Diachronous episodes of Cenozoic erosion in southwestern North America and their relationship to surface uplift, paleoclimate, paleodrainage, and paleoaltimetry

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ABSTRACT
The history of erosion of southwestern North America and its relationship to surface uplift is a long-standing topic of debate. We use geologic and thermochronometric data to reconstruct the erosion history of southwestern North America. We infer that erosion events occurred mostly in response to surface uplift by contemporaneous tectonism, and were not long-delayed responses to surface uplift caused by later climate change or drainage reorganization. Rock uplift in response to isostatic compensation of exhumation occurred during each erosion event, but has been quantified only for parts of the late Miocene–Holocene erosion episode. We recognize four episodes of erosion and associated tectonic uplift: (1) the Laramide orogeny (ca. 75–45 Ma), during which individual uplifts were deeply eroded as a result of uplift by thrust faults, but Laramide basins and the Great Plains region remained near sea level, as shown by the lack of significant Laramide exhumation in these areas; (2) late middle Eocene erosion (ca. 42–37 Ma) in Wyoming, Montana, and Colorado, which probably occurred in response to episodic uplift from lithospheric rebound that followed the cessation of Laramide dynamic subsidence; (3) late Oligocene–early Miocene deep erosion (ca. 27–15 Ma) in a broad region of the southern Cordillera (including the southern Colorado Plateau, southern Great Plains, trans-Pecos Texas, and northeastern Mexico), which was uplifted in response to increased mantle buoyancy associated with major concurrent volcanism in the Sierra Madre Occidental of Mexico and in the Southern Rocky Mountains; (4) late Miocene–Holocene erosion (ca. 6–0 Ma) in a broad area of southwestern North America, with loci of deep erosion in the western Colorado–eastern Utah region and in the western Sierra Madre Occidental. Erosion in western Colorado–eastern Utah reflects mantle-related rock uplift as well as an important isostatic component caused by compensation of deep fluvial erosion in the upper Colorado River drainage following its integration to the Gulf of California. Erosion in the western Sierra Madre Occidental occurred in response to rift-shoulder uplift and the proximity of oceanic base level following the late Miocene opening of the Gulf of California. We cannot estimate the amount of rock or surface uplift associated with each erosion episode, but the maximum depths of exhumation for each were broadly similar (typically ~1–3 km). Only the most recent erosion episode is temporally correlated with climate change.

Paleoaltimetric studies, except for those based on leaf physiognomy, are generally compatible with the uplift chronology we propose here. Physiognomy-based paleoelevation data suggest that near-modern elevations were attained during the Paleogene, but are the only data that uniquely support such interpretations. High Paleogene elevations require a complex late Paleogene–Neogene uplift and subsidence history for the Front Range and western Great Plains of Colorado that is not compatible with the regional sedimentation and erosion events we describe here. Our results suggest that near-modern surface elevations in southwestern North America were generally not attained until the Neogene, and that these high elevations are the cumulative result of four major episodes of Cenozoic rock uplift of diverse origin, geographic distribution, and timing.

INTRODUCTION
Southwestern North America (herein the Colorado Plateau, the central and southern Rocky Mountains, the western Great Plains, and northern Mexico) encompasses one of Earth’s great orogenic plateaus. Despite a long history of scientific investigation, no consensus exists concerning the timing and mechanisms of uplift. The main region of high topography is in the southern Rocky Mountains of western and central Colorado (Fig. 1), where elevations are as high as 4.4 km. The southern Rocky Mountains are surrounded by broad regions of moderate to high elevation in the Colorado Plateau, central Rocky Mountains (Fig. 1), northern Rio Grande rift, and the western Great Plains. These regions are separated from the high terrain of the Sierra Madre Occidental (Fig. 1) of Mexico by the low-lying southern Basin and Range province. Modern rivers have deeply dissected some areas, including the Grand Canyon of northwestern Arizona, the barrancas of the western Sierra Madre Occidental, and numerous deep canyons in the northern Colorado Plateau–Rocky Mountains region.

The Cenozoic uplift and erosion history of southwestern North America has been a topic of research and scientific debate for more than a century (Powell, 1875). Geologists and geophysicists, invoking a variety of mechanisms, have interpreted that major episodes of regional uplift occurred during the Late Cretaceous–Eocene Laramide orogeny (McQuarrie and Chase, 2000; Humphreys et al., 2003; Huntington et al., 2010), post-Laramide magmatism (Spencer, 1996; Cather et al., 2008; Roy et al., 2004, 2009; Eaton, 2008; Liu and Gurnis, 2010; Cather and Chase, 2000; Humphreys et al., 2003; Huntington et al., 2010), post-Laramide magmatism (Spencer, 1996; Cather et al., 2008; Roy et al., 2004, 2009; Eaton, 2008; Liu and Gurnis, 2010; Cather, 2011a), the middle and late Cenozoic (Sahagian et al., 2002a; Roy et al., 2009), the middle Miocene–Holocene (Eaton, 1986, 1987, 2008; Moucha et al., 2009; Pelletier, 2009),...
Figure 1. Color-ramped digital elevation model of southwestern North America showing modern drainages and geographic regions discussed in text. Abbreviations: CRM—central Rocky Mountains; SMOc—Sierra Madre Occidental; SMOr—Sierra Madre Oriental; SRM—Southern Rocky Mountains; Gun. R.—Gunnison River.
and the late Miocene–Holocene (McMillan et al., 2006; Karlstrom et al., 2008; van Wijk et al., 2010; Levander et al., 2011). Diverse uplift mechanisms have been proposed, but most authors have emphasized the role of a single geologic process to explain the majority of uplift.

Here, we use the term rock uplift to denote the upward movement of a body of rock relative to a datum; in this case, sea level adjusted for eustatic changes. Surface uplift is rock uplift minus erosion depth. We use the term tectonic uplift for all rock uplift processes except for passive isostatic compensation in response to erosion. Climate change or drainage capture can induce deep erosion and cause rock uplift without tectonic uplift, through isostatic uplift. Erosion-induced isostatic rock uplift, however, causes a net lowering of surface elevations (Molnar and England, 1990). To explain the post-Cretaceous increase in surface elevations in southwestern North America we must invoke tectonic processes, although isostatic compensation in response to erosion was also a significant rock uplift process (Pederson et al., 2002; Lazear et al., 2011; Karlstrom et al., 2012).

In this report, we summarize stratigraphic and thermochronometric data that bear on the Cenozoic exhumation history and drainage evolution of southwestern North America. Stratigraphic relationships, such as where a younger unit is topographically inset below older rocks, and regionally extensive lacunae (times of non-deposition and erosion) provide unambiguous evidence of erosional regimes. Thermochronometry allows discernment of exhumation and cooling events, even in areas where associated degradational deposits have been removed by erosion. Stratigraphic and thermochronometric data thus provide prima facie evidence of erosion.

We argue that four episodes of deep erosion affected the region: (1) the Late Cretaceous–middle Eocene (ca. 75–45 Ma) Laramide orogeny, when local positive areas in the Colorado Plateau–Rocky Mountain region were deeply eroded; (2) the late middle Eocene (ca. 42–37 Ma), during which extensive post-Laramide pediments (the Rocky Mountain erosion surface) became incised by valleys locally >0.6 km deep, and Laramide basins in Wyoming and Colorado underwent widespread exhumation; (3) the late Oligocene–early Miocene (ca. 27–15 Ma), when a broad region (including northeastern Mexico, central and western Texas, the southern Great Plains, and the southern Colorado Plateau) along the northeast flank of the Sierra Madre Occidental volcanic field was deeply eroded, locally to 2–3 km depth; and (4) the late Miocene–Holocene uplift (ca. 6 Ma to present), during which erosion was renewed throughout the region with a locus of maximum incision (1–3 km depth) in eastern Utah and western Colorado.

It is clear that southwestern North America has undergone major surface uplift since the final withdrawal of the sea in the Late Cretaceous. It is a topic of contention, however, if erosion was synchronous with surface uplift or if erosion occurred long after uplift in response to climate change or drainage reorganization. We show that Cenozoic erosion patterns in this region cannot be reasonably explained solely by climate change or drainage reorganization. We infer that the four episodes of Cenozoic erosion in southwestern North America were related to tectonism. The loci of these four episodes are distinct in space and time, and each is associated with distinct uplift processes. These are (1) local contractile tectonism during the Laramide orogeny, (2) post-Laramide (late middle Eocene) lithospheric rebound following cessation of Laramide dynamic subsidence, (3) increased mantle buoyancy from basalt melt extraction and thermal expansion related to voluminous Oligocene–Miocene ignimbrite volcanism in the Sierra Madre Occidental and in the southern Rocky Mountains, and (4) late Cenozoic mantle-driven uplift related to dynamic uplift and thermal effects.

Although the net post-Cretaceous rock uplift is calculable or known for most areas (e.g., Pederson et al., 2002, fig. 5 therein; Kelley and Chapin, 2004), we cannot quantify the amount of rock uplift associated with individual erosion episodes. The maximum depths of erosion for each episode, however, were broadly similar. Rock uplift in response to isostatic compensation of deep erosion undoubtedly contributed to each phase of uplift, but these effects have been quantified only for parts of the youngest erosional episode. The effects of climate change were also important, but they mostly influenced regional depositional episodes and the late Miocene–Holocene interval of erosion.

We compare the timing and geographic extent of inferred uplift episodes to paleoelevation determinations from paleobotanical studies, vesicle palaeoaltimetry, stable isotopic analyses, and clumped isotopic analyses, and note that only palaeoaltimetry based on leaf physiognomy is difficult to reconcile with the surface uplift history we delineate here. We conclude that the Cenozoic surface uplift and erosion history of southwestern North America was both complex and polygenetic.

**INITIAL CONDITIONS**

Marine beds preserved in basins of the Colorado Plateau, the Rocky Mountains, and the western Great Plains provide an unambiguous record of the last time these regions were near sea level. The final regression of the Late Cretaceous Western Interior Seaway occurred during the Campanian in the Piceance Basin (Fig. 2), following deposition of the Buck tongue of the Mancos Shale, which contains *Baculites perplexus* (Gill and Hail, 1975) (ca. 79 Ma; Obradowich, 1993). The final marine regression in the San Juan Basin was in the Campanian (ca. 74 Ma; Cather, 2004), based on radioisotopically constrained biozones. The youngest marine deposits in the Black Mesa Basin are Santonian (ca. 84 Ma), but this area is landward of the subsequent Campanian transgression (Molenaar, 1983). Today the modern surface elevation of the Colorado Plateau averages ~2 km above sea level (Pederson et al., 2002).

Final withdrawal of the Western Interior Seaway from the western Great Plains was in the late Campanian (ca. 72–71 Ma in the Raton Basin; Cather, 2004; ca. 69 Ma in the Denver Basin; Raynolds, 2003). Sporadic marine inundations occurred in the Hanna Basin of south-

![Figure 2](https://pubs.geoscienceworld.org/gsa/geosphere/article-pdf/8/6/1177/3343200/1177.pdf)
Figure 2.
central Wyoming as late as the early Paleocene (Boyd and Lillegraven, 2011). The Cannonball Sea withdrew from the Williston Basin (~300 km north of the Black Hills uplift; Fig. 2) of North Dakota during the early Paleocene (ca. 64 Ma; Pepe et al., 2009). The western Great Plains today are ~0.8–1.8 km above sea level; only ~220 m of this elevation is the result of global sea-level fall since the beginning of the Laramide orogeny (Haq et al., 1987).

**LARAMIDE INTRAFORELAND EROSION**

The Laramide orogeny (ca. 75–45 Ma) produced a series of basement-involved uplifts and intervening basins throughout the Colorado Plateau–Rocky Mountain area. Deformation began in the Late Cretaceous (Campanian, ca. 75 Ma) and was mostly finished by the early middle Eocene (ca. 50–45 Ma). Evidence for Laramide erosion and cooling in the Colorado Plateau–Rocky Mountain area is restricted to local uplifts. Apatite fission track (AFT) dating has demonstrated significant exhumational cooling on many Laramide uplifts. These include the Kaibab uplift (Fig. 2; ca. 80–45 Ma; Naeser et al., 1989, 2001; Kelley et al., 2001), the western Grand Canyon (ca. 75–60 Ma; Kelley et al., 2001), the Zuni uplift (ca. 70–40 Ma; Kelley, 2003), the Nacimiento uplift (81–33 Ma; Kelley, 2003), the Gunnison uplift (ca. 55 Ma; Bryant and Naeser, 1980), the Sawatch Range (51–52 Ma; Bryant and Naeser, 1980), the Montosa uplift (65–37 Ma; Kelley et al., 1992), the northern Sierra uplift (57–45 Ma, based on clasts derived from the uplift, which subsequently subsided into the Rio Grande rift; Kelley et al., 2009), parts of the Front Range (79–45 Ma; Bryant and Naeser, 1980; 67–57 Ma; Kelley and Chapin, 2004), the Sangre de Cristo uplift (Santa Fe Range portion, 74–44 Ma; Kelley and Duncan, 1986; Kelley, 1990; Kelley and Chapin, 1995), the Laramie uplift (82–65 Ma; Kelley, 2005), the northern Medicine Bow uplift (79–60 Ma; Kelley, 2005), the Sierra Madre uplift (79–49 Ma; Kelley, 2005), the Park Range uplift (75–45 Ma; Kelley, 2005), the White River uplift (70–40 Ma; Naeser et al., 2002), the Wind River uplift (80–50 Ma; Cerveny and Steidtmann, 1993), and the Sierra Madre Oriental of northeastern Mexico (74–64 Ma; Gray et al., 2001).

