USA occurred between ca. 7 and 5 Ma, as opening of the Gulf of California (Oskin and Stock, 2003) intensified the North American monsoon. The erosional response to the energized monsoonal climate was to integrate drainages and effectively terminate in many areas the middle to late Miocene deposition of continental sediments in closed basins. The latest Miocene (ca. 7–5 Ma) also saw the spread of C4 grasses, as outlined in the previous section, and the conversion of savanna habitats to largely treeless, arid steppe, such as the southern Great Plains (MacFadden, 1992; Webb and Opdyke, 1995), or to scrub deserts to the southwest. The effects of these degraded environments were recorded in the greatest land mammal extinction of the Neogene in North America (Webb, 1984; Webb and Opdyke, 1995). These dramatic changes in the Southwest occurred during a global climate change often referred to as the terminal Miocene event, or the Messinian, after the European stage during which major changes occurred in southern Europe, including desiccation of the Mediterranean Sea. As summarized by Cita and McKenzie (1986) and Hodell et al. (1986), the increased cooling and drying during the terminal Miocene event was probably caused by increased Antarctic glaciation as reflected in growth of the West Antarctic Ice Sheet, beginning ca. 7 Ma, and a 50- to 70-m drop in sea level with extensive erosion of continental margins. Thus, the regional cooling and drying trend in the southwestern USA reinforced a similar global trend.

Considering the long-running controversy over the relative effects of tectonism and climate (e.g., Epis and Chapin, 1975; Trimble, 1980; Molnar and England, 1990; Mears, 1993; Gregory and Chase, 1994; Steven et al., 1997; McMillan et al., 2002; Eaton, 1987, 2008), how do you sort out which is dominant? For the deposits in extensional basins, tectonism was obviously important, but climate played a role in aridifying the landscape, increasing vulnerability of uplands to erosion, and producing interbedded evaporites. For the coeval deposits on unextended plains, climate was obviously important in changing from erosional stripping to aggradation in paleovalleys lacking sufficient stream flow to flush sediment from the area. But tectonism also played a role through modifying climate by changing ocean circulation and by increasing topographic relief through isostatic uplift of flanks of the Rio Grande rift (Chapin and Cather, 1994). For the latest Miocene integration of drainages, incision of uplifts, and exhumation of basin fills, climate was obviously important, but it was tectonism that opened the Gulf of California and intensified the monsoon. The answer is not one or the other but rather the complex interplay of both tectonics and climate on scales from regional to global.

**APPENDIX 1: FORMATION SUMMARIES FOR FIGURE 2**

**Ogallala Group, Northern High Plains**

**Geographic Extent.** W. Nebraska, S.E. Wyoming, S. South Dakota, E. Colorado, W. Kansas.

**Depocenter Type.** East-trending paleovalleys up to 24 km wide and 100 m deep; complex cut-and-fill structures with reworking of earlier paleovalley fills; deposits merge to form widespread blanket by ca. 12 Ma (Tedford et al., 2004).

**Lithology.** Interbedded sandstones, siltstones, conglomerates, lenses of volcanic ash, minor pond deposits, clasts of Rocky Mountain provenance.

**Thickness.** 30 to 245 m in W. Nebraska, average 180 m.

**Climatic Indicators.** Caliche horizons, volcanic ashfall beds, episodic cutting and filling, numerous internal unconformities, braided streams.

**Supercrop.** Pliocene Broadwater Formation, average age 15 m (maximum 90 m) of fluvial sand and gravel following 1.5 m.y.: hiatus.

**Basal Lacuna.** Minor (0.5 m.y.).

**Age at Base.** Circa 19 Ma (early Hemingfordian, Tedford et al., 2004); 16 ± 3.7 Ma fission-track zircon age on volcanic ash 50 m above base (Zett, 1975).

**Age at Top.** Circa 5 Ma (late Hemphillian, Tedford et al., 2004); 4.6 ± 1.0 Ma fission-track zircon age on ash in uppermost Ogallala (Izett, 1975).

**Key References.** Skinner et al. (1977); Diffendal (1982); Swinehart et al. (1985); Swinehart and Diffendal (1990).

**Ogallala Formation, Southern High Plains**

**Geographic Extent.** S.E. Colorado, S. Kansas, W. Oklahoma, W. Texas, E. New Mexico.

**Depocenter Type.** SE-trending paleovalleys that headed on the Southern Rocky Mountains and were incised 60–150 m into the pre-Ogallala erosion surface; blanket deposit formed as valleys were filled and interfluves covered.

**Lithology.** Gravelly and sandy braided-stream deposits interbedded with eolian sands and loess-like deposits; extensive caprock of calcrete deposits.

