History and causes of post-Laramide relief in the Rocky Mountain orogenic plateau

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ABSTRACT

The Rocky Mountain orogenic plateau has the highest mean elevation and topographic relief in the contiguous United States. The mean altitude exceeds 2 km above sea level and relief increases from 30 m in the river valleys of the Great Plains to more than 1.6 km deep in the canyons and basins of the Rocky Mountains and Colorado Plateau. Despite over a century of study, the timing and causes of elevation gain and incision in the region are unclear. Post-Laramide development of relief is thought to either result from tectonic activity or climatic change. Interpretation of which of these causes dominated is based upon reconstruction of datums developed from, and supported by, paleoelevation proxies and interpretations of landscape incision. Here we reconstruct a datum surface against which regional incision can be measured in order to evaluate late Cenozoic tectonic and climatic influences. The distribution, magnitude, and timing of post-Laramide basin filling and subsequent erosion are constrained by depositional remnants, topographic markers, and other indicators across the region. We suggest that post-Laramide basin filling resulted from slow subsidence during Oligocene to Miocene time. Incision into this basin fill surface began in late Miocene time and continues today. The pattern of incision is consistent with control by localized extensional tectonism superimposed upon regional domal surface uplift. Localized extension is associated with the projection of the Rio Grande Rift into the central Rockies, and the domal uplift generally coincides with the position of buoyant mantle anomalies interpreted at depth. If the magnitudes of incision directly reflect magnitudes of surface elevation gain, they are less than can be resolved by existing paleoelevation proxy methods. In addition, the combination of post-Laramide subsidence followed by regional surface uplift reduces the net magnitude of surface elevation change since Laramide time.

Keywords: Cenozoic, tectonic uplift, climate change, incision, Rocky Mountains, Colorado Plateau.

INTRODUCTION

The Rocky Mountain orogenic plateau (Fig. 1), an area that includes the southern and central Rockies and adjacent western Great Plains and the Colorado Plateau, has the highest mean elevations and topographic relief found in the contiguous United States. Relief increases from ~30 m in the central Great Plains to over 1.6 km deep in the Rocky Mountains and Colorado Plateau. Steep-walled canyons, broad exhumed basins, and dissected subsummit erosion surfaces all record the history of deep incision mostly by rivers that began sometime after the end of the Laramide orogeny (ca. 80–40 Ma; Dickinson et al., 1988). Traditionally, this cycle of incision has been assumed to record late Cenozoic regional epeirogeny beginning ca. 5 Ma (Trimble, 1980; Epis and Chapin, 1975; Eaton, 1987). More recently, however, climate change as the cause of incision has been invoked (c.f. Molnar and England, 1990), implying that the Rocky Mountains are a dead orogen and have gained little or possibly even lost mean elevation since the end of Laramide deformation (Gregory and Chase, 1992, 1994; McQuarrie and Chase, 2000). This latter view is supported by paleoelevation estimates derived from leaf margin analyses, which suggest no net change since Eocene time (Gregory and Chase, 1992; 1994). Additional paleoelevation proxies based on geochemical, stratigraphic, paleobarometric, and other paleobotanical methods have yielded varying results, both for and against each interpretation (Axelrod and Bailey, 1976; Wolfe et al., 1998; Sahagian et al., 2002; Poulson and John, 2003; Hay et al., 1989). As a result, the history of regional elevation change remains unresolved (cf. McQuarrie and Chase, 2000; Spencer and Patchett, 1997; Fricke, 2003).

In this study, we quantify the distribution and timing of post-Laramide aggradation and incision as a way to provide a framework for constraining the evolution of relief development and thus the possible roles of climate and tectonics in the history of elevation changes in the Rocky Mountain orogenic plateau. The pattern of post-Laramide relief is reconstructed by (1) analyzing the magnitude of mid-to-late Cenozoic deposition in basins, which records the reduction of regional relief; (2) projecting the maximum elevation of preserved basin fill deposits to reconstruct a reference surface, or envelope map, that coincides with the transition from net accumulation to net degradation; and (3) calculating the difference between the reference surface and the modern topography in order to map the magnitude of late Cenozoic incision across the entire region. This study builds on an approach taken by Pederson et al.

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GSA Bulletin; March/April 2006; v. 118; no. 3/4; p. 393–405; doi: 10.1130/B25712.1; 5 figures; 1 table; Data Repository item 2006045.