Apatite (U-Th)/He thermochronometry (AHe) has documented Laramide erosion and cooling on the Bighorn uplift (ca. 65 ± 5 Ma; Crowley et al., 2002) and the north flank of the Mogollon Highland (ca. 68–37 Ma; Flowers et al., 2008). AHe data for the southwestern part of the Colorado Plateau have been interpreted by some to support uplift of the southwest part of the Colorado Plateau and the cutting of a proto–Grand Canyon during the Laramide orogeny. Erosion estimates range from limited incision of post-Paleozoic strata by the early Eocene (Flowers et al., 2008) to the cutting of a proto–Grand Canyon to nearly its present depth during the Campanian (ca. 70 Ma; Wernicke, 2011). Evidence for deep (~1200 m) incision of paleodrainages of Laramide age on the flank of the Mogollon Highland south of the Grand Canyon (Young and McKee, 1978; Young, 1979, 2001; Young and Hartman, 2011) may support such interpretations. A recent combined AFT and AHe study in the Grand Canyon area (Lee et al., 2011), however, did not find evidence of a Laramide western Grand Canyon, but instead supports deep erosion (~3 km) and canyon cutting on the Kaibab uplift (eastern Grand Canyon area) during the early Miocene.

Paleodrainage patterns during the Laramide orogeny are fairly well known (Fig. 2). Rivers entered the Colorado Plateau from the south and west. During the Late Cretaceous these rivers fed deltaic systems on the margin of the Western Interior Seaway as shorelines retreated to the east during nascent Laramide basin development. Paleogene rivers on the western and northern Colorado Plateau terminated in lake basins south of the Uinta uplift. A major Laramide river headed on the continental divide in eastern California (the California River of Wernicke, 2011) entered the southwestern Colorado Plateau possibly near the present Grand Canyon, and terminated in the Uinta Basin (Davis et al., 2010). Another major river (the Idaho River of Chetel et al., 2011) flowed into the northwestern part of the greater Green River Basin. An east-flowing precursor to the Salt River entered the Baca Basin of the southeastern Colorado Plateau from south-central Arizona (Potochnik, 1989; Cather, 2004). Paleogene rivers draining the east flank of the Laramide orogen and the southeastern Colorado Plateau flowed southeastward across the Great Plains region to depocenters in the Gulf of Mexico (Galloway et al., 2011). Laramide depocenters in the northwest Gulf of Mexico basin received sediments from sources as distant as the Cordilleran arc terranes of northern Mexico and southeastern California (Mackey et al., 2012). The Colorado Mineral Belt formed a subregional drainage divide within the Laramide foreland because of high topography associated with constructional volcanism, but perhaps also as a result of mantle buoyancy related to magmatism (see following discussions of mantle-buoyancy effects associated with magmatism in the Jemez lineament and the Sierra Madre Occidental).

At the end of the Laramide orogeny in the middle Eocene, an extensive, low- to moderate-relief pediment or subsummit surface beleved parts of Laramide uplifts in Colorado and Wyoming (late Eocene surface of Epis and Chapin, 1975) and the Defiance uplift of Arizona–New Mexico (Tsaile surface of Cooley, 1958; Schmidt, 1991). We use the term Rocky Mountain erosion surface (Evans and Chapin, 1994; Chapin and Kelley, 1997) for these surfaces. In central Colorado, the initial carving of the Rocky Mountain erosion surface was in the early middle Eocene (Bridgegerian), as the surface can be projected approximately to the top of the Laramide fill of the Denver Basin (Chapin and Kelley, 1997), which is ca. 50 Ma (Raynolds, 2003). In parts of Colorado and Wyoming, the Rocky Mountain erosion surface was modified during the onlap of the Miocene Ogallala Group (Evans, 1990).

**LATE MIDDLE EOCENE EROSION IN THE WYOMING-COLORADO AREA**

The late middle Eocene (ca. 42–37 Ma; largely the late Uintan–Duchesnean; Fig. 3) was a time of widespread erosion in Wyoming, Colorado, and parts of Montana, but not in areas to the south and west. In contrast to the Laramide orogeny when erosion was localized on discrete uplifts, the late middle Eocene erosional episode involved both uplifts and basins of the preceding Laramide deformation. There are two lines of evidence for this erosional episode. First, the Rocky Mountain erosion surface (Fig. 4) was incised by deep paleovalleys. These paleovalleys are widely developed in the Front Range and Wet Mountains of Colorado (Epis and Chapin, 1975; Scott and Taylor, 1986), in the eastern Bighorn Mountains (Weibel, 1985), Laramie Range, Medicine Bow Mountains, and southeastern Wind River Mountains of Wyoming (Evans, 1990), and locally in central Montana (Wolfe, 1964). Paleovalley depth ranges from ~60 to 300 m in central Colorado (Scott and Taylor, 1986; Chapin and Kelley, 1997) to at least 600 m in parts of northern Colorado, Wyoming, and central Montana (Evanson, 1990; Wolfe, 1964). In general, the locus of deepest paleovalley incision (~20 km; Fig. 4) in the central Rockies encompasses the southeast ern Wind River uplift, the Medicine Bow uplift, the northern Laramie uplift, and part of central Montana (Evanson, 1990; Wolfe, 1964).

In central Colorado, volcanic and volcanioclastic rocks as old as the 36.7 Ma (early Chadronian) Wall Mountain Tuff (McIntosh and Chapin, 2004) began to accumulate in paleovalleys (Chapin and Kelley, 1997) and on the low-relief interfluves between paleovalleys.
The central Colorado paleovalleys underwent episodic cutting and filling until ca. 36–35 Ma, when a dominantly aggradational regime began. In Wyoming, the lower parts of these paleovalleys typically are filled with conglomeratic and tuffaceous sediments of the Chadronian White River Group (Evanoff, 1990). The incision of paleovalleys therefore occurred between ca. 50 Ma (the age of initial development of the Rocky Mountain erosion surface) and ca. 37 Ma (the age of the earliest fill in the paleovalleys). Deep erosion (~3 km) also occurred 36.4–34.3 Ma in the Sawatch Range of central Colorado (Zimmerer, 2011), but this may be the result of local volcano-tectionism.

The second line of evidence for a late middle Eocene erosional episode is the nearly complete lack of Duchesnean-age strata in the basins of the Wyoming-Colorado region (Fig. 4; Table 1). As noted by Lillegraven and Ostresh (1988) and Lillegraven (1993), in most Wyoming basins Chadronian beds disconformably overlie Uintan or older strata; Duchesnean beds are rare. A regional lacuna in the Paleogene strata of Wyoming (Lillegraven, 1993) suggests that erosion began in the late Uintan or early Duchesnean (ca. 42–40 Ma) and continued until the early Chadronian (ca. 37 Ma). This stratigraphic relationship is clearly illustrated in the Powder River Basin, where the Chadronian White River Group overlies lower Eocene Wasatch beds at Pumpkin Buttes (Love, 1952), and in the gangplank area near Cheyenne where the White River Group overlies Upper Cretaceous strata (Lillegraven, 1993).

The scarcity of Duchesnean beds in basins of the Wyoming region is probably the result of erosion rather than nondeposition. This is shown by contemporaneous drainage incision of the Rocky Mountain surface on nearby uplifts, and the fact that these incised drainages were locally graded to the adjacent, exhumed basins. The depth of erosion in the basins is not well constrained, although AFT data from a well on the northern margin of the greater Green River Basin indicate ~2 km of erosion ca. 42 Ma (Cerveny and Steidtmann, 1993). Stable isotope data from mammalian teeth suggest that kilometer-scale relief was present in southwestern Wyoming during the Chadronian (Barton and Fricke, 2006). The late middle Eocene episode of erosion in the Wyoming region was not recognized by McMillan et al. (2006), who interpreted continuous slow subsidence and sedimentation in the Wyoming region from the end of the Laramide to the late Miocene.

South and west of the Wyoming-Colorado area, Duchesnean beds are widespread in Laramide basins (Fig. 4). Examples include the Uinta Basin (Duchesne River Formation; Lucas, 1992), the Claron Basin (Brian Head Formation; Eaton et al., 1999), the Baca Basin (Baca Formation; Cather, 2004; Prothero et al., 2004), the Galisteo Basin (Galisteo Formation; Lucas, 1982; Lucas and Williamson, 1993), the Big Bend region of Texas (Robinson et al., 2004; Kirk and Williams, 2011), and possibly the Carthage–La Joya Basin (Baca Formation; Cather, 2004, 2009). In contrast to the Wyoming-Colorado area, the Rocky Mountain erosion surface in the Chuska Mountains of the central Colorado Plateau (Tsaile surface of Cooley, 1958; Schmidt, 1991) was not dissected by fluvial incision prior to its burial by the Chuska Sandstone during the latest Eocene (ca. 34 Ma; Cather et al., 2008). We therefore interpret the area south and west of the Wyoming-Colorado region to be outside the area of late middle Eocene erosion.

The eastern and northern boundaries of the area of late middle Eocene erosion are poorly constrained. The northwestern part of the eroded area may have been contiguous with an area of middle Eocene uplift in the Idaho region that formed in response to southward-migrating magmatism and extension, as suggested by stable isotope data (e.g., Mix et al., 2011). These processes, however, were unlikely to have caused the late middle Eocene episode of erosion in Wyoming and Colorado, given the scarcity of significant magmatism (except locally along the Colorado Mineral Belt; Mutschler et al., 1987; Chapin, 2012) and the lack of contemporaneous extension. Instead, we suggest that Duchesnean erosion in Wyoming and Colorado was caused by uplift related to the cessation of Laramide subduction-driven dynamic subsidence (see following).

Insofar as is known, paleodrainage patterns during the late middle Eocene erosion event were similar to preceding Laramide patterns. An exception is the southern Front Range, where Laramide paleodrainages that carried volcanioclastic detritus northeastward from the South Park area to the Denver Basin (Tweto, 1975) were supplanted by southeast-flowing paleodrainages that cut deep paleovalleys into the Rocky Mountain surface during the late middle Eocene (Eps and Chapin, 1975; Steven...
Figure 4. Map showing areas of late middle Eocene (ca. 42–37 Ma) erosion, deposition, and inferred paleodrainages relative to isopachs for early Laramide deposition. Abbreviations for erosion control points (red) and depositional control points (blue) are keyed to Table 1. Other abbreviations: AFT—apatite fission track; Twr—eroded area beneath the White River Group; RMS—incised Rocky Mountain erosion surface; TS—unincised Tsaille erosion surface. See text for discussion.
TABLE 1. STRATIGRAPHIC AND THERMOCHRONOMETRIC CONSTRAINTS ON THE AREAL EXTENT OF LATE MIDDLE EOCENE EROSION

<table>
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<th>Locality symbol*</th>
<th>Locality Description</th>
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<tr>
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<tr>
<td>BH</td>
<td>Claron Basin, Utah</td>
<td>Duchesnean strata present in Baca Formation</td>
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<tr>
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<tr>
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<td>Duchesnean strata present in Baca Formation</td>
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<td>CJ</td>
<td>Carthage—La Joya Basin, New Mexico</td>
<td>Possible Duchesnean strata present in Baca Formation</td>
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<td>Galisteo Basin, New Mexico</td>
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<tr>
<td>BB</td>
<td>Big Bend region, Texas</td>
<td>Duchesnean strata present in various units</td>
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*Locality symbols correspond to those in Figure 4.

et al., 1997). Detritus eroded from Wyoming and northern Colorado during the late middle Eocene erosion event was mostly deposited in the Gulf of Mexico basin. The Yegua Formation (late Uintan–Duchesnean; Holroyd, 2002) records a rapid clastic influx to the northwestern Gulf of Mexico basin (Galloway et al., 2011). We suggest that the Yegua Formation is the depositional record of late middle Eocene erosion in the Rocky Mountain area.

LATE OLIGOCENE–EARLY MIOCENE EROSION IN THE SOUTHERN CORDILLERA

Beginning ~10 m.y. after the end of the late middle Eocene erosional episode in Wyoming and Colorado, a broad region to the south was deeply exhumed during the late Oligocene–early Miocene (ca. 27–15 Ma). Deep erosion associated with this event encompassed the southern Colorado Plateau, the southern Great Plains, central and western Texas, and part of northeastern Mexico.

Southern Colorado Plateau

Widespread fluvial erosion occurred on the southern Colorado Plateau during the late Oligocene–early Miocene. Recognition of this phase of erosion on the Colorado Plateau was first enabled by the reconstruction of a major eolian sand sea, the Chuska erg (Cather et al., 2008). During most of the Oligocene (ca. 33.5–27 Ma) the central and southern plateau was occupied by a large eolian erg in which sand >0.5 km thick accumulated (Fig. 5). The reconstructed top of the erg would be at a modern elevation of at least 3.0 km on the central and southern Colorado Plateau (see discussion of the erg-reconstruction methodology in Cather et al., 2008) and provides a paleogeographic datum to which the elevation of younger, inset fluvial deposits can be compared.

On the southern Colorado Plateau near the valley of the Little Colorado River (Fig. 5) and its tributaries, the Bidahochi Formation of Arizona and correlative Fence Lake Formation of New Mexico (McIntosh and Cather, 1994) are inset below the reconstructed erg top. Major erosion occurred after eolian deposition ended ca. 27 Ma and before the beginning of deposition of the Bidahochi Formation (ca. 16 Ma; Dallegge et al., 2001, 2003). In east-central Arizona, the pre-
Figure 5.
Bidahochi depth of erosion was at least 1.2 km (Fig. 5; Table 2), using a conservative estimate of the elevation of the reconstructed erg top. In contrast, post-Bidahochi (post–6 Ma) erosion there is only ~520 m. Near Escondida Mountain in west-central New Mexico (Fig. 5), the base of the Fence Lake Formation (middle to upper Miocene; McIntosh and Cather, 1994) is ~300 m beneath top of the erg deposits (Chamberlin and Harris, 1994; Chamberlin et al., 1994). In contrast, post–Fence Lake (post–6 Ma) erosion there is ~150 m. Thus, in eastern Arizona and west-central New Mexico, the late Oligocene–early Miocene depth of erosion exceeded late Miocene and younger erosion by a factor of two or more. Stratigraphic evidence for deep, early Miocene erosion on the southern Colorado Plateau has been corroborated by AHe thermochronometry near the Little Colorado River valley (Flowers et al., 2008) and in the eastern Grand Canyon area (Lee et al., 2011).