**Thickness.** 30–210 m on S. High Plains but locally 560–580 m in solution-subidence depressions above Permian evaporite deposits.

**Climatic Indicators.** Eolian sands and loess, calcic paleosols, local playa facies, calcrete caprock; fossil flora and fauna indicate savanna-like environment.

**Supercrop.** Pliocene Blanco Formation (3.5–2 Ma); eolian Plio-Pleistocene Gatauna and Blackwater Draw formations.

**Key References.** Progressively older formations from north to south; Cretaceous to Permian rocks beneath major erosion surface.

**Pre-Ogallala Neogene Erosion.** Post–28 Ma Capanian granite pluton (Allen and McLemore, 1991), pre–12 Ma basal Ogallala (Kelley and Chapin, 1995).

**Age at Base.** 12 Ma based on Clarendonian vertebrate fossils (Gustavson, 1996; Tedford et al., 2004); 13.0 ± 0.6 Ma based on K-Ar date (glass) on volcanic ash near Orla, Texas, initially mapped as Gatauna Formation (Powers and Holt, 1993).

**Age at Top.** 6–7 Ma based on late Miocene jaw of fossil proboscidean (Morgan and Lucas, 2001) and 6–7 Ma 40Ar/39Ar ages of basal flows in Raton and Ocate areas of New Mexico (Stroud, 1997; Olimsted and McIntosh, 2004); 6.6 ± 0.8 Ma fission-track zircon age on ash overlying Hemphillian vertebrate fossils at Coffee Ranch, Texas (Izett, 1975); 4.5 Ma NALMA (North American land mammal “ages”) (Tedford et al., 2004).

**Key References.** Holiday (1987); Winkler (1987); Gustavson and Winkler (1988); Hawley (1993); Gustavson (1996).
Northern and Central Rio Grande Rift

Geographic Extent. Central Colorado to central New Mexico; San Luis, Santo Domingo, Española, Albuquerque, Socorro basins.

Depocenter Type. Linked, oppositely tilted, half grabens.

Lithology. Diverse suite of siliciclastic deposits ranging from conglomerates to mudstones but typically poorly indurated, buff-colored, sandstones and siltstones; also ash beds, mafic flows, dune fields, playas, deposits, and caliche horizons.

Thickness. As much as 6 km of sedimentary fill on deep side of larger half grabens.

Climatic Indicators. Closed basins, dune fields, gypsum playa deposits, caliche horizons; savannas-like fossil vertebrate assemblage similar to Clarendonian assemblage of the Great Plains and transitional to that of the Great Basin (Tedford, 1981).

Stratigraphic Summary. Santa Fe Group—sedimentary and volcanic fill of Rio Grande rift; lower Santa Fe Group—deposits of internally drained basins (ca. 30 to 7–5 Ma); upper Santa Fe Group—axial river and piedmont deposits following integration of drainage ages at 7–5 Ma; Pleistocene entrenchment (0.7 Ma) of the Rio Grande Valley—end of Santa Fe Group.

Supercrop. Latest Miocene–Pliocene fluvial deposits of ancestral Rio Grande and bordering piedmont fan complexes; basalt flows.

Subcrop. Oligocene–early Miocene volcanic and volcanioclastic units transition from Oligocene volcanic fields to middle Miocene well-defined rift basins; lower Potosopa Formation (ca. 26–16 Ma) in structurally fragmented basins of Socorro area; unit of Isleta #2 (ca. 36–16 Ma) in Albuquerque basin; volcanioclastic aprons of San Juan and Latir volcanic fields and Orlitz porphyry belt in Española Basin, including Los Pinos Formation on west side and Espinosa Formation on east side; Conejos Formation in Monte Vista graben of San Luis Basin; Oligocene nonvolcanic sedimentary units in the Española Basin, including the lower Namibe member of the Tesuque Formation and the lower Abiquiu and Picuris formations (Smith, 2004).

Basal Lacuna. Variable, local, inadequate age control.

Age at Base. Sedimentation began ca. 26 Ma, but rapid subsidence and accumulation of thick fills began about 18 to 16 Ma (Cather et al., 1994; Connell, 2004; Smith, 2004).

Age at Top. The transition from closed basins to through-going epicontinental drainage began northwest of Santa Fe, New Mexico, at about 7 Ma as recorded by river axial deposits underlying volcanic rocks dated at 6.93 ± 0.05 Ma (40Ar/39Ar sanidine, McIntosh and Quade, 1995); late Hemphillian vertebrate fossils in the San Juan and Rac quarries in the Chimita Formation were dated at 6.75 ± 0.05 Ma (40Ar/39Ar, sanidine). The ancestral Rio Grande emplaced into the fluvial-deltaic complex of the Potosopa Formation north of Socorro until between 7 and 5 Ma (Connell, 2004), after which by progressive basin spillover, it integrated the drainage southward to the El Paso area by 5 Ma (Mack, 2004).