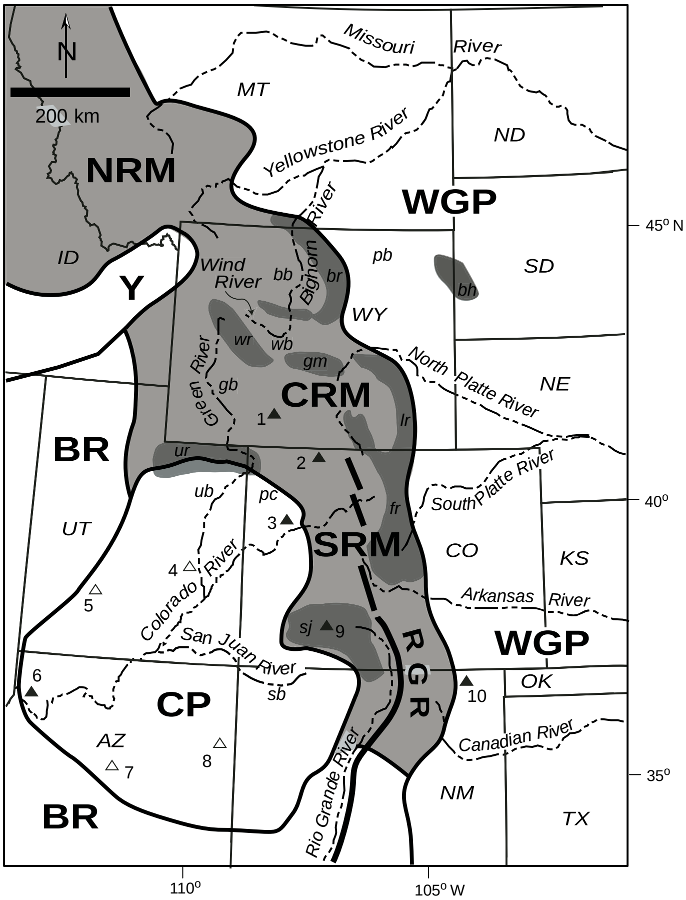

Figure 1. Location map showing major features of the Rocky Mountain orogenic plateau. Major tectonic features include the Rio Grande Rift (RGR), and the track of the Yellowstone hotspot (Y). Physiographic features include the Colorado Plateau (CP), the western Great Plains (WGP), the Basin and Range Province (BR), and the northern, central, and southern Rocky Mountains (light gray and labeled NRM, CRM, and SRM, respectively). Specific ranges (dark gray) include the Black Hills (bh), Bighorn Range (br), Wind River Range (wr), Granite Mountains (gm), Uinta Range (ur), Laramie Range (lr), Front Range (fr), and San Juan Mountains (sj). Basins include the Bighorn basin (bb), Powder River basin (pb), Wind River basin (wb), Uinta basin (ub), Green River basin (gb), San Juan basin (sb), and Piceance basin (pc). Approximate locations of volcanic flows and shallow intrusives used for age control (black triangles) or in reconstructing the basin fill surface (hollow triangles) include the Luecite Hills (1), Elkhead Mountains (2), west-central Colorado fields (3), La Sal and Henry Mountains (4), western Colorado Plateau fields (5), western Grand Canyon flows (6), southeast Colorado Plateau fields (7), south-central Colorado Plateau fields (8), San Juan volcanics (9), and Raton-Clayton field (10). Other acronyms are state name abbreviations.

(2002), except we establish a younger datum over a broader area, therefore allowing us to more tightly constrain the young history of the entire Rocky Mountain orogenic plateau. We use definitions following Molnar and England (1990), wherein rock uplift is the upward displacement of rock relative to a datum, surface uplift is rock uplift minus exhumation, and exhumation is the thickness of rock removed due to erosion or tectonism.

BACKGROUND

At the close of the Laramide orogeny (ca. 40 Ma; Dickinson et al., 1988; Erslev, 1993), the Rocky Mountains and Colorado Plateau were characterized by narrow basement-cored ranges and monoclines separated by broad structural-sedimentary basins. Overall the difference in elevation between the ridge tops and the valley floors reached proportions similar to modern, based on the depth to which preserved paleovalleys cut into an Eocene datum (referred to as the Eocene erosion surface or Rocky Mountain surface; Chapin and Kelley, 1997; Steven et al., 1997; Evanoff, 1990; Mears, 1993) and late Laramide structural and unroofing histories that suggest topographic relief of 1.5–3 km (DeCelles, et al., 1991; Hoy and Ridgway, 1997). After deformation ceased and throughout most of the middle Cenozoic, relief was progressively reduced by erosion of highlands and accumulation of predominantly fine-grained, tuffaceous fluvial and eolian sediments within the relict basins (Fig. 2; McKenna and Love, 1972). These deposits, henceforth referred to as basin fill, include the widely distributed White River Group (late Eocene to Oligocene), Arikaree Group (Oligocene to Miocene), Santa Fe Group (late Oligocene to late Miocene), the Ogallala Formation (Miocene to Pliocene), and several locally distributed correlatives (Fig. 3; Flanagan and Montagne, 1993; Lillegraven, 1993; Larson and Evanoff, 1998; Eaton, 1986). Although the youngest of these units represents the final stages of basin filling, local deposition continued during the subsequent early stages of incision, so that, in places, Ogallala Formation and equivalents onlap early incisional features. The region has continued to undergo incision since deposition of the Ogallala Formation.

The southern part of the study area experienced an early phase of erosion prior to deposition of the Ogallala Formation, as evidenced by the presence of a regional disconformity below the unit in New Mexico and Colorado (Chapin and Kelley, 1997; Steven et al., 1997). Oligocene erosion is inferred to record a period of lithosphere buoyancy in the southern Rockies (Roy et al., 2004). This event is earlier than, and separate from, the incision event described here. Thus, we agree with McKenna and Love (1972) that the Ogallala Formation and its equivalents mark a nearly continuous aggradational surface that covered most of the basins formed during the Laramide orogeny and merged with the low-relief topography in the adjacent Great Plains and Colorado Plateau (Lillegraven and Ostresh, 1988; McKenna and Love, 1972; Eaton, 1987).

METHODOLOGY

Our approach to documenting the late Cenozoic relief development in the Rocky Mountain orogenic plateau is to reconstruct the aggradational surface at the top of the basin fill and to measure subsequent incision relative to this reference surface. Our reconstruction is based on a compilation of various markers of the former topography. One data set is the distribution and thickness of post-Laramide sedimentary basin fill deposits. We use these data to determine the original extent and magnitude of net basin filling following the end of the Laramide orogeny. A second data set is the elevation and age of the top of basin fill remnants (sedimentary and associated volcanic units). We use these data to reconstruct a reference surface that represents the topography during the transition from net aggradation to net degradation and to constrain the time of turnaround from basin filling to incision. The last data set is the difference between the modern topography and the reconstructed reference surface, which we use as a measure of the magnitude of late Cenozoic incision within the Rocky mountain orogenic plateau. All of these data come from basin fill of different ages, and thus the reconstructed surface and the measurement of incision should be viewed as somewhat diachronous features representing the transition from basin filling to erosion within the region.

Figure 2. Example of basin fill remnants, in this case the White River (late Eocene–Oligocene) and Split Rock (Oligocene–Miocene) Formations capping the Beaver Rim, southern Wind River Basin, Wyoming. This example shows typical outcrop characteristics of basin fill deposits: light-colored, fine-grained, tuffaceous, fluvial, and eolian facies that predominantly form badland-type topography.