Late Oligocene–early Miocene deep erosion on the Colorado Plateau was limited largely to its southern part. Thermochronometric studies show that post-Laramide erosion on the central and northern Colorado Plateau was mostly late Miocene and younger (Kelley and Blackwell, 1990; Stockli et al., 2002; Stockli, 2005; Braschayko et al., 2008; Hoffman, 2009; Hoffman et al., 2011; Lee et al., 2011). Mild erosion during the late Oligocene–early Miocene, however, occurred as far north as Wyoming, where upper Arikareean beds unconformably overlie Chadronian or older strata in several basins (Lillegraven, 1993). At about the same time, sediment was shed from the rejuvenated Wind River uplift ca. 30–23 Ma (Steidtmann et al., 1989; Steidtmann and Middleton, 1991) and thermochronometric data indicate that the Uinta Basin began to be exhumed during the Oligocene (Kelley et al., 2007). Deep erosion (3–4 km) and cooling also occurred during the late Oligocene–early Miocene in north-central Utah, but may be the result of local footwall uplift (Armstrong et al., 2003; Ehlers et al., 2003).

**Southern Basin and Range and Rio Grande Rift**

Because extension and footwall uplift began in the Oligocene in both the southern Basin and Range and the Rio Grande rift (Spencer and Reynolds, 1991; Chapin and Cather, 1994; Mack, 2004), it is difficult to discern what effect, if any, late Oligocene–early Miocene regional erosion had in these areas. Although thermochronometric evidence for late Oligocene–early Miocene cooling has been described in mountain ranges of southern Arizona and southern New Mexico (Bryant et al., 1991; Kelley et al., 1992; Kelley and Chapin, 1997; Faxon et al., 2000; Carter et al., 2004), such cooling may be mostly the result of local footwall uplift and exhumation.

**Great Plains**

Aspects of the late Oligocene–Miocene geologic history of the southern Great Plains are similar to that of the southern Colorado Plateau. The Great Plains are capped with remnants of the middle to upper Miocene Ogallala Group, a thin (~180 m; Gustavson, 1996) fluvial and eolian succession that was derived from the uplands to the west and was deposited at much the same time (ca. 18–5 Ma in the northern plains, ca. 13–5 Ma in the south; McMillan et al., 2002; Gustavson and Winkler, 1988; Hawley, 1993; Gustavson, 1996; Chapin, 2008) as the Bidahochi Formation of the Colorado Plateau (ca. 16–6 Ma; Dallegge et al., 2001, 2003). The base of the Ogallala Group on the Great Plains is a regional unconformity that typi-

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**TABLE 2. STRATIGRAPHIC AND THERMOCHRONOMETRIC CONSTRAINTS ON DEPTH AND AREAL EXTENT OF LATE OLIGOCENE–EARLY MIOCENE EROSION**

<table>
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<td>a Estua bolson, trans-Pecos (western) Texas</td>
<td>1</td>
<td>minimum thickness of volcanic rocks missing beneath Miocene bolson fill</td>
<td>Stevens and Stevens, 2003</td>
<td></td>
</tr>
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<td>b southeastern New Mexico</td>
<td>-1.5</td>
<td>thickness of strata missing beneath Ogallala Group</td>
<td>This study; Fig. 5</td>
<td></td>
</tr>
<tr>
<td>c Capitan mountains area, New Mexico</td>
<td>&gt;2</td>
<td>pre-Ogallala erosion near 28 Ma Capitan pluton</td>
<td>This study; Fig. 5</td>
<td></td>
</tr>
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<td>d northeastern New Mexico</td>
<td>-1.0</td>
<td>thickness of strata missing beneath Ogallala Group</td>
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<td>This study; Fig. 5</td>
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<td>f southwestern Nebraska</td>
<td>-0.1</td>
<td>thickness of strata missing beneath Ogallala Group</td>
<td>Chamberlin and Harris, 1994, Chamberlin et al., 1994; Cather et al., 2008; Cather, 2011a</td>
<td></td>
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<tr>
<td>g west-central New Mexico</td>
<td>0.3</td>
<td>depth of inset of Fence Lake Formation beneath top of Chuska erg eolianites near Escondida Mountain</td>
<td>Cather et al., 2008</td>
<td></td>
</tr>
<tr>
<td>i east-central Arizona</td>
<td>1.2</td>
<td>depth of inset of Bidahochi Formation beneath top of Chuska erg eolianites near Escondida Mountain</td>
<td>Flowers et al., 2008</td>
<td></td>
</tr>
<tr>
<td>j northern Arizona</td>
<td>-1.5</td>
<td>AHe data for early Miocene exhumation</td>
<td>Lee et al., 2011, Figure 4 therein</td>
<td></td>
</tr>
<tr>
<td>k Kaibab uplift, northern Arizona</td>
<td>-2</td>
<td>AHe data for early Miocene exhumation</td>
<td>Lee et al., 2011</td>
<td></td>
</tr>
<tr>
<td>l Lee’s Ferry area, northern Arizona</td>
<td>minor</td>
<td>AHe data show no evidence of significant exhumation prior to late Miocene</td>
<td>Stockli et al., 2002; Hoffman et al., 2011</td>
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<td>m Monument uplift, southeastern Utah</td>
<td>minor</td>
<td>AHe data show no evidence of significant exhumation prior to late Miocene</td>
<td>Hoffman et al., 2011</td>
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<tr>
<td>n Canyonsland area, southeastern Utah</td>
<td>minor</td>
<td>AHe data show no evidence of significant exhumation prior to late Miocene</td>
<td>Hoffman et al., 2011</td>
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<td>o Book Cliffs, east-central Utah</td>
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<td>AHe data show no evidence of significant exhumation prior to late Miocene</td>
<td>Hoffman et al., 2011</td>
<td></td>
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<tr>
<td>p San Juan Basin, southwestern New Mexico</td>
<td>minor</td>
<td>AFT and AHe data show no major exhumation prior to late Miocene</td>
<td>Braschayko et al., 2008</td>
<td></td>
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<td>q south of Monterey, northeastern Mexico</td>
<td>-3</td>
<td>thermochronometric data show deep exhumation ca. 27–15 Ma</td>
<td>Gray et al., 2001, Armstrong et al., 2003; Ehlers et al., 2003</td>
<td></td>
</tr>
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<td>r Wasatch Mountains area, central Utah</td>
<td>3–4</td>
<td>thermochronometric data show deep exhumation during the late Oligocene–middle Miocene</td>
<td>Flowers et al., 2008</td>
<td></td>
</tr>
<tr>
<td>s Little Colorado River area, east-central Arizona</td>
<td>-1.5</td>
<td>AHe data for early Miocene exhumation</td>
<td>Corrigan et al., 1998; Ewing, 2005</td>
<td></td>
</tr>
<tr>
<td>t Edwards Plateau and Llano uplift, central Texas</td>
<td>-1.3</td>
<td>AFT data from Llano uplift; unroofing succession in lower Miocene Oakville Sandstone nearby to the east</td>
<td>Kelley and Chapin, 1995; House et al., 2003</td>
<td></td>
</tr>
<tr>
<td>u western Great Plains–southern Sangre de Cristo Mountains area</td>
<td>-2</td>
<td></td>
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</tbody>
</table>

*Locality symbols correspond to those in Figure 5, AFT—apatite fission track; AHe—apatite (U-Th)/He.
Cenozoic uplift and erosion, southwestern North America
cally exhibits 20–60 m of relief across buried
paleovalleys, and as much as 150 m of relief in
areas affected by solution subsidence of under-
lying Permian evaporites (Cronin, 1969; Borman
and Meredith, 1983; Borman et al., 1984;
Fallin, 1988; Gustavson, 1996). The Ogallala
Group laps across progressively older Paleo-
gegene and Mesozoic strata to the south (Fig. 6).
In the gangplank area (or Cheyenne tablelands)
near Cheyenne, Wyoming, the Ogallala Group
overlies lower Miocene beds of the Arikaree
Formation with a few million years of deposi-
tional hiatus. In southeastern New Mexico, the
Ogallala Group unconformably overlies Lower
Cretaceous, Triassic, and Permian beds (Cronin,
Where the Ogallala Group overlies pre-Creta-
ceous strata, the effects of southward beveling
beneath the sub-Cretaceous regional uncon-
formity must also be considered (Repenning
and Page, 1956; Bilodeau, 1986; Dickinson
et al., 1989; Potochnik, 2001; Cather, 2012).
Strata missing beneath the Ogallala Group
in southeastern New Mexico represent, at
minimum, the entire Upper Cretaceous sec-
tion. Nearby Upper Cretaceous successions
are ~1.2 km thick in the Sierra Blanca Basin
(D.J. Koning, 2010, written commun.) and
~1.2 km thick in the Raton Basin (Woodward,
1984; Cather, 2004). These thicknesses may
be extrapolated along depositional strike to the
south (e.g., Molenaar, 1983) to provide an
estimate of the thickness of Upper Cretaceous
strata now missing beneath the Ogallala in east-
er and southeastern New Mexico. Also absent
there is a thin Paleogene succession (possibly a
few hundred meters thick) that must have existed
between source areas in Laramide uplifts and
Paleogene volcanic fields to the west and devo-
centers in the Gulf of Mexico to the southeast.
The missing Upper Cretaceous and Paleogene
strata suggest that ~1.5 km of erosion occurred
prior to deposition of the Ogallala Group on the
Great Plains of eastern and southeastern New
Mexico. Thermochronometric data from the
southeastern Sangre de Cristo Mountains (Kelley
and Chapin, 1995) and from the southern Great
Plains (Kelley, 1997; Flowers and Kelley, 2007)
document widespread late Oligocene–middle
Miocene cooling that resulted both from erosion
and from decreased heat flow.
The southward-deepening erosion beneath
the Ogallala on the southern Great Plains largely
occurred between the late Oligocene and the
middle Miocene. This is most clearly illustrated
by the geologic relationships near the Capitan
intrusion, a large, late Oligocene (ca. 28 Ma;
Campbell et al., 1994; Lundberg, 1999) gra-
nitic stock in southeastern New Mexico (Fig. 5).
The Capitan intrusion attains an elevation of

Figure 6. North-south cross section from 32°N to 43°N along 103.5°W. Vertical exaggeration is 100×. Line of section is shown in Figure 5.
Stratigraphic units: P—Permian; Tri—Triassic; J—Jurassic; Kl—Lower Cretaceous; K—Cretaceous; Tw—White River Group; A—Arkose Formation (upper Eocene to lower Miocene); Gp—group.
The holocrystalline nature of the granite and the epidote-actinolite-magnetite skarn mineralogy associated with it suggest that at least 1 km of strata must have overlain the intrusion when it cooled (N.W. Dunbar, 2010, personal commun.; V.W. Lueth, 2012, personal commun.; see also Phillips, 1990). In areas north and south of the Capitan intrusion, pediment gravels correlated to the Ogallala Group by Darton (1928), Kelley (1971), Segerstrom and Ryberg (1974), Hawley (1993), and Rawling (2008) are approximately on the projected grade of the Ogallala on the Great Plains (Fig. 7). The elevation of these gravels shows that, following intrusion of the Capitan pluton at 28 Ma, at least 2 km of erosion occurred before deposition of the Ogallala Group in the middle to late Miocene. A similar argument for deep late Oligocene–early Miocene erosion near the Spanish Peaks intrusion of southern Colorado (Fig. 5) was proposed by Kelley and Chapin (1995).

Pre-Ogallala erosion significantly exceeded post-Ogallala erosion in southeastern New Mexico. As noted here, ~1.5 km of pre-Ogallala erosion occurred on the southern Great Plains in southeastern New Mexico, and post–28 Ma, pre-Ogallala erosion was at least 2 km near the Capitan intrusion. Post-Miocene erosion in the region, in contrast, was rather modest. For example, as a result of post-Ogallala (post-Miocene) fluvial incision and solution collapse (Gustavson, 1986; Hawley, 1993) the Pecos River is inset ~250–400 m below the top of the Ogallala Group on the Great Plains, and the Canadian River is currently ~220 m below the same surface. Post-Ogallala incision along modern drainages south of the Capitan intrusion is at most ~270 m. The depth of erosion during the late Oligocene–early Miocene thus exceeded that of the late Miocene–Holocene by a factor of ~4–7 in eastern and southeastern New Mexico. The depth of pre-Ogallala erosion diminishes northward on the Great Plains; north of the latitude of southern Colorado the post-Ogallala erosional event is predominant (e.g., McMillan et al., 2006).

**Western and Central Texas**

South of the Great Plains in Big Bend National Park, deposits of early Hemphillian age (ca. 8 Ma; Stevens and Stevens, 2003) are preserved in the Estufa bolson (Tor-nillo graben of Dickerson and Muehlberger, 1994). These rocks overlie an erosion surface that was cut, prior to graben subsidence, on Upper Cretaceous and Eocene strata. A thick sequence of upper Eocene–Oligocene volcanic and volcaniclastic rocks, present in the Chisos Mountains nearby to the west, was eroded before Hemphillian deposition began in the Estufa bolson. The thickness of the volcanic beds eroded between Oligocene and Hemphillian time is at least 1 km, based on the preserved thickness of such rocks in the Chisos Mountains. In contrast, since deposition ended in the Estufa bolson in the Blancan(?), incision by the Rio Grande has been ~430 m.