Key References. Chapin and Cather (1994); Cather et al. (1994); Smith et al. (2001); Connell (2004); Koning et al. (2004); Smith (2004).

Fen Lake Formation

Geographic Extent. West-central New Mexico, northwest corner of Mogollon-Datil volcanic field and bordering Colorado Plateau.

Depocenter Type. NW-trending paleovalleys and alluvial fans.

Lithology. Fluvial sandstones and conglomerates, clasts of mafic to rhyolitic rocks of Mogollon-Datil volcanic field, boulders to 1 m, interbedded caliche.

Thickness. 20–100 m.

Climatic Indicators. Caliche horizons; 2-m-thick Stage VI petrocalcic soil caps Fence Lake Formation and underlies 6.7 Ma base surge deposits (Love et al., 1994).

Supercrop. Basalt flows dated by 40Ar/39Ar at 6.8 to 6.0 Ma with local basalt insets possibly continuing until ca. 5 Ma (McIntosh and Cather, 1994); Plio-Pleistocene (ca. 4 to <1 Ma) Quemado Formation (McIntosh and Cather, 1994).

Subcrop. Inset against Bearwall Mountain Andesite (ca. 24–26 Ma) and older units; unconformably overlies Eocene Baca and Eager formations and upper Cretaceous Moreno Hill Fill.


Age at Base. About 14.5 Ma or less, based on fossil mammal bone of a proboscidean whose oldest record in North America is ca. 14.5 Ma (Lucas and Anderson, 1994); fossil bone located ~0.3 m above basal unconformity of Fence Lake Formation (Love et al., 1994).

Age at Top. Four 40Ar/39Ar ages between 6.80 and 6.0 Ma on overlying basalt flows; one 5.20 Ma 40Ar/39Ar age on possibly interbedded basalt (McIntosh and Cather, 1994).

Key References. McLelllan et al. (1982); Lucas and Anderson (1994); McIntosh and Cather (1994).

Bighiichi Formation

Geographic Extent. Approximately 16,000 km² of southern Colorado Plateau in northeastern Arizona and western New Mexico; original extent ~50,000 km².

Depocenter Type. Broad, shallow basin incorporating paleovalleys and upland alluvial plains.

Lithology. In north, mostly fine-grained lacustrine(? member); middle volcanic member includes Hopi Buttes volcanic field; thin beds of distal silicic volcanic ash; lacustrine claystone, siltstone, and marl. In south, increasing fluvial deposits.

Thickness. Ranges up to 240 m thick.

Climatic Indicators.olian sand sheets, calcic soils, lacustrine facies with fresh-water fossils, selcetnites.

Supercrop. Plio-Pleistocene basalt flows, Quaternary alluvium.

Subcrop. Cretaceous to Permian formations beneath an erosion surface with up to 60 m of relief.

Basal Lacuna. Late Eocene-middle Oligocene eolian beds of Chuska Sandstone (Cather et al., 2007) completely removed after eruption of 25 Ma mafic flow and prior to onset of Bighiichi deposition at ca. 16 Ma.

Age at Base. About 16 Ma based on 40Ar/39Ar age of 15.84 ± 0.05 Ma (geochronal correlation with ash bed in Buffalo Canyon), and weighted average age of 15.46 ± 0.36 Ma from two 4Ar/39Ar (bioite) dates on east point ash bed; ca. 16 Ma onset of sedimentation based on above 4Ar/39Ar ages and strata accumulation rates (Dallegrave et al., 2003).

Age at Top. About 6 Ma based on correlation of ash bed at top of member 5 with 6.02 ± 0.03 40Ar/39Ar age of Blacktail Creek ash bed and late Hemphillian (ca. 6–5 Ma, NALMA) age of White Cone fauna (Baskin, 1979). See Dallegge et al. (2003) for additional 4Ar/39Ar ages and correlations with dated ash beds in University of Utah database.

Key References. Love (1989); Dallegrave et al. (2003).

Fill of Extensional Basins, Southern Arizona

Geographic Extent. Southern Basin and Range Transition Zone. A depth-to-bedrock map of the Arizona Basin and Range province (Oppenheimer and Sumner, 1981) outlines ~64 basins that contain, or may contain, middle and upper Miocene sedimentary fill; however, most are undissected and overlain by alluvial fans and pediment gravel. Figure 1 shows portions of 18 basins where outcrops, drill holes, and geophysical data provide a basic framework of Miocene sedimentation.