Reconstruction of Post-Laramide Basin Fill

We use the distribution and thickness of preserved basin fill remnants to determine the minimum depositional thicknesses and original extent of the fill deposits across the Rockies and western Great Plains (Fig. 4A). Remnants are preserved today as isolated paleovalley fills on range flanks, as butte caps in the middle of basins, as inverted topography on interfluves, as headward eroded rims on basin margins, and, particularly in the western Great Plains, as a continuous mantle of inset paleovalley fills. Remnants are perched at similar elevations within and between basins and have similar stratigraphy. We compiled the basin fill thickness from published and unpublished (e.g., theses) stratigraphic sections and available mappings (see GSA Data Repository DR1[1]). There are insufficient data to draw isopach maps of basin fill, so, instead, we show spot thicknesses of fill deposits across the region.

We reconstructed the reference surface, i.e., an envelope map encompassing the maximum elevation of the basin fill, by interpolating elevations between the highest, youngest preserved remnants. Input elevations were identified by extracting the maximum elevation over a 10 by 10 km square moving analysis window centered over every digital elevation model (DEM) cell from within the mapped extent of the basin fill units (Fig. 4A; data set DR2, see footnote 1).

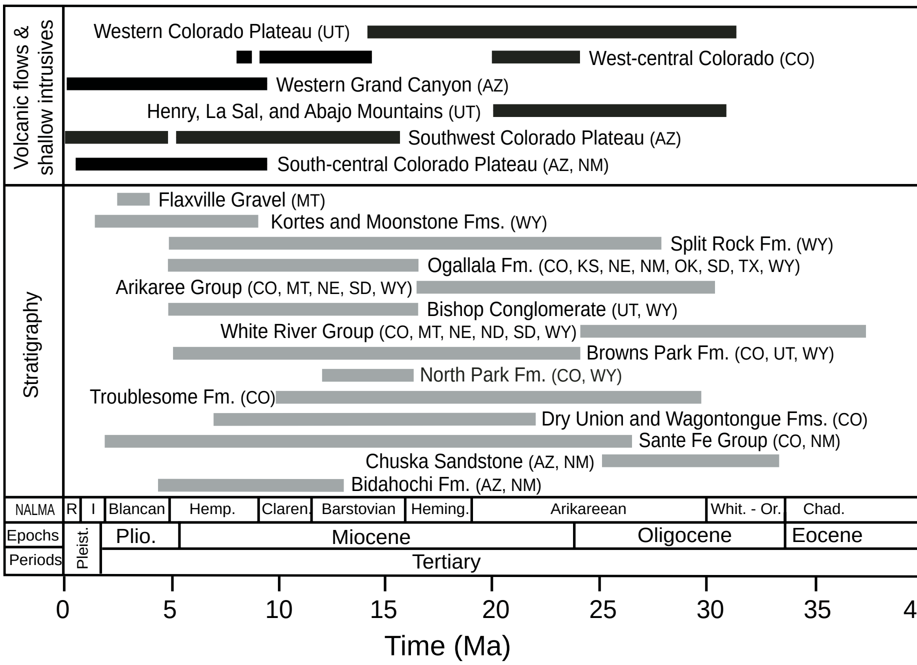

Figure 3. Age distribution of sedimentary and volcanic units used in this study. Sedimentary units were used in reconstructing basin fill thicknesses and the basin fill surface. Volcanic units were used in reconstructing the basin fill surface. The length of each bar depicts the time range of deposition or volcanic activity. Stratigraphic units are generalized and include mappable post-Laramide basin fill units. Disconformities and hiatuses are not shown. Abbreviated North American Land Mammal ages (NALMA) are Rancholabrean (R), Irvingtonian (I), Hemphillian (Hemp.), Clarendonian (Claren.), Hemingfordian (Heming.), Whitneyan-Orellan (Whit.–Or.), and Chadronian (Chad.). No vertical scale.

[1]GSA Data Repository item 2006045, data sets DR1 and DR2 used in constructing maps of incision, is available on the Web at http://www.geosociety.org/pubs/ft2006.htm. Requests may also be sent to editing@geosociety.org.