The Edwards Plateau–Llano uplift region of central Texas was exhumed beginning in the early Miocene. The depth of exhumation near the Llano uplift was ~1.3 km (Corrigan et al., 1998). Erosion of the Edwards Plateau–Llano uplift region is recorded by an unroofing sequence in the lower Miocene Oakville Sandstone in the Gulf of Mexico basin to the east (Ewing, 2005).
Norththeastern Mexico

Gray et al. (2001) presented thermochronometric data from oil wells that document an area of late Oligocene–early Miocene deep erosion and cooling in northeastern Mexico near Monterrey (Fig. 5). Exhumation and cooling mostly occurred ca. 27–15 Ma based on AFT and AHe data; the depth of erosion during this interval was ~3 km (Gray et al., 2001, fig. 16 therein). The northern extent of this area of deep erosion is poorly defined, so it is unknown if deep, late Oligocene–early Miocene exhumation occurred in the region between Monterrey and Big Bend.

Oligocene–Miocene Drainage Network

The Oligocene–Miocene is perhaps the least understood interval of the Cenozoic drainage evolution of southwestern North America, and was a time of profound drainage reorganization and drier climates. During the Oligocene there was erg development (Chuska Sandstone and equivalents) on the Colorado Plateau and major volcanism in surrounding areas, both of which resulted in the disruption of former Laramide drainages (Cather et al., 2008). The northern Great Plains were blanketed with loess and volcanic ash of the White River Group.

Oligocene volcanism and erg development were followed by the extensional collapse of former Laramide highlands south and west of the plateau during the Neogene. Extension and subsidence in southern Arizona caused drainage reversal of the ancestral Salt River during the late Miocene (ca. 11 Ma; Potochnik and Faulds, 1998; Potochnik, 2001). Sedimentologic evidence suggests that no major drainage exit the Colorado Plateau near the present Grand Canyon during deposition of the Muddy Creek Formation (ca. 11–6 Ma; Pederson, 2008; Forrester, 2009). Beginning ca. 6 Ma, following its poorly understood capture by the lower Colorado River near the Kaibab uplift, drainage on the western and northern Colorado Plateau exited the plateau to the southwest.

The fate of sediment eroded from the southern Colorado Plateau during late Oligocene–early Miocene erosion is uncertain. It is possible that this sediment was stored on the northern Colorado Plateau, but no evidence for a depositional episode of appropriate volume is preserved. Rivers probably did not exit the Colorado Plateau to the south or west until the extensional foun-dering of former Laramide highlands allowed for drainage reversal (cf. Wernicke, 2011). In the case of the ancestral Salt River in Arizona, this reversal did not occur until ca. 11 Ma (Potochnik and Faulds, 1998; Potochnik, 2001). It is possible that rivers draining the southern Colorado Plateau during the late Oligocene–early Miocene exited the plateau eastward to the Gulf of Mexico, before rapid subsidence in the Rio Grande rift that began in the middle Miocene (Chapin and Cather, 1994) became a major impediment to transverse rivers. No evidence for such paleo-rivers, however, is known in the rift.

It is perhaps more likely that late Oligocene–early Miocene rivers exited the Colorado Plateau to the north or northwest. Pre-Bidahoci paleovalleys on the southern Colorado Plateau clearly drained to the northwest (e.g., Love, 1989; Dickinson, 2011), and the fluvial and lacustrine deposits of the Bidahoci Formation that accumulated in these paleovalleys during the middle to late Miocene contain fossil fish with affinities to the fish faunas of the Pacific Northwest, implying a hydrologic connection to that region (Spencer et al., 2008). If this is correct, then by the Miocene the continental divide had become established near its present position in western New Mexico, far to the east of its last known position during the Laramide.

The geometry of fluvial systems that may have connected the southern Colorado Plateau with the Pacific Northwest during the Miocene is speculative. At least two possibilities exist (x and y in Fig. 5). First, as suggested by Lucchitta et al. (2011), a major drainage (x) may have exited the Colorado Plateau in southwestern Utah and then flowed northward along the relict front of the Sevier thrust belt in western Utah. This drainage geometry would explain thermochronometric evidence for the cutting of an early Miocene paleocanyon across the Kaibab uplift (Lee et al., 2011). Such a drainage pathway, however, would potentially have been compromised in the northern Basin and Range by the late Eocene–middle Miocene southward sweep of volcanism and high elevations (e.g., Mix et al., 2011) in that region.

Second, it is plausible that much or all of the Colorado Plateau was tilted northward during the late Oligocene–early Miocene in a manner similar to the southern Great Plains. This would have produced the observed deep erosion of the southern plateau and may have caused development of north-flowing drainage across the northern plateau, which did not become uplifted and deeply incised until later (late Miocene–Holocene). The possibility of a paleodrainage that flowed northward from the eastern Grand Canyon area (y in Fig. 5) was discounted by Lucchitta et al. (2011) on the basis of topographic incompatibilities that result from presumed steep fluvial paleo-gradients and a rationale that regional tilting was insufficient to allow northward drainage. However, fluvial systems did drain northward from the northern plateau during the late Oligocene–early Miocene (as recorded by the lower Browns Park Formation of northwestern Colorado; Buffler, 2003), and a major north-flowing river exited the northern Colorado Plateau in southern Wyoming (Ferguson, 2011), although probably during the late Miocene.

Sediment eroded from the Great Plains during the late Oligocene–early Miocene was deposited primarily in the Rio Grande embayment of the Gulf of Mexico basin (e.g., Galloway et al., 2011). Northward tilting of the Great Plains during the late Oligocene–early Miocene induced local development of northerly paleodrainage, as is shown by evidence for a north-flowing ancestral Pecos River in southeastern New Mexico (Cather, 2011b). Paleovalleys incised into bedrock beneath the Ogallala Group on the Great Plains, however, generally drained eastward toward the Gulf of Mexico (Fig. 5; Cronin, 1969; Weeks and Gutentag, 1981; Borman and Meredith, 1983; Borman et al., 1984; Fallin, 1988). These paleovalleys may represent a quasi-radial, late Oligocene–Early Miocene paleo-drainage system that developed in response to buoyancy modification beneath the ca. 37–23 Ma San Juan–central Colorado volcanic field to the west (e.g., Roy et al., 2004). Alternatively, the sub-Ogallala paleovalleys may be younger, middle Miocene features related to developing rift-shoulder topography following the onset of rapid extension in the Rio Grande rift.

LATE MIocene–HOLOCene EROsion

Following an episode of aggradation on the Colorado Plateau (producing the Bidahoci, Fence Lake, and Browns Park formations) and the Great Plains (Ogallala Group), the most recent phase of deep Cenozoic regional erosion began in the late Miocene and continues today. To evaluate the post-Miocene depth of incision on the western Great Plains and Colorado Plateau, we compared the elevations of modern rivers to those of adjacent, preserved tops of the Ogallala Group, the Bidahoci Formation, and the Fence Lake Formation, the upper parts of which are upper Miocene (ca. 6–5 Ma).

The depth of post-Ogallala incision on the Great Plains increases westward. At longitude 101°W–102°W, the top of the Ogallala is near modern river grade. Westward from there, the Ogallala Formation on the Great Plains rises more steeply than modern drainages (e.g., McMillan et al., 2002, 2006; Leonard, 2002; Duller et al., 2012), resulting in modern rivers becoming more deeply incised to the west. On the western Great Plains near longitude 103.5°W, the depth of post-Miocene incision is ~220–420 m (Fig. 8; Table 3). An exception

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Figure 8. Map showing area of deep late Miocene–Holocene (ca. 6–0 Ma) erosion, the Jemez volcanic lineament, incision depths of modern rivers, and exhumation depths from thermochronometric data (lowercase letters, keyed to localities in Table 3). AHe—apatite (U-Th)/He thermochronometry. See text for discussion.
TABLE 3. STRATIGRAPHIC AND THERMOCHEMOMETRIC CONSTRAINTS ON DEPTH OF LATE MIocene–HOLOCENE EROSION

<table>
<thead>
<tr>
<th>Locality</th>
<th>Erosion depth (km)</th>
<th>Remarks</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td>a North Platte River, western Nebraska</td>
<td>0.42</td>
<td>incision depth below top of Ogallala Group</td>
<td>this study, Fig. 6</td>
</tr>
<tr>
<td>b Laramie River near Laramie, Wyoming</td>
<td>0.38</td>
<td>incision depth below Ogallala remnants on Sherman surface, Laramie Range</td>
<td>this study, Fig. 6</td>
</tr>
<tr>
<td>c South Platte River, northeastern Colorado</td>
<td>0.42</td>
<td>incision depth below top of Ogallala Group</td>
<td>this study, Fig. 6</td>
</tr>
<tr>
<td>d South Platte River, central Colorado</td>
<td>–1</td>
<td>incision depth below projected top of Ogallala Group</td>
<td>this study, Fig. 7</td>
</tr>
<tr>
<td>e Arkansas River, southeastern Colorado</td>
<td>0.96</td>
<td>incision depth below top of Ogallala Group</td>
<td>this study, Fig. 6</td>
</tr>
<tr>
<td>f Canadian River, northeastern New Mexico</td>
<td>0.22</td>
<td>incision depth below top of Ogallala Group</td>
<td>this study, Fig. 6</td>
</tr>
<tr>
<td>g Pecos River, east-central New Mexico</td>
<td>0.26</td>
<td>incision depth below top of Ogallala Group</td>
<td>this study, Fig. 7</td>
</tr>
<tr>
<td>h Pecos River, east-central New Mexico</td>
<td>0.40</td>
<td>incision depth below top of Ogallala Group</td>
<td>this study, Fig. 6</td>
</tr>
<tr>
<td>i Rio Grande, trans-Pecos Texas</td>
<td>0.43</td>
<td>incision depth below top of Miocene fill in Estufa Bolson</td>
<td>Stevens and Stevens, 2003; this study</td>
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<tr>
<td>j Escondida Mountain area, west-central New Mexico</td>
<td>0.15</td>
<td>incision depth below top of Fence Lake Formation</td>
<td>Chamberlin and Harris, 1994; Chamberlin et al., 1994; Cather et al., 2008; Cather, 2011a</td>
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<tr>
<td>k Little Colorado River near Winslow, Arizona</td>
<td>0.52</td>
<td>incision depth below top of Bighiho Formation</td>
<td>Cather et al., 2008</td>
</tr>
<tr>
<td>l San Juan Basin, northwestern New Mexico</td>
<td>–1</td>
<td>AFT and AHe evidence for exhumation</td>
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</tr>
<tr>
<td>m Grand Canyon area, northwestern Arizona</td>
<td>–1</td>
<td>AHe evidence for exhumation</td>
<td>Lee et al., 2011, Fig. 4 therein</td>
</tr>
<tr>
<td>n Lee’s Ferry area, northwestern Arizona</td>
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<td>AHe evidence for exhumation</td>
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</tr>
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<td>o Monument uplift, southeastern Utah</td>
<td>1.5–2.0</td>
<td>AHe evidence for exhumation</td>
<td>Stockli et al., 2002; Hoffman et al., 2011</td>
</tr>
<tr>
<td>p Canyonlands area, southeastern Utah</td>
<td>2.0–3.0</td>
<td>AHe evidence for exhumation</td>
<td>Hoffman et al., 2011</td>
</tr>
<tr>
<td>q Colorado and Gunnison Rivers near Grand Mesa, west-central Colorado</td>
<td>1.59</td>
<td>incision depth below ca. 10 Ma basalt flows</td>
<td>Aslan et al., 2011; Cole, 2011</td>
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<tr>
<td>r Book Cliffs, east-central Utah</td>
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<td>AHe evidence for exhumation</td>
<td>Hoffman et al., 2011</td>
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<td>s Wasatch Mountains area, central Utah</td>
<td>2.0–5.0</td>
<td>thermochronometric evidence for exhumation</td>
<td>Armstrong et al., 2003; Ehlers et al., 2003</td>
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<td>t South-southeast of Monterrey, Mexico</td>
<td>–0.7</td>
<td>Burial and exhumation history for El Abra-1 well</td>
<td>Gray et al., 2001</td>
</tr>
<tr>
<td>u Rio Ruidoso, southeastern New Mexico</td>
<td>0.27</td>
<td>incision depth below top of Ogallala Formation</td>
<td>this study</td>
</tr>
</tbody>
</table>

*Locality symbols correspond to those in Figure 8. AFT—apatite fission track; AHe—apatite (U-Th)/He.

Cenozoic uplift and erosion, southwestern North America

is the Arkansas River of southeastern Colorado, where the depth of incision is nearly 1 km. Along the east flank of the Front Range, the incision depth of major drainages below the projected top of the Ogallala Group is ~1 km.

The Ogallala Group and underlying strata are broadly arched in northern New Mexico and southern Colorado. The crest of the arch corresponds approximately to the Jemez volcanic lineament, a zone of volcanism represented by the Raton volcanic field (9.0 Ma to 34 ka; Stroud and Love, 1972). Upper Miocene strata, however, are not present in these remnants, so we did not attempt to reconstruct the geometry of the Neogene fill and the depth of post-Miocene erosion in these basins. The post-Miocene depth of erosion in northern Wyoming, however, was probably large, and was modeled by McMillan et al. (2006) to be ~1 km. Elsewhere in Wyoming, northern Colorado, and along the Rio Grande rift, upper Miocene strata typically were deposited in actively subsiding basins and are thus not useful for the determination of regional incision depths.