Depocenter Type. Grabsen, half grabens, structural troughs bounded by relatively wide-spaced, steep normal faults that contrast with the thin-skinned structural style of the late Oligocene–early Miocene Miocene detachment terranes with their relatively shallow supradetachment basins.

Lithology. Locally derived clastic sediments interbedded with basalt flows, red-brown lacustrine clay, thin nonmarine evaporates of halite, anhydrite, and gypsum.

Thickness. Drill holes and seismic data indicate 200–3000 m of massive evaporites in some basins and comparable thicknesses of clastic basin fill.

Climatic Indicators. Thick, nonmarine evaporites, red-brown oxidized lacustrine clays, petrocalcic soils.

Supercrop. Deposits of external drainage systems, such as gravels of the Gila and Colorado rivers (ca. 5–6 Ma) and the Piocene Bouse Formation (ca. 5.4 Ma, Spencer et al., 2001a); pediment alluvium fans and pediment gravels; basalt flows.

Subcrop. Late Oligocene to middle Miocene tilted fault blocks of volcanic rocks (ca. 23–16 Ma), alluvial fan and playa deposits that postdate detachment.
faulting and predate Basin and Range block faulting (Spencer et al., 2001b).

**Basal Lacuna.** Variable, regional unconformity developed between ca. 17 and 10 Ma; inadequate age control to determine missing interval.

**Age at Base.** Overlies 16.4 Ma andesitic tuff in Exno No. 1 Yuma Federal drill hole (Yuma basin) and 14.9 Ma maﬁ c ﬂ ow in Exno No. 1 State 74 drill hole (Picacho basin) (Eberly and Stanly, 1978); basal Hickey Formation (16.2 Ma, Leighty, 1998); 15 Ma depositional overlap across deformed middle Miocene units (Menges and Pearthree, 1989); middle Miocene transition from felsic volcanism and core–complex extension to high-angle normal faulting and deep basin formation (Spencer et al., 2001b).

**Age at Top.** Deep paleochannels cut perpendicularly to Mogollon Rim in Verde Valley prior to 5.66 Ma reﬂ ecting lowered base level by opening of Gulf of California prior to 5.5 Ma (Nations et al., 1985); river gravels below 6.0 Ma basal near Gillespie Dam indicate external drainage of Gila River began in late Miocene (Nations and Stanly, 1978); basin-ﬁ ll sediments and basalts of Perkinsville Formation (6.3–4.6 Ma, Leighty, 1998) deposited in basins of Transition Zone; upper basin-ﬁ ll units depositionally overlap basin-bounding faults and bury bedrock pediments fringing adjacent mountain ranges (Menges and Pearthree, 1989).

**Key References.** Eberly and Stanley (1978); Menges and Pearthree (1989); Leighty (1998); Spencer et al. (2001b).

### Lake Mead Area

**Geographic Extent.** Virgin River depression at junction of southwest corner of Colorado Plateau and Basin and Range, including adjacent Grand Wash trough and Overton Arm basin.

**Depocenter Type.** Two deep half grabens (Mormon and Mesquite basins); pull-apart basin of Overton Arm; Grand Wash structural trough; post-extension Muddy Creek structural sag.

**Lithology.** Interbedded sandstone, conglomerate, limestone, gypsiferous playa deposits, diagenetically altered tuffs in saline lakes, fanglomerates.

**Thickness.** 2–6 km (Bohannon et al., 1993); 5–9 km in Mesquite and Mormon subbasins based on gravity and seismic data (Langenheim et al., 2005); 1–2 km, Muddy Creek basin (Moder-Ray, 1986).

**Climatic Indicators.** Nonmarine limestone, bedded gypsum, red oxidized clastic sediments.

**Supercrop.** Pliocene-Holocene alluvium, river gravels, playa deposits, calcrite soils, basalts ﬂ ows.

**Subcrop.** Pre-extension Rainbow Gardens Member of Horse Springs Formation deposited on major angular unconformity and overlain by Thumb Member of Horse Springs Formation (Beard, 1996).

**Basal Lacuna.** 2–6 m? lacuna indicated between karst and pedogenically altered upper surface of 24.3–18.8 Ma Rainbow Gardens Member and base of 16.4–14.6 Ma Thumb Member of Horse Springs Formation (Lamb et al., 2005).

**Age at Top.** Pebble to cobble ﬂ uvial gravels interbedded with top of Muddy Creek Formation reﬂ ect initial dissection of the basin following integration of the Virgin and Colorado rivers at about 5 Ma (Lamb et al., 2005).