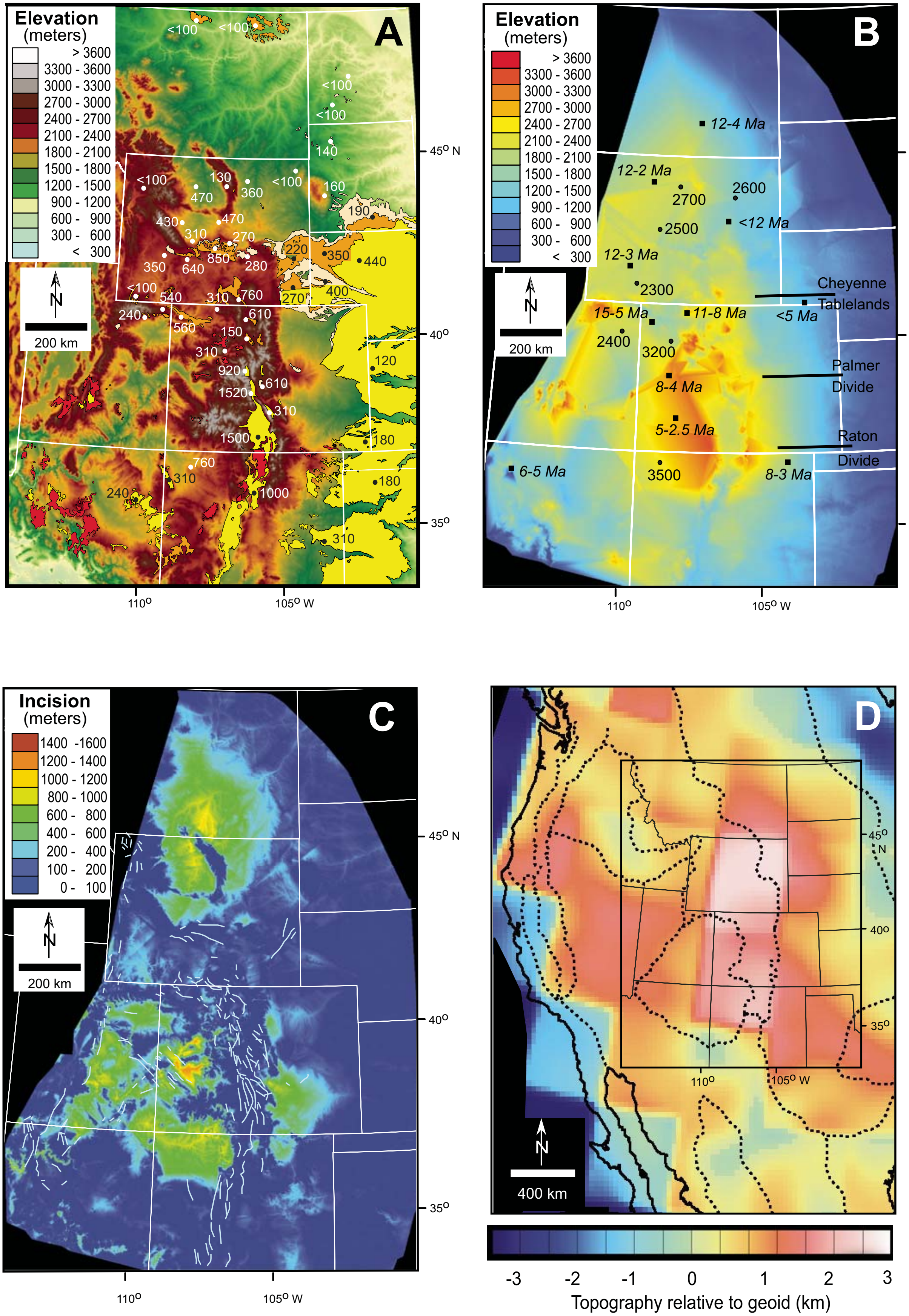
Elevation
(meters)
> 3600
3300 - 3600
3000 - 3300
2700 - 3000
2400 - 2700
2100 - 2400
1800 - 2100
1500 - 1800
1200 - 1500
900 - 1200
600 - 900
300 - 600
< 300
200 km
A
45° N
40°
35°
110°
105° W
B
12-4 Ma
12-2 Ma
2700
2600
<12 Ma
2500
12-3 Ma
2300
Cheyenne
Tablelands
15-5 Ma
11-8 Ma
<5 Ma
2400
3200
Palmer
Divide
8-4 Ma
5-2.5 Ma
Raton
Divide
6-5 Ma
3500
8-3 Ma
Incision
(meters)
1400 -1600
1200 - 1400
1000 - 1200
800 - 1000
600 - 800
400 - 600
200 - 400
100 - 200
0 - 100
C
D
400 km
-3
-2
-1
0
1
2
3
Topography relative to geoid (km)

Figure 4. (A) Shaded relief and generalized distribution of late Cenozoic sedimentary and volcanic units from which maximum elevations of preserved basin fill remnants were extracted. Mapped units are the late Eocene–Oligocene White River Group (tan), the late Oligocene–mid Miocene Arikaree Group (orange), the Miocene–early Pliocene Ogallala Formation (yellow) and their correlatives, and Miocene volcanic flows (red). Data points are thicknesses of late Cenozoic sedimentary basin fill remnants in meters. See data set DR1 (see text footnote one) for thickness references and data set DR2 for detailed list of stratigraphic units used in our analysis. (B) Color-shaded map of reconstructed basin fill surface relative to modern topography. This surface was interpolated from the maximum elevation of basin fill deposits and volcanic units (see text for explanation). Black east-west transect lines are along the interfluves between the North and South Platte Rivers (Cheyenne Tablelands), the South Platte and Arkansas Rivers (Palmer Divide), and the Arkansas and Canadian Rivers (Raton Divide). Black circles are elevations of fill top over now-eroded basins reconstructed from indirect indicators of former topography. Black squares are compiled age constraints of turnaround from degradation to incision. The first value is before incision began and the second value after incision was initiated. See data set DR3 (see text footnote one) for references. (C). Distribution of incision into the reconstructed basin fill surface in the Rocky Mountains. Color shading depicts regional incision calculated by subtracting the modern digital elevation model (DEM) (A) from the DEM of the reconstructed basin fill surface (B). Light blue solid lines are Neogene and younger extensional faults. (D). Isostatic topography calculated from lithospheric density structure down to 250 km (after Goes and van der Lee, 2002). Much of the highest topography and the difference in elevation between the Rocky Mountains and the Great Plains predicted by this model matches the pattern of maximum topography of the reconstructed basin fill surface. Scale is in kilometers relative to the geoid. Dotted lines are North American physiographic provinces and solid lines are U.S. western states and study area boundary as in Figure 1.

The distribution of basin fill units is from previously mapped digital geology at 1:500,000 scale for Montana, Wyoming, Utah, Colorado, New Mexico, and Arizona and at 1:2,500,000 scale for North Dakota, South Dakota, Nebraska, Kansas, Oklahoma, and Texas (Table 1). Elevations (Fig. 4A) are from the publicly available GTOPO30 DEM data set (~1 km horizontal resolution; U.S. Geological Survey, 1993). Interpolation between adjacent input data points was completed using a nearest-neighbor triangulation algorithm to create the reconstructed basin fill surface. While triangulation produces an artificial-looking faceted surface, it is the best method for variable density sampling points, because it preserves all of the precision of the input data while simultaneously modeling the simplest relationship between known points (Jenson and Domingue, 1988; Montgomery, 1994). Other methods such as spline interpolation produce smoother, more "realistic" surfaces (e.g., Pederson et al., 2002), but can introduce extremely steep gradients where input density changes.

Within the Colorado Plateau, where sedimentary basin fill remnants are sparse, elevations from volcanic flows and exhumed igneous intrusive units were used to reconstruct topography. Flows mark former land surface positions, at least the low parts of the topography, and exhumed intrusives provide minimum estimates of the land surface position at the time of emplacement (Larson et al., 1975; Nelson et al., 1992). These data, therefore, provide a minimum estimate of the basin fill equivalent surface in areas where no other remnants are preserved.