AHe thermochronometric studies also show that deep erosion and rapid cooling occurred on much of the central and northern Colorado Plateau beginning in the late Miocene (Fig. 8). The depth of erosion was 1.0–1.5 km in the Grand Canyon–Lees Ferry region (Lee et al., 2011), 1.5–2.0 km on the Monument uplift (Hoffman, 2009; Hoffman et al., 2011), ~1.0 km in the San Juan Basin (Brashayko et al., 2008), 2–3 km in the Canyonlands area (Hoffman et al., 2011), and 1.1–1.9 km in the Book Cliffs area (Hoffman et al., 2011). In contrast, the southern Uinta Basin has undergone only modest post-Miocene erosion and cooling (Hoffman et al., 2011). Late Cenozoic cooling also occurred along the western flank of the Colorado Plateau in the Wasatch Mountains region (Armstrong et al., 2003; Ehlers et al., 2003) and in ranges along the Rio Grande rift (Kelley et al., 1992; Kelley and Chapin, 1997), but in both these areas cooling may have resulted mostly from erosion of local footwall uplifts.
Inset stratigraphic relationships and thermochronometry document a broad zone of ~1–3-km-deep late Miocene–Holocene incision that includes the southwestern, central, and northern Colorado Plateau, and the Colorado Rockies (Fig. 8). This zone encompasses much of the watershed of the upper Colorado River as well as most of the present areas of high relief and high elevation in the southwestern USA (Karlstrom et al., 2012). It forms the core of a broad region of post–late Miocene erosion that encompasses much of southwestern North America. A second focus of deep, late Miocene–Holocene incision occurred on the western flank of the Sierra Madre Occidental of Mexico, where canyons commonly exceed 1 km in depth.

RELATIONSHIP OF EROSION EPISODES TO SURFACE UPLIFT, CLIMATE CHANGE, AND DRAINAGE REORGANIZATION

There have been profound changes in the surface elevation, climate, and drainage network of southwestern North America during the Cenozoic. The surface elevation has increased from near sea level to ~1–4 km since the Late Cretaceous. It is a matter of contention, however, if episodes of erosion were associated in time and space with surface uplift or if uplift substantially preceded erosion. In the latter case, shifts in climate or the drainage network may cause deep erosion long after uplift.

In the following sections, we show that each of the four episodes of deep Cenozoic erosion in southwestern North America was associated with a plausible contemporaneous tectonic uplift process. We further argue that, except for late Miocene–Holocene erosion, none of the erosion episodes were related in a major way to climate change or drainage reorganization.

Laramide

Thermochronometric evidence for deep erosion in the Laramide orogen (ca. 75–45 Ma) is restricted to local areas of crustal contraction and uplift. Cooling ages on uplifts span the age range of the Laramide orogeny and the immediate postorogenic interval, and were broadly synchronous with unroofing successions and growth faulting in adjacent basins. There is thus no evidence for significant diachrony between Laramide uplift, erosion, cooling, and deposition. Because the structural relief between Laramide basin uplift pairs in the Colorado Plateau–Rocky Mountain area is typically ~2–10 km, thermochronometric evidence for cooling on Laramide uplifts can be reconciled without invoking regional uplift and exhumation.

No Laramide basin in the Rocky Mountain region has yielded stratigraphic or thermochronometric evidence for deep erosion during the Laramide orogeny. AFT thermochronometry in the San Juan, Piceance, and Uinta basins indicates that no Laramide exhumation occurred in these areas (Fig. 2; Kelley and Blackwell, 1990; Kelley et al., 2007; Braschayko et al., 2008). If the Laramide orogeny had produced high regional elevations, as has been interpreted by some (e.g., McQuarrie and Chase, 2000; Humphreys et al., 2003; Huntington et al., 2010), it is difficult to envision why falling base levels did not cause fluvial incision along the downstream margins of basins, particularly in the eastern Laramide basins that drained to the sea.

Perhaps more important, thermochronometric data record only post-Laramide erosion (~1–2 km deep since ca. 40 Ma) in areas southeast of the Laramide Rockies (Fig. 2; the southern Great Plains, Anadarko Basin, Llano uplift, Marathon uplift, and Wichita uplift areas; Kelley, 1997; Flowers and Kelley, 2007; Carter et al., 1998, Corrigan et al., 1998, Winkler et al., 1999), despite the fact that these areas were traversed by fluvial systems that drained the Laramide orogen and flowed to the northern Gulf of Mexico basin (Galloway et al., 2011). Given the lack of Laramide structural impoundment of surface drainage in these regions, it is unlikely they would have escaped deep erosion if near-modern elevations were attained during the Laramide orogeny.

McQuarrie and Chase (2000) argued that the present crustal thickness of the Colorado Plateau (42 ± 5 km; Sheehan et al., 1997) is too thick for the plateau to have been at sea level in the Late Cretaceous, and they noted that globally there are no areas of continental crust the breadth and thickness of the Colorado Plateau at sea level today. To explain this apparent discrepancy, McQuarrie and Chase (2000) invoked ~15 km thickening of 30-km-thick Colorado Plateau crust by Laramide mid-crustal flow eastward from the thickened Sevier hinterland, thus producing ~2 km of uplift in the Colorado Plateau. However, the crustal thickness of the Colorado Plateau is similar to much of cratonic North America (Braile et al., 1989; Mooney et al., 1998; Eaton, 2008), including areas that were near sea level in the Late Cretaceous but were largely unaffected by the Laramide orogeny (North Dakota, South Dakota, Nebraska, Kansas, Iowa, Missouri, and much of Minnesota). Kilty (2000) and Morgan (2003) also questioned the viability of the Laramide mid-crustal flow model on the basis of other geological and geophysical parameters.

The Sierras Pampeanas of Argentina are commonly invoked as a modern analogue for Laramide deformation in the Rocky Mountains, in that they host active, amagmatic, basement-involved crustal contraction over a shallow subducted slab (Alvarado et al., 2009, and references therein). There, contractual deformation has deformed the 50–60-km-thick crust of the foreland basin into thrust-bounded, basement-cored uplifts and intervening basins (Fielding and Jordan, 1988; Alvarado et al., 2009).

Despite these similarities to the Laramide Rocky Mountains, basin-floor elevations in the Sierras Pampeanas region are relatively low (~600–1400 m).

We suggest that the Laramide orogeny probably did not produce large-magnitude regional surface uplift, but rather discrete areas of high elevation resulting from local contraction and crustal thickening. Morgan (2003) calculated the effects of regional crustal contraction on Laramide surface uplift, assuming 10% pure-shear contraction (this is a reasonable average shortening value for the central and southern Rocky Mountains, but an order of magnitude too large for the Colorado Plateau). Such contraction would produce ~300 m of surface uplift (Morgan, 2003). This value, combined with the global sea-level fall during the Laramide orogeny (~50 m from 75 to 45 Ma; Haq et al., 1987) suggests that the average surface elevation of the Rocky Mountains may have been as little as ~350 m above sea level at the end of the Laramide orogeny. This is compatible with low paleoelevation determinations for late Laramide beds of the Wind River and Bighorn basins using the stable isotope geochemistry of paleosol carbonate (~500 m; Fan et al., 2011) and for the Piceance Basin using the nearest living relative paleobotanical approach (~300 m; MacGinitie, 1969; Axelrod, 1968), but is far less than the paleoelevation estimate for the Piceance Basin based on leaf physiognomy (~2.9 km; Wolfe et al., 1998, see following).

Late Middle Eocene

The onset of the late middle Eocene (ca. 42–37 Ma) erosion episode was probably related to concurrent tectonic uplift, and not to climate change or drainage-reorganization events. During the Late Cretaceous to middle Eocene Laramide orogeny, a greenhouse climate prevailed. A global transition to cooler climates began ca. 50 Ma, ending with the onset of major Antarctic glaciation at 33.5 Ma (Zachos et al., 2001). The late middle Eocene episode of erosion occurred in the Wyoming–Colorado region during this transition. This erosional interval does not correspond to any known event within the climatic transition, and the geographic restriction of the erosional event suggests that
it was not a response to climatic forcing, which tends to be subcontinental or greater in extent (e.g., Chapin, 2008).

Late middle Eocene erosion was probably not the result of drainage reorganization. This is best demonstrated for the western Great Plains of Colorado, where there is no evidence that paleo-drainage toward the Gulf of Mexico was interrupted by closed-drainage development between the Laramide orogeny and the beginning of late middle Eocene erosion. It is thus unlikely that this erosion interval was a delayed response to uplift that occurred during the preceding Laramide deformation. In fact, the eroded area corresponds approximately to an area of deep Laramide subsidence, as we discuss next.

We propose a plausible tectonic uplift explanation for late middle Eocene erosion in Wyoming and Colorado. During the early part of the Laramide orogeny, an area of deep subsidence and thick sedimentary accumulation developed in Wyoming, northeastern Utah, and northern Colorado (Cross and Pilger, 1978; Roberts and Kirschbaum, 1995; Liu et al., 2011; Jones et al., 2011; Chapin, 2012; see upper Campanian–Maastrichtian isopach map in Fig. 4). Although disrupted by major intralateral deformation, deep subsidence persisted into the late Laramide, as shown by the great thicknesses of Paleocene and lower Eocene deposits within Laramide intermontane basins of this region (e.g., McDonald, 1972). Soon after the end of Laramide deformation, however, a widespread erosional regime ensued in the same region that underwent deep subsidence in the early Laramide (see the Upper Cretaceous isopach map in Fig. 4). The area of deep Laramide subsidence has been interpreted as a response to dynamic processes. These processes include subcrustal loading by the underlying flat slab (Cross and Pilger, 1978), dynamic pull of the sinking slab (Liu et al., 2011), and slab suction induced by flow in the asthenospheric wedge between the shallowly subsiding Farallon slab and the Wyoming Archean craton (Jones et al., 2011). Regardless of the subsidence mechanism, the former region of deep Laramide subsidence became topographically inverted ca. 42–37 Ma, soon after the end of the Laramide deformation and during the initial stages of post-Laramide magmatism in the Rocky Mountains. We suggest that late middle Eocene erosion in Wyoming and Colorado was caused by lithospheric rebound related to the cessation of Laramide dynamic subsidence following the fooring of the underlying Farallon slab.

Climate change may have caused the end of the late middle Eocene erosion episode. The transition to global icehouse conditions during the late Eocene–early Oligocene (Zachos et al., 2001) corresponded to a shift to cooler and drier conditions in southwestern North America, and was marked by deposition of the Chuska Sandstone on the Colorado Plateau and the White River Group on the Great Plains (Cather et al., 2008; Evanoff et al., 1992). Deposition may have occurred in response to drier climates and diminished fluvial transportation capacities, although aggradation of the White River Group was also driven by deposition of great volumes of volcanic ash from the ignimbrite flare-up of southwestern North America. The ignimbrite flare-up itself may have contributed to the onset of the Cenozoic icehouse (Cather et al., 2009).

Late Oligocene–Early Miocene

The onset of late Oligocene–early Miocene (ca. 27–15 Ma) interval of deep erosion in the southern Cordillera ca. 27 Ma does not correspond to a known climatic event. Moreover, its subregional expression and southward-deepening character are difficult to explain by climate change. The end of this erosive period, however, may correspond to major climatic shifts in the middle Miocene that were driven by global changes in oceanic circulation, including the middle Miocene climate optimum ca. 17–15 Ma and an enhanced El Niño–Southern Oscillation beginning ca. 12 Ma (see summary in Chapin, 2008). Climate change in the middle Miocene was probably largely responsible for the onset of sedimentation on the Great Plains (Ogallala Group) and the Colorado Plateau (Bidakochi and Fence Lake formations), as well as in other areas farther west (Chapin, 2008).

An important role of drainage reorganization in producing late Oligocene–early Miocene deep erosion in the southern Cordillera can be discounted. Major closed paleo-drainages on the Colorado Plateau region (the early middle Eocene Green River lakes) preceded this erosion episode by more than 20 m.y., and no major structural impediment of Cenozoic paleo-drainage occurred in the weakly deformed Great Plains region. Therefore, closed-basin conditions in these regions could not have produced a long-delayed erosional response to uplift. Because the eroded region encompassed separate watersheds that straddled the paleo-continenal divide, it is also unlikely that drainage capture events would have induced coeval deep erosion on both the southern Colorado Plateau and the southern Great Plains.

Deep erosion during the late Oligocene–early Miocene in the southern Cordillera appears to have been spatially and temporally linked to the voluminous Sierra Madre Occidental volcanic field (Fig. 5). The broad extent of the erosion northeast of the Sierra Madre Occidental, as shown by the ~1000 km north–south extent of significant pre-Ogallala erosion on the western Great Plains, suggests that uplift was related to a buoyancy source in the mantle. In the Sierra Madre Occidental, ~3–5 x106 km3 of silicic ignimbrite was erupted ca. 46–15 Ma; peak volcanism was ca. 36–27 Ma (Cather et al., 2009).

A potential source of uplift northeast of the Sierra Madre Occidental was thermal expansion and increased lithospheric buoyancy related to basalt melt extraction from the mantle during the middle Eocene–middle Miocene ignimbrite flare-up of southwestern North America (e.g., Roy et al., 2004, and references therein). Contemporaneous, but smaller volume, volcanism along the eastern Colorado Plateau margin may have produced as much as ~4 km of rock uplift, about half of which was due to permanent buoyancy caused by basalt-melt extraction from the lithospheric mantle (Roy et al., 2004). Volcanism in the Sierra Madre Occidental requires basalt melt extraction from a huge volume of mantle (>45 x 106 km3; Farmer et al., 2008), and thus may have modified the buoyancy of the mantle over a large area. The Sierra Madre Occidental was presumably being eroded during the Oligocene–Miocene, but erosion generally did not keep pace with constructional volcanism. Steven et al. (1997) also noted evidence for regional northeast tilting and erosion of the northern Great Plains area during the early to middle Miocene, but attributed it to development of the northern Rio Grande rift.

Southward-increasing erosion, coupled with evidence for a north-flowing Miocene ancestral Pecos River in southeastern New Mexico (Cather, 2011b), indicates that a broad area of the Great Plains was uplifted and tilted gently to the northeast during the late Oligocene–early Miocene. The depth of pre-Ogallala erosion on the Great Plains increases southward over a distance of ~1000 km from western Nebraska to southeastern New Mexico, where as much as 1.5–2.0 km of pre-Ogallala section is missing. The rate of southward beveling beneath the Ogallala is thus ~1.5–2.0 m/km and appears reasonably monotonic (Fig. 6). The amount of rock uplift that occurred in southeastern New Mexico is unknown, but it was at least as large as the resultant erosion depth. Development of a broad, northeast-facing topographic ramp along the flank of the Sierra Madre Occidental may also explain the southward younging of the basal Ogallala Formation. Deposition in the northern Great Plains began ca. 18 Ma, but not until ca. 13 Ma did the Ogallala lap southward across the uplifted and deeply eroded southern Great Plains region.