### Mojave Desert

**Geographic Extent.** Southeastern California from Nevada westward through the Baker and Barstow areas and northwest to the El Paso basin.

**Depocenter Types.** Basin and Range half grabens and sag basins that generally postdate and overprint early Miocene supradepositional basins except in Shadow Valley basin where thin-skinned, ductile extension began ca. 13.5 Ma and continued until ca. 10.5 Ma (Friedmann et al., 1996).

**Lithology.** Barstow Formation—basal ﬂ uvial conglomerate and sandstone overlain by lacustrine claystone, sandstone, and limestone containing tuff beds and grading laterally into course fanglomerates; Dover Spring Formation—ﬂ uvial and lacustrine deposits with interbedded basaltic ﬂ ows, tuff beds, bedded cherts, and caliche and silcrete horizons; Shadow Valley basin—chiefly alluvial fan and lacustrine facies, including gypsiferous playa deposits, rock avalanche breccias, gravity-ﬂ ight blocks changing to cobble-boulder fanglomerates at ca. 10.5 Ma associated with normal faulting of upper plate.

**Thickness.** Barstow Formation, ~1000 m; Dover Springs Formation, ~1800 m; Shadow Valley basin, ~3000 m.

**Climatic Indicators.** Nonmarine limestone, caliche, silcrete, gypsiferous playa deposits, diagenetically altered tuffs in saline lakes, fanglomerates.

**Supercrop.** Barstow Formation—unconformably overlain by Black Mountain Basalt (2.5 Ma) and Quaternary alluvium; Dover Spring Formation—unconformably overlain by Quaternary alluvium; Shadow Valley basin—unconformably overlain by 5.1–4.5 Ma basalt ﬂ ows of Cima volcanic ﬁ eld.

**Subcrop.** Barstow Formation—unconformably overlies lower Miocene Pickhandle Formation (23.7–19.3 Ma, Fillmore and Walker, 1996) and lower Miocene Mud Hills Formation (20–19 Ma, Ingersoll et al., 1996); Dover Spring Formation—unconformably overlies lower Miocene Camp Forest Formation (Whistler and Burbank, 1992), Shadow Valley strata unconformably or nonconformably overlie rocks ranging from Proterozoic to middle Miocene in age.

**Basal Lacuna.** Variable, inadequate age control.

**Age at Base.** Barstow Formation, ca. 17 Ma based on paleomagnetic and biostratigraphic data (Woodburne, 1998); Dover Spring Formation, ca. 12.5 Ma based on geochemical correlation of a tuff about 100 m above base with the Cougar Point Tuff dated by 40Ar/39Ar at 12.07 ± 0.04 Ma (Perkins et al., 2001).

**Age at Top.** Barstow Formation, ca. 12.5 Ma coeval with base of Dover Spring Formation based on biostratigraphic estimate of Tedford et al. (2004) and 13 ± 0.2 Ma isotopic age of the Lapilli Tuff that occurs ~30 m below the top of the Barstow Formation (Woodburne et al., 2000); Dover Spring Formation, ca. 7 Ma based on magnetic polarity stratigraphy and biostratigraphic data (Whistler and Burbank, 1992); Shadow Valley strata, ca. 7 Ma based on end of extension and oldest basalt ﬂ ows of cutting Cima volcanioclastic ﬁ eld (Friedmann et al., 1996).

### Key References.

Woodburne et al. (1990); Whistler and Burbank (1992); Fillmore and Walker (1996); Friedmann et al. (1996); Tedford et al. (2004).

### Monterey Formation

**Geographic Extent.** 1700 km along California coast; as far as 300 km seaward and 300 km inland.

**Depocenter Type.** Extensional and oblique-slip basins along transform margin; offshore borderlands and interior basins.

**Lithology.** Laminated diatomites, diatomaceous and siliceous mudrocks, porcelanites, calcareous and phosphatic mudrocks, dolostone, limestone.

**Thickness.** Typically 300–500 m on land, locally much thicker and thinner, varies between basins.

**Climatic Indicators.** Marine benthic and planktonic foraminifers, diatoms, radiolarians; organic-rich deposits of coated upwelling.

**Supercrop.** Late Miocene Pismo Formation; Plio-Pleistocene Foremost Formation.

**Subcrop.** Early Miocene Rincon Shale; middle Miocene Obispo Formation.

### Key References.

Woodburne et al. (1990); Whistler and Burbank (1992); Fillmore and Walker (1996); Ingersoll et al. (1996); Tedford et al. (2004).

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### REFERENCES CITED