Because the reconstructed surface is interpolated based on the maximum elevation of preserved erosional remnants, it represents the minimum estimate of the position of the former landscape (relative to modern elevations) at the time of turnaround from net accumulation to net erosion. To provide additional limits for the reconstructed basin fill surface, we collected estimates of the position of the basin fill top by using proxy indicators of erosion depth, such as thermal burial histories, sediment compaction curves, and fission-track thermochronology profiles from wells that penetrate the major basin floors within the region (Hagen and Surdam, 1984; Nuccio and Finn, 1994; Naeser, 1986, 1992; Bond, 1984; Pitman et al., 1982). While these estimates were not incorporated into the reconstructed surface or the incision map, they do provide an independent check of the reconstructed surface and the measurement of incision depths across the region. Our results honor all of these data.

Calculating Regional Incision

The distribution and magnitude of regional incision were calculated by subtracting the

TABLE 1. POST-LARAMIDE BASIN FILL UNITS FROM WHICH MAXIMUM ELEVATIONS WERE EXTRACTED FOR RECONSTRUCTING THE REFERENCE SURFACE

State	Mappable stratigraphic units	Reference
AZ	Basaltic rocks, Pliocene to late Miocene; sedimentary rocks, Pliocene to middle Miocene (includes Bidahochi Formation); sedimentary rocks, Oligocene to Eocene (includes Chuska Sandstone)	Hirschberg and Pitts (2000)
CO	Gravels on old erosion surfaces in Front Range–Never Summer Ranges, Ogallala Formation; basalt flows (3.5–26 Ma), Browns Park Formation, North Park Formation, Arikaree Formation, Los Pinos Formation, Dry Union Formation, Santa Fe Formation, Troublesome Formation; Oligocene sedimentary rocks, White River Formation	Green (1992)
MT	Flaxville Gravel, Arikaree Formation, White River Formation	Raines and Johnson (1996)
NM	Ogallala Formation, Santa Fe Group, Chuska Sandstone, Upper Tertiary sedimentary units (includes Bidahochi Formation)	Green and Jones (1997)
UT	Basalt flows of southwestern Utah (Pliocene and Miocene), Browns Park Formation, Bishop Conglomerate, Laccolith intrusions	Hintze et al. (2000)
WY	Upper Miocene rocks, Miocene rocks, Lower Miocene rocks, Lower Miocene and Upper Oligocene rocks, Salt Lake Formation, Bishop Conglomerate, White River Formation (includes Upper Conglomerate Member, Brule Member, and Chadron Member), Red Conglomerate on top of Hoback and Wyoming Ranges	Green and Drouillard (1994)
KS, OK, SD, TX, ND, NE	Pliocene continental deposits, Miocene continental deposits, Oligocene continental deposits	Schruben et al. (1994)

Note: See appendix DR2 (see text footnote one) for detailed list and map symbols of stratigraphic units.

modern elevation from the elevation of the reconstructed basin fill surface. Positive values represent areas where rock section has been removed since the late Miocene, zero represents areas of either no incision, or, in places along mountain ranges, areas where the modern elevation is greater than the reconstructed surface. The latter case are areas where range crests and peaks were likely never covered by the basin fill (McKenna and Love, 1972; Mears, 1993; Epis and Chapin, 1975) or basinal areas that have not yet begun to incise, generally due to continued aggradation, such as by volcanism in the Yellowstone and San Juan Mountain regions. Since our reconstructed surface is based on the top of preserved basin fill deposits, we are ignoring erosion of basement rocks exposed along range crests. While range crests have experienced erosion over the past ~40 m.y., these rates are comparatively low, and erosion processes there need not directly relate to fluvial incision (Small and Anderson, 1998; Small et al., 1997). The incision data set, therefore, represents a minimum estimate of the maximum magnitude of incision for the Rocky Mountain orogenic plateau because (1) we ignore range crest erosion, and (2) in many places, the youngest, highest part of the basin fill has been stripped away leaving only older, lower parts of the section preserved.

The timing of regional incision is constrained by a compilation of the youngest ages of the highest basin fill deposits that have been cut by incision and the oldest ages of material filling (e.g., lava flows) or associated with the incision (e.g., cut and fill terraces), using published isotopic, fission-track (depositional age), and biostratigraphic age data.

RESULTS

Basin Fill Pattern

Based on the distribution of remnants, the original extent of basin fill (Fig. 4A) was nearly continuous across the western Great Plains and throughout most of Wyoming and Montana. The extent of basin fill may have been more limited in southwestern Colorado and most of Utah, Arizona, and western New Mexico if the present scarcity of data points reflects the original distribution. Fill thickness (Fig. 4A) ranges from less than 100 m up to 1500 m, with the thicker accumulations concentrated in the central portion of the Rocky Mountains. Fairly consistent magnitudes of greater than 300 m reaching up to near 900 m accumulated in a broad swath from western Nebraska into central and northern Wyoming, southern Montana, northeastern Utah, and northern Colorado. These magnitudes are considered to be minimum thicknesses, because the deposits may have once reached higher than what is preserved today (Flanagan and Montagne, 1993; Roy et al., 2004). Evidence that the fill was originally even thicker than present preservation suggests, likely over 1 km thick in many places, comes from the compilation of proxy indicators of eroded stratigraphy over now dissected basins (discussed in the following; Fig. 4B).

In Colorado and New Mexico, the thickest preserved basin fill deposits are restricted to grabens in the Rocky Mountains associated with the trajectory of the Rio Grande Rift (e.g., Eaton, 1986), with only a thin veneer of basin fill over the adjacent western Great Plains. In the Colorado Plateau, maximum basin fill thickness is hard to calculate, because there are few occurrences of late Cenozoic sedimentary units. It is possible that very little deposition occurred on the Colorado Plateau during this interval, however, exhumed Oligocene-Miocene intrusives and perched Oligocene-Miocene basalt flows indicate that the land surface, even without basin fill deposits, remained relatively high and likely merged with the basin fill surface to the northeast and east (Eaton, 1987) and that timing of turnaround from net aggradation to net degradation was similar to that of the Rocky Mountains (Hunt, 1956; Nelson et al., 1992).