The Rocky Mountain erosion surface in the Front Range also appears to have been tilted.
northward. The erosion surface generally slopes eastward (Chapin and Kelley, 1997), but it also rises gently to the south. Remnants of the Rocky Mountain erosion surface near the eastern edge of the Laramie Range and the northern Front Range are at 2.0–2.2 km elevation (Scott and Taylor, 1986), and rise southward to elevations of ~2.4–2.7 km in the southern Front Range and northern Wet Mountains. Where last seen at Greenhorn Mountain in the southern Wet Mountains, the erosion surface is beneath ca. 34 Ma andesite flows at ~3.5 km elevation. With the exception of the southern Wet Mountains area (Fig. 2), which underwent significant post-Miocene northward tilting due to arching along the Jemez volcanic lineament (Cather, 2011a), the northward component of tilt of the Rocky Mountain erosion surface (~2.0–2.5 m/km) is comparable to that required to produce the observed southward beveling beneath the Ogallala Group on the western Great Plains. Tilting of the Rocky Mountain erosion surface, however, can only be constrained to be post-Laramide and pre-Ogallala in age.

**Late Miocene–Holocene**

Beginning in the late Miocene and continuing to the present day (ca. 6–0 Ma), southwestern North America has undergone a generally erosive regime, except where the subsidence rates of local extensional basins have exceeded degradation rates. The causes of late Miocene–Holocene erosion are diverse. The episode of erosion corresponds temporally to the intensification of the North American monsoon due to the opening of the Gulf of California ca. 6 Ma (Chapin, 2008) and, beginning ca. 3 Ma, the onset of Northern Hemisphere glaciation (e.g., Pelletier, 2009). Erosion was thus probably enhanced by high-discharge events related to increased summertime thunderstorm activity beginning in the late Miocene, and by increased snow-melt runoff during the Pleistocene (e.g., Chapin, 2008; Pelletier, 2009). Part of two of the great rivers of the southwestern USA, the upper Rio Grande (Connell et al., 2005) and the lower Colorado (House et al., 2011), underwent top-down integration by spillover during the late Miocene, probably in response to increased monsoonal precipitation (Chapin, 2008).

Following the opening of the Gulf of California in the late Miocene (e.g., Umhoefer, 2011), deep erosion occurred along the western flank of the Sierra Madre Occidental. Rift-shoulder uplift and the new proximity of oceanic base level increased the gradients and erosive potential of rivers, causing deep incision of middle Miocene and older ignimbrites in the western Sierra Madre Occidental. In contrast, the Sierra Madre Oriental region (Gray et al., 2001) and the eastern flank of the Sierra Madre Occidental have undergone only modest erosion since the Miocene.

Western Colorado hosts some of the highest elevations in western North America, and forms the core of the southern Rocky Mountains epeirogen of Eaton (2008). Deep, post-6 Ma fluvial incision of western Colorado and adjacent areas occurred at the same time as the western Great Plains were being tilted eastward, as shown by steepened fluvial paleogradients of the Ogallala Group (Steven et al., 1997; McMillan et al., 2002, 2006; Leonard, 2002; Duller et al., 2012). This tilting (mostly 6–3.7 Ma; Duller et al., 2012) requires significant post-Miocene uplift of areas west of the Great Plains, and is not compatible with the concept that regional elevations were simply inherited from earlier uplift events. Moreover, thermochronometric evidence for deep, late Miocene–Holocene exhumation of the central and northern Colorado Plateau is not easily reconciled with major surface uplift inherited from Paleogene or early Neogene events. Because of the prevalence of easily erodible Mesozoic and Paleogene sedimentary successions in this region, it is difficult to envision how deep erosion would have been delayed until the late Miocene if uplift had occurred much earlier. Other than the middle Miocene Lake Hopi beds of the Bidadari Formation on the southern Colorado Plateau, there is no evidence for major Neogene internal drainage in the region. Lake Hopi (actually better regarded as a series of ephemeral lakes and ponds) can only account for a small fraction of the expected sediment budget of the upper Colorado River drainage (Dallegge et al., 2003), and thus cannot represent the principal depocenter for paleodrainages of the central and northern Colorado Plateau. Evidence for late Oligocene–Miocene rivers that flowed northward from the Colorado Plateau (Buffer, 2003; Ferguson, 2011) favors external drainage of the central and northern plateau, and thus suggests that a long-delayed erosional response to uplift of this region is unlikely.

The area of deep erosion in western Colorado and eastern Utah encompasses much of the watershed of the upper Colorado River, where surface elevations are generally lower than in surrounding regions (Fig. 1; Pederson et al., 2002; Karlstrom et al., 2012). Low surface elevations and deep fluvial incision suggest an important role for isostatic rock uplift in response to erosion in the upper Colorado River watershed. The isostatic response to rapid erosion in the upper Colorado watershed, which was the result of drainage integration with the Gulf of California and climate change, caused as much as ~1 km of late Miocene–Holocene isostatic rock uplift in southeastern Utah beginning in the late Miocene (Lazear et al., 2011; Karlstrom et al., 2012). This accounts for about one-third of the post-Cretaceous rock uplift in this region (Pederson et al., 2002, Fig. 5 therein). Similarly, isostatic compensation in response to exhumation can account for only 25%–50% of the observed eastward tilting of the western Great Plains (McMillan et al., 2002; Leonard, 2002). Although the analyses of McMillan et al. (2002) and Leonard (2002) neglected isostatic uplift related to erosion of the Front Range (Pelletier, 2009), such erosion was modest, as shown by the widespread preservation of the Rocky Mountain erosion surface. Isostatic response to erosion thus cannot solely account for the late Miocene–Holocene regional uplift in eastern Utah–western Colorado; an important tectonic component is necessary.

The great breadth of the region of late Miocene–Holocene uplift, which extends ~850 km from central Utah to the plains of eastern Colorado, implies a source of buoyancy in the mantle. Uplift has been attributed to phase changes in the lithosphere (Morgan, 2003) or thermal expansion (Eaton, 2008; Roy et al., 2009). Late Miocene–Holocene uplift of the western and northern Colorado Plateau and the Colorado Rockies also corresponds approximately to areas of inferred lithospheric delamination, mantle upwelling, and dynamic uplift (Karlstrom et al., 2008, 2011, 2012; Moucha et al., 2009; Levander et al., 2011). Some combination of these mechanisms may account for the non-isostatic portion of rock uplift in this region.

In summary, surface uplift by contemporaneous tectonism appears to be largely responsible for the erosion episodes we document here. None of the four Cenozoic erosion events in southwestern North America can be attributed exclusively to the effects of climate change or paleodrainage reorganization, although these were important factors in the late Miocene–Holocene episode. Climate change, however, probably played an important role in the initiation of regional depositional episodes, such as the Chuska–White River (late Eocene–late Oligocene) and the Ogallala–Bidadari–Browns Park (middle to late Miocene) aggradational intervals.

**RELATIONSHIP OF SURFACE UPLIFT HISTORY TO PALEOALTIMETRY**

Several methods have been employed to evaluate the Cenozoic paleoelavation of southwestern North America, including oxygen stable isotope analysis, carbonate clumped isotope thermometry, paleobotany, and paleopressure determinations from vesicles in lava. The utility of several of these methods has been disputed.
Stable Isotope Paleoaltimetry

The paleoelevation and topographic relief of Laramide basins in the Rocky Mountain region have been assessed using the oxygen stable isotope geochemistry of geologic materials (Norris et al., 1996; Dettman and Lohmann, 2000; Morrill and Koch, 2002; Fricke, 2003; Carroll et al., 2008; Doebbert et al., 2010). Such analyses have produced paleorelief estimates ranging from ~0.5 km for interbasin relief (Fricke, 2003) to as much as 3–4 km between basins and adjacent mountains (Norris et al., 1996; Dettman and Lohmann, 2000; Fan and Dettman, 2009; Fan et al., 2011). Oxygen stable isotopes from precipitation runoff, however, reflect the hypsometric mean elevation of basin catchments and the effects of evaporation (e.g., Rowley and Garzione, 2007; Davis et al., 2008, 2009), but do not uniquely constrain the paleoelevation of the basin floor. The $\delta^{18}O$ of basin-floor calcic paleosols, in contrast, records in situ precipitation and suggests that the early Eocene paleoelevations of the Wind River and Bighorn basins were relatively low (~0.5 km; Fan et al., 2011). Carbonate clumped isotope thermometry has been applied to the Bidahochi Formation, and suggests that the middle to late Miocene elevation of the south-central Colorado Plateau was similar to that of today (Huntington et al., 2010). Interpretations derived from the stable isotope compositions, however, often hinge on assumptions that are poorly known, such as terrestrial lapse rates and the initial isotopic composition of vapor source regions.

Paleobotanical Paleoaltimetry

Paleobotanical methods have been used to estimate Paleogene basin-floor elevations. To facilitate the discussion of this complex topic, we broadly divide these methods into two types: (1) the nearest living relative (or floristic) method and (2) the physiognomic (leaf shape) method. Paleoelevation estimates using leaf phlogynometry can be further divided into two types, those based on paleoenthalpy calculations and those based on the calculation of paleotemperatures using terrestrial lapse rates (see Meyer, 2007, for an excellent discussion of terrestrial lapse rates and other complexities in the calculation of paleoelevation from fossil floras).

The paleoelevation of fossil leaf assemblages has been derived from comparison to the modern temperature range of their nearest living relatives (NLR). Early studies using the NLR technique were simple qualitative comparisons (e.g., MacGinitie, 1953, 1966, 1969), whereas later NLR analyses used the modern global lapse rate (Axelrod, 1968, 1997; Axelrod and Bailey, 1976). NLR analyses typically yield low paleoelevation estimates (~300 m for ca. 45 Ma lacustrine beds of the Green River Formation and 300–900 m for the 34.1 Ma lakebeds at Florissant, Colorado; MacGinitie, 1953, 1969; Axelrod, 1968, 1997; Leopold and Clay-Poole, 2001). Axelrod (1968, p. 729) noted, “...nearly all known early Eocene, Paleocene, and Late Cretaceous floras probably lived within 1000 feet of sea level...” in the Rocky Mountain region. Low Eocene paleoelevations have also been interpreted from the presence of fossil frost-intolerant plants (palms and cycads) and animals (crocodiles and tortoises) in Laramide basins in the Wyoming region (Wing and Greenwood, 1993). Fossil tsetse flies and palms at Florissant have been inferred to support low to moderate paleoelevations and a subtropical to temperate paleoclimate (Leopold and Clay-Poole, 2001; Moe and Smith, 2005).

Subsequent authors have criticized the NLR approach because it is not quantitative, it does not adequately address the effects of global climate change, and it assumes that the habitat tolerance of plants has not evolved through time. Stable isotope data from mammalian teeth, however, also suggest that Florissant and adjoining areas were at low elevations in the late Eocene (Barton and Fricke, 2006).

Physiognomic techniques combined with terrestrial lapse-rate calculations have been used to estimate paleoelevations in the Rocky Mountain region. Such studies have yielded widely varying paleoelevation estimates (Fig. 9; Table 4), in part because of differing assumptions about terrestrial lapse rates and the mean annual temperatures in coastal regions. Ancient terrestrial lapse rates are poorly understood and difficult to model unambiguously (Meyer, 2007), and the methodology of terrestrial lapse-rate determination has been questioned (Forest et al., 1995; Wolfe et al., 1998). Important effects of sampling biases in physiognomic methods of paleoelevation determination have also been noted. For example, overrepresentation of woody plants that grew along rivers and lakes may result in underestimation of mean annual temperature by 2.5–5 °C, and thus cause overestimation of paleoelevation (Burnham et al., 2001).

The Climate-Leaf Analysis Multivariate Program (CLAMP; Wolfe, 1993) is a physiognomic method used to estimate paleotemperature, paleoenthalpy, and paleoelevation from fossil leaf assemblages without using terrestrial lapse rates. Several Paleogene leaf assemblages in the Rocky Mountain region have been analyzed using this technique (Forest et al., 1995; Gregory and McIntosh, 1996; Wolfe et al., 1998). Enthalpy-based analyses suggest high regional elevations were achieved during the Paleogene in the Rocky Mountain region (Fig. 9). For example, Wolfe et al. (1998) analyzed three late Laramide (early to middle Eocene) sites in Wyoming and Colorado, and three post-Laramide (late Eocene–Oligocene) sites in Colorado. All were interpreted to have been deposited at their present elevation, within uncertainty (±890 m), except the 34.1 Ma beds at Florissant, Colorado, which were inferred to have been deposited ~1.2 km higher than their present elevation (Wolfe et al., 1998).

Peppe et al. (2010) questioned aspects of the CLAMP method. They described several sources of error in the paleoenthalpy estimates, some of which are difficult to quantify or are unquantifiable, and noted that leaf-size bias alone may produce uncertainties in the resultant paleoelevation estimates of ±2 km or more, similar to the magnitude of Paleogene elevation estimates. Peppe et al. (2010, p. 12) concluded, “...cumulative errors and uncertainties associated with the paleoenthalpy approach are large enough to make most estimates uninformative about true paleoelevation. Therefore, we do not recommend using CLAMP to estimate enthalpy, or the paleoenthalpy method to estimate elevation.” Spicer and Yang (2010) disputed the results of Peppe et al. (2010) and argued that leaf-size biases add no more than ±50 m (1σ error) to the uncertainties of the CLAMP method.