The reconstructed basin fill surface (Fig. 4B) defines a broad domal highland paralleling the spine of the modern Rocky Mountains that decays away gradually to the east and north across the western Great Plains. To the south and west across the Colorado Plateau, the fill surface does not decrease much until very near the boundary with the Basin and Range Province. Many of the highest elevations on the map coincide with the locations of perched volcanic flows and the flanks of several Laramide ranges, whereas lows coincide with extensional features, such as along the Rio Grande Rift in Colorado and along a corridor centered along southwest to south-central Wyoming.

The smoothest and most robust portion of the reconstructed surface is over the western Great Plains where remnants are densely preserved. This part of the reconstructed surface has a gradient that rises from ≤1–2 m/km at the eastern edge to greater than 10 m/km near the Rocky Mountain front (Figs. 4B and 5). Farther west, peaks and depressions in the reconstructed surface may indicate that (1) the top of the fill was generally higher than found today, but did not get preserved in most places, (2) isolated deposits collected in high depressions on range crests and were not part of a continuous blanket of fill (e.g., wind-blown as opposed to water-laid deposits), and (3) structural blocks have been differentially uplifted or down-dropped after deposition. Our method of extracting elevations based on the mapped extent of highest preserved landscape markers within individual basins reduces the impact such outliers have on the overall distribution.

The reconstructed basin fill surface is lower in all cases than is suggested by our compilation of proxy indicators of basin erosion (Fig. 4B; data set DR3, see footnote one). These proxy data show that, in places, the surface once reached considerably higher than remnant elevations found today. These values are particularly useful to constrain our interpreted incision history where remnants are scarce, particularly over central Wyoming to southern Montana and over the Colorado Plateau. These data suggest that our elevation map of the reconstructed basin fill surface only shows a minimum. As such, our interpreted map of regional incision (Fig. 4C) should be taken as a minimum estimate as well.

Timing of Incision

Landscape markers that provide limits on the timing of incision are infrequent, but scattered across the region. In the Colorado Plateau, the time incision began is constrained by volcanic units along both the upper and lower reaches of the Colorado River drainage. Incision began after 6 Ma and before 5 Ma (McKee and McKee, 1972; Lucchitta, 1989) in the western Plateau, after 6 Ma and before 1.2 Ma (Pederson et al., 2002) in the Grand Canyon, and after

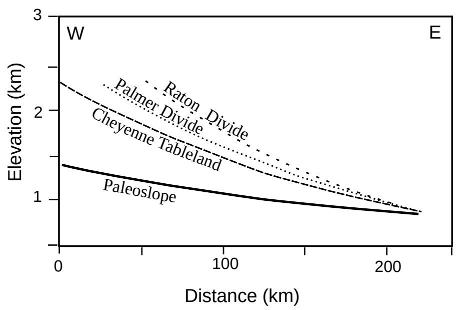

Figure 5. Gradients of major interfluves (the Cheyenne Tablelands, Palmer Divide, and Raton Divide; Fig. 4B) in the western Great Plains. The modern gradients are up to an order of magnitude greater than the paleoslope calculated from preserved gravels in the Miocene-Pliocene Ogallala Formation in the Cheyenne Tablelands (see McMillan et al., 2002). This difference suggests postdepositional tilting up to the west and of greater magnitude in the south relative to the north, of the western Great Plains.

8 Ma and before 5 Ma (Larson et al., 1975; Izett, 1975) in the northeastern Colorado Plateau. In the western Great Plains, timing is constrained by the age of volcanic flows and interbedded ash layers within the basin fill. These indicate that incision began after 5 Ma (Naeser, et al., 1980) in the northwestern Great Plains, and after 8 and before 3 Ma (Stroud, 1997; Stormer, 1972) in the southwestern Great Plains. Volcanic flows and ash-fall tuffs also constrain the timing in southern Wyoming and northern Colorado to be after 11 Ma and (Buffler, 2003; Segerstrom and Young, 1972; Snyder, 1980; Izett, 1975) and before 8 Ma (Buffler, 2003; Lange et al., 2000). Basaltic flows and evaporite collapse structures in central Colorado constrain the timing of incision to be after 8 Ma and before 4 Ma (Kunk et al., 2002; Leat et al., 1991) and after 5 Ma but before 2.5 Ma along the east-central edge of the plateau (Rye et al., 2000). North of this region, within central and northern Wyoming, Montana, and the Dakotas, timing is the least well constrained, as volcanic units are sparse or remain undated. In this area, the best available age estimations for the turnaround to incision are: after 8 Ma in the Granite Mountain area of central Wyoming, based upon U-Pb ages on detrital zircon (Scott and Chamberlain, 2002), after Clarendonian North American Land Mammal Age (ca. 8–12 Ma; Tedford et al., 1987) and before Pliocene (ca. 2 Ma) in the Bighorn basin, based on pollen fossils, and before 4–6 Ma to the north in central Montana, based on fission-track ages from zircon in ash deposits (Wayne et al., 1991). This compilation suggests that incision began regionally after 8 Ma and was established by 3–4 Ma (Fig. 4B; data set DR3, see footnote one).

Incision Pattern

The difference between the reconstructed fill surface and the modern topography represents the pattern of incision in the Rocky Mountain orogenic plateau over the past 4–8 m.y. (Fig. 4C). Incision depths are highly variable and range from 0 m to over 1.6 km. Maximum magnitudes are concentrated over the major river drainages adjacent to the flanks of the highest elevations of the Rocky Mountains in three wide (>200 km) zones: (1) the Colorado Plateau, (2) the western Great Plains of Colorado and Wyoming, and (3) central Wyoming to southern Montana. The largest of these zones is within the Colorado Plateau, where incision broadly parallels the Colorado River drainage. The thickest section has been removed from the plateau interior, where incision depths reach up to 1.6 km. Locations of deep incision occur along the upper Colorado and Green Rivers (1.6 km and 1.4 km, respectively), along the San Juan River (1.5 km), and in the Grand Canyon (1.6 km). To the south and west, away from the main drainage, incision decreases to zero.

The second zone of major incision occurs in northern Wyoming and southern Montana (Fig. 4C). Here incision has exhumed the sedimentary basins and cut several deep canyons through the mountain ranges. Erosion is greater in the more erodible finer-grained Cenozoic basin fill units, exposing the more resistant older units around the edges of each basin. Maximum incision depths occur along the major river drainages with magnitudes of 0.8 km in the Wind River Basin, 1.2 km in the Bighorn basin, and 1.2 km in the Powder River basin.

The third zone of incision is in the western Great Plains along the Rocky Mountain front. Erosion follows each major river draining the Front Range from northern New Mexico, through Colorado and into southern Wyoming (Fig. 4C). The pattern forms steep narrow canyons in the crystalline rocks and broader, yet still deep, valleys where the streams enter the softer sedimentary units of the Great Plains. Greatest incision depths occur nearest the Rocky Mountain front and narrow and decay away to zero in the downstream direction. Maximum incision decreases from south to north along the Arkansas, South Platte, and North Platte Rivers, with magnitudes of 1.2 km, 0.7 km, and 0.3 km, respectively.

DISCUSSION

Post-Laramide Basin Filling

The thickness and distribution of basin fill remnants are consistent with a sustained period of slow subsidence, aggradation, and relief reduction across at least the central Rockies following the end of the Laramide orogeny. Conditions that force basin filling include increased sediment supply and/or trapping sediment either in structurally bounded basins that previously existed but were underfilled, or in basins that undergo subsidence. Following the Laramide orogeny, there is a source of increased sediment supply in the region. The basin fill includes tuffaceous units that locally contain significant volumes of volcanic ash (up to 60%) related to ignimbrite activity in the Great Basin farther west (Larson and Evanoff, 1998), carried as ash fall into the Rocky Mountain region. These sources were abundant until ca. 25 Ma (Best et al., 1989) and so could be responsible for much of the early basin fill accumulation including the late Eocene-Oligocene age White River Group and its equivalents (Larson and Evanoff, 1998). However, the tuffaceous units make up less than half of the basin fill deposits. Reworking of these deposits may have extended the duration of increased sediment supply into the Miocene. The late Miocene turnaround of aggradation to incision is not coincident with any known demise of source area supply at that time.

Lithofacies, paleocurrent data, and clast provenance data within the thickest, most continuous accumulations indicate that much of the fill was deposited by long-lived, through-flowing river systems (Buffler, 2003; Flanagan, 1990; Love, 1970) with eolian activity on the interfluves. There is no evidence that rivers flowed into large lakes in closed basins during late Oligocene–Pliocene time in the central Rocky Mountains (Flanagan, 1990; Buffler, 2003; Love, 1970; Montagne, 1991). Large lakes would be expected if aggradation was limited to structurally distinct, isolated basins inherited from the end of Laramide deformation. Instead, accumulation took place in integrated fluvial systems that drained a significant portion of the Rocky Mountains (Lillegraven and Ostresh, 1988; Swinehart et al., 1985; Gross et al., 2001; Seeland, 1985). The exceptions to this are the thick but relatively restricted basin fill deposits of the Rio Grande Rift in southern Colorado and New Mexico that did not integrate until early Pliocene time (Chapin and Cather, 1994).

If structural traps did not force accumulations, then broadly distributed subsidence keeping pace with aggradation could account for the distribution of basin fill. Subsidence here is a relative term describing the creation of space (accommodation) in the basin in which sediment is stored. River gradients during deposition in Neogene time reached up to ~1 m/km, where calculated in the western Great Plains of Nebraska (McMillan et al., 2002). Thus, if the basins had simply aggraded with no concomitant subsidence, the near kilometer (minimum of 850 m) of basin filling in central Wyoming would have required aggradation at least 1000 km out into the Great Plains in order to maintain the calculated river gradients. There is nowhere near enough depositional thickness of late Cenozoic units in the Great Plains to allow this scenario (McMillan et al., 2002), even when isostatic adjustments are considered. Basin fill deposits in the western Great Plains are thickest, ~400 m, in the Nebraska panhandle but thin to the east to <100 m within 200 km. Instead, aggradation of basins in the Rockies must have been accompanied by subsidence of the basin floors to maintain through-flowing rivers of generally constant slope and fine grain size with the geometry of basin filling seen in this part of the Rockies.

Long-lived regional subsidence appears to have occurred across the central Rocky Mountains and western Great Plains from at least Cretaceous through much of Cenozoic time. Analysis of subsidence during pre-Laramide (Liu and Nummedal, 2004), syn-Laramide (Cross and Pilger, 1978; Mitrovica et al., 1989), and post-Laramide time (McMillan et al., 2002; Heller et al., 2003) suggests that loading due to Laramide deformation was superimposed on long-term, albeit slow, regional subsidence. Background subsidence rates across southern Wyoming varied little over time, from ~40 m/m.y. during the Sevier orogeny (Liu and Nummedal, 2004), to 20–30 m/m.y. during the Laramide orogeny (Cross and Pilger, 1978), and ~30 m/m.y. in Neogene time. While these rates are unremarkable, the results suggest that this part of North America experienced long-lived continued subsidence from Cretaceous through Miocene time. Various workers have proposed that this subcontinental-scale subsidence was due to dynamic effects associated with long-term underthrusting of western North America by Pacific basin plates (Mitrovica et al., 1989; Burgess et al., 1997; Heller et al., 2003; Liu and Nummedal, 2004).

Late Cenozoic Incision

To a first approximation, the timing and distribution of late Cenozoic incision is fairly consistent across the Rocky Mountain orogenic plateau. Everywhere incision began by late Miocene–Pliocene time (8–4 Ma) with similar maximum magnitudes, within a factor of two, from the Grand Canyon to northern Wyoming to the western Great Plains. Incision is centered over major rivers, increasing downstream relative to the crest of the Rocky Mountains and then decreasing again farther downstream.