Vesicle Paleoaltimetry

Paleoelevation estimates have been derived from atmospheric paleopressure determinations from late Cenozoic lava flows in the Colorado Plateau–Rocky Mountain region (Sahagian et al., 2002a). This method calculates paleopressure by comparing the vesicle size at the top and the bottom of lava flows of known thickness (Sahagian et al., 2002b). Vesicle paleoaltimetry requires that lava flows did not undergo inflation or deflation in thickness during emplacement. Such thickness changes, however, are common in modern lavas, and their recognition in ancient lavas is problematic (e.g., Bondre, 2003). Sahagian et al. (2002a) interpreted their vesicle data to indicate that slow uplift began ca. 25 Ma, followed by rapid uplift after ~5 Ma. Libarkin and Chase (2003) reinterpreted the data of Sahagian et al. (2002a) and concluded that the most rapid period of uplift was 20–17 Ma.

Paleoaltimetry Discussion

Figure 9 and Table 4 summarize the Cenozoic paleoaltimetry data for the Colorado Plateau–Rocky Mountain region. A great deal of variation is evident. Some of this variation results from the...
previously described regional differences in the uplift histories (Fig. 10) and the local effects of extensional tectonism in southern Utah and along the Rio Grande rift. We note also that constructional volcanic topography and increased lithospheric buoyancy beneath volcanic fields may have produced higher elevations at some vesicle-study localities than was typical regionally. Additional scatter results from differences between techniques and from the low precision of the paleoelevation determinations (standard errors [1σ] are indeterminate for the qualitative NLR method, ±400–800 m for lapse-rate methods [see summary in Meyer, 2001], ±700–890 m for the paleoenthalpy method [Forest et al., 1995; Wolfe et al., 1998; cf. Peppe et al., 2010, who claimed ±2000 m], and ±400 m for vesicle paleoaltimetry [Sahagian et al., 2002a]).

Despite these sources of variation, several trends are apparent in Figure 9. The NLR paleobotanical technique generally produces significantly lower paleoelevation estimates (and thus greater post depositional uplift) for individual localities than the paleoenthalpy and the lapse-rate physiognomic techniques, both of which suggest that near-modern surface elevations were attained in the Paleogene. Vesicle-based paleoelevation determinations, which are largely from rocks younger than those of the paleobotanical sites, favor attainment of near-modern surface elevations during the late Cenozoic, although the details of the uplift history within the Neogene are unclear.
Paleoelevation results from vesicles and paleosols are generally compatible with those of the NLR technique. The time-space aspects of Cenozoic uplift and associated erosion described here can be used to assess paleoelevation interpretations in the Colorado Plateau–Rocky Mountain region. Most of the vesicle-based paleoelevation estimates (Sahagian et al., 2002a) are from areas that flank the northern and central Colorado Plateau, a region that has had ~2 km of surface uplift since the Late Cretaceous (Pederson et al., 2002). These vesicle data suggest that low to moderate paleoelevations (relative to modern elevations) in the early Miocene gave way to near-modern paleoelevations after the late Miocene. This interpretation is compatible with thermochronometric and stratigraphic data indicating that uplift and deep erosion in this region was largely late Miocene and younger (Fig. 10). Vesicle paleoaltimetric evidence for rapid uplift 20–17 Ma (Libarkin and Chase, 2003) is restricted to the San Juan Mountains region (localities O–T; see Fig. 10). Such uplift is temporally compatible with the late Oligocene–early Miocene event nearby to the south.

### TABLE 4. PALEOELEVATION ESTIMATES AND UPLIFT DETERMINATIONS FOR SELECTED LOCALITIES IN THE COLORADO PLATEAU–ROCKY MOUNTAIN REGION

<table>
<thead>
<tr>
<th>Locality symbol*</th>
<th>Locality</th>
<th>Age (Ma)</th>
<th>Paleoelevation (km)</th>
<th>Modern elevation (km)</th>
<th>Postdepositional uplift† (km)</th>
<th>References and remarks</th>
</tr>
</thead>
<tbody>
<tr>
<td>GR</td>
<td>Green River Fm, Piceance Basin, CO</td>
<td>Uintan (ca. 45)</td>
<td>-0.3</td>
<td>2.4</td>
<td>2.1</td>
<td>MacGinitie (1969); see also Wing (1987)</td>
</tr>
<tr>
<td>WB</td>
<td>Wagonbed Fm, Bighorn Basin, Rate Homestead, WY</td>
<td>Uintan (ca. 45)</td>
<td>-0.5</td>
<td>1.7</td>
<td>1.2</td>
<td>MacGinitie (1969); see also Wing (1987)</td>
</tr>
<tr>
<td>FL</td>
<td>Florissant Fm near Florissant, CO</td>
<td>34.1†</td>
<td>-0.6</td>
<td>2.6</td>
<td>2.0</td>
<td>MacGinitie (1953)</td>
</tr>
<tr>
<td>GF</td>
<td>Galisteo Fm, northwest of Bernalillo, NM</td>
<td>Duchesnean (ca. 40)</td>
<td>0.5</td>
<td>1.8</td>
<td>1.3</td>
<td>Axelrod and Bailey (1976); also see Leopold and MacGinitie (1972)</td>
</tr>
<tr>
<td>BR</td>
<td>Barilla flora, Huester Fm, trans-Pecos, TX</td>
<td>Duchesnean (ca. 38)</td>
<td>0.3</td>
<td>1.6</td>
<td>1.3</td>
<td>Axelrod and Bailey (1976)</td>
</tr>
<tr>
<td>RR</td>
<td>Red Rock Ranch Fm, San Mateo Mts, NM</td>
<td>Ca. 35</td>
<td>1.3</td>
<td>1.8</td>
<td>0.5</td>
<td>Axelrod and Bailey (1976)</td>
</tr>
<tr>
<td>HR</td>
<td>Hermosa flora, Black Range, NM</td>
<td>ca. 35</td>
<td>2.0</td>
<td>2.6</td>
<td>0.6</td>
<td>Axelrod and Bailey (1976)</td>
</tr>
<tr>
<td>HL</td>
<td>Hillsboro flora, Black Range, NM</td>
<td>ca. 35</td>
<td>1.7</td>
<td>2.7</td>
<td>1.0</td>
<td>Axelrod and Bailey (1976)</td>
</tr>
<tr>
<td>FL</td>
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<td>34.1†</td>
<td>0.5</td>
<td>2.6</td>
<td>2.1</td>
<td>Axelrod (1997)</td>
</tr>
<tr>
<td>CR</td>
<td>Creede flora, San Juan Mountains, CO</td>
<td>27.2</td>
<td>1.5</td>
<td>2.7</td>
<td>1.2</td>
<td>Axelrod and Bailey (1976)</td>
</tr>
<tr>
<td>TF</td>
<td>Tesuque Fm near Santa Fe, NM</td>
<td>ca. 15</td>
<td>0.7</td>
<td>1.8</td>
<td>1.1</td>
<td>Axelrod and Bailey (1976)</td>
</tr>
<tr>
<td>KL</td>
<td>Kisinger Lakes flora, Ayocosa Fm, Wind River Basin, WY</td>
<td>Bridgerian (ca. 50)</td>
<td>-0.8</td>
<td>2.7</td>
<td>1.9</td>
<td>Wolfe (1994); also see Wing (1987)</td>
</tr>
<tr>
<td>RR</td>
<td>Red Rock Ranch Fm, San Mateo Mts, NM</td>
<td>ca. 35</td>
<td>-3.7</td>
<td>1.8</td>
<td>-1.9</td>
<td>Meyer (1986)</td>
</tr>
<tr>
<td>HR</td>
<td>Hermosa flora, Black Range, NM</td>
<td>ca. 35</td>
<td>-3.8</td>
<td>2.6</td>
<td>-1.2</td>
<td>Meyer (1986)</td>
</tr>
<tr>
<td>HL</td>
<td>Hillsboro flora, Black Range, NM</td>
<td>ca. 35</td>
<td>3.0</td>
<td>2.6</td>
<td>-0.4</td>
<td>Meyer (1986)</td>
</tr>
<tr>
<td>FL</td>
<td>Florissant Fm near Florissant, CO</td>
<td>34.1†</td>
<td>2.4</td>
<td>2.6</td>
<td>0.2</td>
<td>Meyer (1986)</td>
</tr>
<tr>
<td>FL</td>
<td>Florissant Fm near Florissant, CO</td>
<td>34.1†</td>
<td>2.5</td>
<td>2.6</td>
<td>0.1</td>
<td>Wolfe (1992)</td>
</tr>
<tr>
<td>FL</td>
<td>Florissant Fm near Florissant, CO</td>
<td>34.1†</td>
<td>2.3</td>
<td>2.6</td>
<td>0.3</td>
<td>Gregory and Chase (1992)</td>
</tr>
<tr>
<td>FL</td>
<td>Florissant Fm near Florissant, CO</td>
<td>34.1†</td>
<td>1.9</td>
<td>2.6</td>
<td>0.7</td>
<td>Gregory and McIntosh (1996), using Meyer (1992) method</td>
</tr>
<tr>
<td>FL</td>
<td>Florissant Fm near Florissant, CO</td>
<td>34.1†</td>
<td>3.1</td>
<td>2.6</td>
<td>-0.5</td>
<td>Gregory and McIntosh (1996), using Wolfe (1992) Method</td>
</tr>
<tr>
<td>FL</td>
<td>Florissant Fm near Florissant, CO</td>
<td>34.1†</td>
<td>2.3</td>
<td>2.6</td>
<td>0.3</td>
<td>Wolfe (1994)</td>
</tr>
<tr>
<td>PP</td>
<td>Pitch-Pinnacle flora, Sawatch Range, CO</td>
<td>34.0**</td>
<td>3.2</td>
<td>3.0</td>
<td>-0.2</td>
<td>Gregory and McIntosh (1996), using Wolfe (1992) method</td>
</tr>
<tr>
<td>PP</td>
<td>Pitch-Pinnacle flora, Sawatch Range, CO</td>
<td>34.0**</td>
<td>1.9</td>
<td>3.0</td>
<td>1.1</td>
<td>Gregory and McIntosh (1996), using Meyer (1992) method</td>
</tr>
<tr>
<td>PP</td>
<td>Pitch-Pinnacle flora, Sawatch Range, CO</td>
<td>34.0**</td>
<td>-3.4</td>
<td>3.0</td>
<td>-0.4</td>
<td>Meyer (1986), his nearby Marshall Pass flora</td>
</tr>
<tr>
<td>PL</td>
<td>Piñonita flora, San Juan Mountains, CO</td>
<td>28.2</td>
<td>1.8</td>
<td>3.0</td>
<td>1.2</td>
<td>Meyer (1986)</td>
</tr>
<tr>
<td>CR</td>
<td>Creede flora, San Juan Mountains, CO</td>
<td>27.2</td>
<td>1.8</td>
<td>2.7</td>
<td>0.9</td>
<td>Meyer (1986)</td>
</tr>
<tr>
<td>CR</td>
<td>Creede flora, San Juan Mountains, CO</td>
<td>27.2</td>
<td>-2.4</td>
<td>2.7</td>
<td>0.3</td>
<td>Wolfe and Schom (1989)</td>
</tr>
</tbody>
</table>

(cf. Sahagian et al., 2002a; Libarkin and Chase, 2003). Paleoelevation results from vesicles and paleosols are generally compatible with those of the NLR technique. The time-space aspects of Cenozoic uplift and associated erosion described here can be used to assess paleoelevation interpretations in the Colorado Plateau–Rocky Mountain region. Most of the vesicle-based paleoelevation estimates (Sahagian et al., 2002a) are from areas that flank the northern and central Colorado Plateau, a region that has had ~2 km of surface uplift since the Late Cretaceous (Pederson et al., 2002). These vesicle data suggest that low to moderate paleoelevations (relative to modern elevations) in the early Miocene gave way to near-modern paleoelevations after the late Miocene. This interpretation is compatible with thermochronometric and stratigraphic data indicating that uplift and deep erosion in this region was largely late Miocene and younger (Fig. 10). Vesicle paleoaltimetric evidence for rapid uplift 20–17 Ma (Libarkin and Chase, 2003) is restricted to the San Juan Mountains region (localities O–T; see Fig. 10). Such uplift is temporally compatible with the late Oligocene–early Miocene event nearby to the south,
but may also be related to uplift near the San Juan volcanic field (e.g., Roy et al., 2004).

Paleoelevation data from sites that are within the late Oligocene–early Miocene (ca. 27–15 Ma) area of uplift and deep erosion in the southern Cordillera are difficult to assess. The paleobotanical sites (BR, HL, HR, RR and GF; see Fig. 10) are all older than this uplift event. All except GF were deposited within constructional volcanic fields; all except BR are located on the flanks of the Rio Grande rift. The uplift histories of these sites are thus not solely the result of the epeirogenic uplift episodes that we investigate here. The inferred paleoelevations of these paleobotanical sites range from 1.3 km below to 1.9 km above their present elevation. Much of this discrepancy, however, is the result of differing paleobotanical techniques (cf. Axelrod and Bailey, 1976; Meyer, 1986).

Vesicle paleoaltimetry within or near the late Oligocene–early Miocene area of uplift and deep erosion is limited to five post-Miocene sites (J, K, L, M, and N; see Fig. 10). The paleoelevation of each of these localities was assessed by the modern global lapse rate (~5.5 °C/km).

TABLE 4. PALEOELEVATION ESTIMATES AND UPLIFT DETERMINATIONS FOR SELECTED LOCALITIES IN THE COLORADO PLATEAU–ROCKY MOUNTAIN REGION (continued)

<table>
<thead>
<tr>
<th>Locality symbol</th>
<th>Locality</th>
<th>Age (Ma)</th>
<th>Paleoelevation (km)</th>
<th>Modern elevation (km)</th>
<th>Postdepositional uplift (km)</th>
<th>References and remarks</th>
</tr>
</thead>
<tbody>
<tr>
<td>LM</td>
<td>Little Mountain flora, Green River Fm, Green River Basin, WY</td>
<td>Bridgerian (ca. 51)</td>
<td>2.1</td>
<td>2.7</td>
<td>0.6</td>
<td>Wolfe et al. (1998); also see Wing (1987)</td>
</tr>
<tr>
<td>KL</td>
<td>Kisling Lakes flora, Aycross Fm, Wind River Basin, WY</td>
<td>Bridgerian (ca. 50)</td>
<td>1.8</td>
<td>2.7</td>
<td>0.9</td>
<td>Wolfe et al. (1998); also see Wing (1987)</td>
</tr>
<tr>
<td>GR</td>
<td>Green River Fm, Piceance Basin, CO</td>
<td>Uintan (ca. 45)</td>
<td>2.9</td>
<td>2.4</td>
<td>-0.5</td>
<td>Wolfe et al. (1998)</td>
</tr>
<tr>
<td>FL</td>
<td>Florissant Fm near Florissant, CO</td>
<td>34.1^†</td>
<td>2.9</td>
<td>2.6</td>
<td>-0.3</td>
<td>Gregory and McIntosh (1996), using Forest et al. (1995) method</td>
</tr>
<tr>
<td>PP</td>
<td>Pitch-Pinnacle flora, Sawatch Range, CO</td>
<td>34.0**</td>
<td>2.0</td>
<td>3.0</td>
<td>1.0</td>
<td>Gregory and McIntosh (1996), using Forest et al. (1995) method</td>
</tr>
<tr>
<td>CR</td>
<td>Creede flora near Creede, CO</td>
<td>27.2</td>
<td>2.6</td>
<td>2.7</td>
<td>0.1</td>
<td>Wolfe et al. (1998)</td>
</tr>
<tr>
<td>CR</td>
<td>Creede flora near Creede, CO</td>
<td>27.2</td>
<td>2.8</td>
<td>2.7</td>
<td>-0.1</td>
<td>Wolfe et al. (1998)</td>
</tr>
</tbody>
</table>

Vesicle paleoaltimetry††

<table>
<thead>
<tr>
<th>Locality symbol</th>
<th>Locality</th>
<th>Age (Ma)</th>
<th>Paleoelevation (km)</th>
<th>Modern elevation (km)</th>
<th>Postdepositional uplift (km)</th>
<th>References and remarks</th>
</tr>
</thead>
<tbody>
<tr>
<td>G</td>
<td>Junction, UT</td>
<td>23</td>
<td>0.7</td>
<td>1.8</td>
<td>1.1</td>
<td>Sahagian et al. (2002a)</td>
</tr>
<tr>
<td>I</td>
<td>LaBaron Lake, UT</td>
<td>22.8</td>
<td>2.0</td>
<td>3.3</td>
<td>1.3</td>
<td>Sahagian et al. (2002a)</td>
</tr>
<tr>
<td>H</td>
<td>Maryvale, UT</td>
<td>21</td>
<td>1.0</td>
<td>2.0</td>
<td>1.0</td>
<td>Sahagian et al. (2002a)</td>
</tr>
<tr>
<td>R</td>
<td>Devils Hole, CO</td>
<td>ca. 20</td>
<td>1.1</td>
<td>3.2</td>
<td>2.2</td>
<td>Sahagian et al. (2002a)</td>
</tr>
<tr>
<td>S</td>
<td>Devils Hole, CO</td>
<td>ca. 20</td>
<td>1.7</td>
<td>3.3</td>
<td>1.6</td>
<td>Sahagian et al. (2002a)</td>
</tr>
<tr>
<td>T</td>
<td>Devils Hole, CO</td>
<td>ca. 20</td>
<td>1.2</td>
<td>3.3</td>
<td>2.1</td>
<td>Sahagian et al. (2002a)</td>
</tr>
<tr>
<td>O</td>
<td>La Jara Creek, CO</td>
<td>19.8</td>
<td>0.7</td>
<td>2.5</td>
<td>1.8</td>
<td>Sahagian et al. (2002a)</td>
</tr>
<tr>
<td>Q</td>
<td>La Jara Creek, CO</td>
<td>17.7</td>
<td>1.0</td>
<td>2.5</td>
<td>1.5</td>
<td>Sahagian et al. (2002a)</td>
</tr>
<tr>
<td>P</td>
<td>La Jara Creek, CO</td>
<td>17.6</td>
<td>1.1</td>
<td>2.5</td>
<td>1.4</td>
<td>Sahagian et al. (2002a)</td>
</tr>
<tr>
<td>F</td>
<td>Plute Reservoir, UT</td>
<td>ca. 12</td>
<td>1.0</td>
<td>2.0</td>
<td>1.0</td>
<td>Sahagian et al. (2002a)</td>
</tr>
<tr>
<td>A</td>
<td>Grand Mesa, CO</td>
<td>10.2</td>
<td>2.3</td>
<td>3.1</td>
<td>0.8</td>
<td>Sahagian et al. (2002a)</td>
</tr>
<tr>
<td>B</td>
<td>Grand Mesa, CO</td>
<td>10.2</td>
<td>1.9</td>
<td>3.3</td>
<td>1.4</td>
<td>Sahagian et al. (2002a)</td>
</tr>
<tr>
<td>C</td>
<td>Mesa Lakes, CO</td>
<td>9.5</td>
<td>1.6</td>
<td>3.2</td>
<td>1.6</td>
<td>Sahagian et al. (2002a)</td>
</tr>
<tr>
<td>D</td>
<td>Mesa Lakes, CO</td>
<td>9.5</td>
<td>1.7</td>
<td>3.2</td>
<td>1.5</td>
<td>Sahagian et al. (2002a)</td>
</tr>
<tr>
<td>E</td>
<td>Antimony, UT</td>
<td>5</td>
<td>1.8</td>
<td>1.9</td>
<td>0.1</td>
<td>Sahagian et al. (2002a)</td>
</tr>
<tr>
<td>M</td>
<td>Ojo Caliente, NM</td>
<td>4.3</td>
<td>1.4</td>
<td>2.2</td>
<td>0.8</td>
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</tr>
<tr>
<td>N</td>
<td>Tres Piedras, NM</td>
<td>3.8</td>
<td>1.4</td>
<td>2.5</td>
<td>1.1</td>
<td>Sahagian et al. (2002a)</td>
</tr>
<tr>
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<td>2.0</td>
<td>1.6</td>
<td>1.9</td>
<td>0.3</td>
<td>Sahagian et al. (2002a)</td>
</tr>
<tr>
<td>L</td>
<td>Lyman Lake, AZ</td>
<td>2.0</td>
<td>1.1</td>
<td>1.9</td>
<td>0.8</td>
<td>Sahagian et al. (2002a)</td>
</tr>
<tr>
<td>J</td>
<td>Corduroy Creek, AZ</td>
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<td>0.8</td>
<td>1.8</td>
<td>1.0</td>
<td>Sahagian et al. (2002a)</td>
</tr>
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Carbonate clumped-isotope thermometry

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<tr>
<th>Locality symbol</th>
<th>Locality</th>
<th>Age (Ma)</th>
<th>Paleoelevation (km)</th>
<th>Modern elevation (km)</th>
<th>Postdepositional uplift (km)</th>
<th>References and remarks</th>
</tr>
</thead>
<tbody>
<tr>
<td>BF</td>
<td>Bidahochi Fm, central AZ</td>
<td>ca. 16</td>
<td>1.8</td>
<td>1.7</td>
<td>0.1</td>
<td>Huntington et al. (2010)</td>
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</table>

Oxygen stable isotopes from paleosols

<table>
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<tr>
<th>Locality symbol</th>
<th>Locality</th>
<th>Age (Ma)</th>
<th>Paleoelevation (km)</th>
<th>Modern elevation (km)</th>
<th>Postdepositional uplift (km)</th>
<th>References and remarks</th>
</tr>
</thead>
<tbody>
<tr>
<td>WR</td>
<td>Wind River Basin, WY</td>
<td>ca. 53</td>
<td>0.5</td>
<td>2.3</td>
<td>1.8</td>
<td>Fan et al. (2011)</td>
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</tbody>
</table>
Figure 10. Map showing distribution of paleoaltimetry localities relative to loci of the late middle Eocene (ca. 42–37 Ma), late Oligocene–early Miocene (ca. 27–15 Ma), and late Miocene–Holocene (ca. 6–0 Ma) erosion episodes. Paleoaltimetry localities are keyed to Figure 9 and Table 4. G–H is the line of section for Figure 11.
(early middle Eocene) beds, therefore the paleoelevation of these sites must derive from the Laramide orogeny. By our reckoning, geologic and thermochronometric evidence favors the lower range of these paleoelevation estimates. This interpretation is in agreement with the results of $\delta^{18}O$ paleoaltimetry from early Eocene paleosols (site WR) that support low paleoelevations (~0.5 km; Fan et al., 2011).

The most problematic region for the determination of paleoelevation history is the Rocky Mountains of central and southern Colorado, where a great disparity of interpretation exists. This disparity is best exemplified by the paleoflora at Florissant (present elevation ~2.6 km) in the Front Range, for which no fewer than 10 paleobotanical estimates of paleoelevation have been published (site FL in Figs. 9 and 10 and Table 4). These estimates, produced using a variety of techniques, range from 2.1 km below modern elevation to 1.4 km above.

The broad range of Paleogene elevation determinations depicted in Figure 9 allows a variety of uplift-history interpretations. Here we describe some geological constraints on potential uplift histories. To facilitate discussion, we employ two hypothetical curves (X and Y, Fig. 9) to illustrate end members of the possible paleoelevation and surface uplift histories. Curve X favors paleoelevation interpretations based on leaf physiognomic data (lapse-rate and paleoenthalpy methods) and the younger part of the vesicle data. Curve Y favors the NLR data, the paleosol data, and the entirety of the vesicle data. Both curves honor evidence of eastward tilting of the Great Plains, based on evidence from the Ogallala Group, which indicates that the present slope of the Ogallala has been steepened significantly since the late Miocene (Steven et al., 1997; McMillan et al., 2002, 2006; Leonard, 2002; Duller et al., 2012). Restoration of the original gradient of the Ogallala requires that the western Denver Basin was ~1 km lower during the late Miocene (Fig. 11). Because there has been no significant offset between the Front Range and the Denver Basin since the early late Eocene (Jacob and Albertus, 1985; Leonard and Langford, 1994; Chapin and Kelley, 1997), the eastern Front Range and Florissant (star symbol in Fig. 9) must also have been at lower elevations during the late Miocene.

Post-Ogallala tilting of the western Great Plains requires ~1 km of rock uplift of the eastern Front Range and Florissant since the late Miocene (Leonard, 2002; McMillan et al., 2002, 2006). To accommodate this, the uplift history depicted by curve X would require significant subsidence during the Oligocene–Miocene from the near-modern elevations attained in the late Eocene–early Oligocene. Because the eastern...
Cenozoic uplift and erosion, southwestern North America

Front Range and western Great Plains have acted as a structural unit since the late Eocene, a roller-coaster uplift and subsidence history such as represented by curve X should be manifested by sedimentation and erosion patterns in these areas. The sedimentation and erosion history of Colorado, however, is contrary to what would be expected from curve X. The time of high Paleogene elevations might be expected to be erosional, but in fact occurred during depositional episodes on the northern Great Plains (White River Group), in the Denver Basin (Castle Rock Conglomerate), and in the Florissant and other paleovalleys in the Front Range area. Interpretation of these depositional episodes, however, is complicated by the possibility that aggradation was driven by concomitant climatic drying and increased volcanioclastic sediment supply, despite high surface elevations.

Interpretation of the late Oligocene through early Miocene subsidence represented by curve X is less ambiguous. This time interval corresponds not to a depositional regime as might be expected, but to a regime of widespread erosion on the Great Plains prior to Ogallala sedimentation. The deposition and erosion history is thus out of phase with the uplift and subsidence history represented by curve X. We conclude that geologic evidence suggests that attainment of near-modern surface elevations during the Paleogene was unlikely in the Front Range and adjacent Great Plains, and that the leaf physiognomic techniques that support such interpretations are thus questionable. A gradual attainment of modern elevations by episodic uplift throughout the Cenozoic, similar to curve Y, is more consistent with the sedimentation and erosion patterns we discern here.

CONCLUSIONS

In this report, we use stratigraphic and thermochemical data to reconstruct the time-space patterns of Cenozoic erosion in southwestern North America. We identify four erosion events. The locus of each erosion event was distinct in space and time, and each can be plausibly related to a tectonic uplift event of the same age. Only the oldest of these uplift events (the Laramide orogeny) can be attributed to crustal processes (local contractile thickening). The remaining erosion events (late middle Eocene, late Oligocene–early Miocene, and late Miocene–Holocene) are related to epeirogenic uplift that derives from processes in the mantle. Each erosion episode undoubtedly engendered isotopic responses to exhumation, but the relative role of isostatic versus tectonic uplift processes is unclear except in local areas during the late Miocene–Holocene event.

None of the erosion episodes we identify here can be attributed exclusively to the effects of climate change or paleodrainage reorganization. Only the late Miocene–Holocene erosion episode shows evidence of important, non-tectonic processes, including the strengthening of the North American monsoon ca. 6 Ma, the capture of the upper Colorado River ca. 6 Ma, and the beginning of Northern Hemisphere glaciation ca. 3 Ma. Climate events, however, correlate well with the inception of regional sedimentation episodes during the late Eocene and the middle Miocene.

Paleoerosion estimates from vesicle studies, paleosol oxygen isotope geochemistry, carbonate-clumped isotope thermometry, and nearest living relatives paleobotany are generally compatible with episodes of uplift and erosion that we interpret here. Leaf physiognomic techniques, however, typically yield Paleogene erosion estimates that are higher than seem reasonable. We note that leaf physiognomic techniques stand alone in suggesting near-modern Paleogene basin-floor elevations in southwestern North America. All other techniques support low to moderate Paleogene elevations, as does our analysis.

ACKNOWLEDGMENTS

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