While overall we view the Rocky Mountain orogenic plateau as broadly incisional, in detail there is a zone of reduced incision in the upstream parts of the drainages. These zones of low incision are not limited to bedrock massifs but include basins distributed along the spine of the Rockies in Colorado and across southern Wyoming (Fig. 4C). Two reasons may explain the lack of incision in these areas. First, at least in some cases, these areas lie near the headwaters of drainages. As such, if incision is associated with knickpoint retreat, then possibly headward erosion has yet to reach these upstream limits. Secondly, areas of low incision are coincident with a broad axis of young extensional faulting that dominates the backbone of the Rockies (Fig. 4C). The western Great Plains zone of low incision follows the projection of the Rio Grande Rift from central New Mexico through central Colorado to the Wyoming border (Tweto, 1979; Buffler, 2003; Chapin and Cather, 1994). The southern tier of Wyoming also contains many late Cenozoic normal faults of generally low displacement that trend E-W to NW-SE (Love, 1970; Mears 1998). The association of these tectonically active trends with the zones of low incision suggests a causative relationship (discussed in the following). For example, young graben formation may force local basin aggradation that counters any effect of regional incision in these areas.

The localization of incision in the Rocky Mountain orogenic plateau isolates the event from sea-level influences farther downstream. Base-level changes due to sea-level variations are unlikely to have strongly impacted only these upstream areas. Base-level changes due to more proximal events may have played a role in driving some incision. Zaprowski et al. (2001) argued that recent incision in the area surrounding the Black Hills resulted from knickpoint retreat caused by a base-level fall to the north and east along the Missouri River system. They suggest that if this process acted over several million years, it could account for the incision to the west of the Black Hills seen in the adjacent Rocky Mountains as well. However, our pattern of incision with magnitudes in excess of 1 km that decay to tens of meters downstream in the Great Plains indicates that incision is not focused in the downstream areas. Thus, this mechanism alone is likely the major control on regional incision.

Base-level changes along the bounding structures between the Colorado Plateau and Basin and Range have also been suggested as the cause of regional incision along the Colorado River drainage. Knickpoints generated from local faults do exhibit some control on incision. Faulting along the Wasatch Front and from the Hurricane and Toroweap faults at the mouth of the Grand Canyon has apparently generated knickzones that have migrated upstream tens of kilometers (Fenton et al., 2001). Pederson et al. (2002) argued that regional incision along the Colorado River system over the past 6 m.y. is due to drainage reversals across the transition between the southwestern Colorado Plateau and Basin and Range from the bounding faults. However, in our reconstruction (Fig. 4C), incision upstream of this margin dies out before, and is separated from, the regional incision seen upstream along the Colorado River drainage.

Incision, Climate, and Tectonics

One goal of this study is to use the incision history to evaluate the roles of climate change and tectonic activity in the post-Laramide landscape development of the Rocky Mountain region. Incision can result from changes in gradient induced by tectonically driven surface uplift or changes in the ratio of sediment to water supply, possibly driven by climate change (Tucker and Slingerland, 1997). Untangling these two causative factors is not simple, in part because they are not mutually exclusive, feedbacks between them may occur, and rarely does either factor leave a record of unambiguous origin. Nonetheless, we can evaluate aspects of each controlling factor and infer their relative importance.

Role of Climate Change

Climate control, in the general case, encompasses all aspects of geologic-meteorologic interactions, including magnitudes, rates and periodicity of temperature, precipitation and wind changes, as well as cascading effects involving vegetation and localization of flow routing, among others. By way of simplification, we focus on how the abundance of water varies relative to available sediment supply (Q_w/Q_s), which dictates much of the basin-scale behavior of aggradation and degradation (e.g., Paola et al., 1992). A decrease in Q_w/Q_s encourages aggradation, whereas an increase in Q_w/Q_s encourages incision, especially as older deposits are remobilized along range fronts. Climate can increase water supply directly by increasing precipitation amount and/or seasonality. Sediment supply can be influenced indirectly by climatic control of weathering rates, vegetation distribution, and storm events (affecting sediment delivery off hillslopes). Reviews of global and regional climate change during late Cenozoic time (Zachos et al., 2001; Flower and Kennett, 1993; Hay et al., 2002) provide a basis for comparing the timing of basin filling and incision with climate change in the Rocky Mountain orogenic plateau.

One proxy indicator of climate change is an excursion in the global marine $\delta^{18}O$ curve at ca. 3 Ma (late Pliocene) that marks a major cooling trend (Flower and Kennett, 1993; Zhang et al., 2001; Moore et al., 1987). This change from a monotonic climate to a strongly oscillating regime with infrequent but intense precipitation events, perhaps with a concomitant change in vegetative cover (Zhang et al., 2001), has been cited to explain the production of relief in the Rocky Mountains during late Cenozoic time (e.g., Molnar and England, 1990; Gregory and Chase, 1994). Incision rates significantly accelerated in association with the onset of the cooling trend that began ca. 3 Ma (Sahagian et al., 2002; Dethier, 2001; Lange et al., 2000; Kirkham and Scott, 2002), coincident with a time of globally increased sediment flux delivered to the major ocean basins at 2–4 Ma (Zhang et al., 2001). However, dating of successive incised volcanic flows (Buffler, 2003; Kunk et al., 2002; Stroud, 1997; Lange et al., 2000) demonstrates that incision in most of the Rockies was well on its way prior to this event.

A second proxy indicator of climate change seen in Asia and North America is the transition from C3 to C4 plants at ca. 6–8 Ma (Quade and Cerling, 1995; Cerling et al., 1997; Cerling et al., 1993). Relative to C3 plants, C4 plants have more growth during the warm, water-stressed season and less cool-season growth, perhaps resulting either from less summer precipitation, colder winters, or both. This transition has been interpreted as a proxy for conditions that favor fluvial downcutting. While this transition is coincident with the onset of widespread incision in the Rockies, it does not appear to be a global event (Fox and Koch, 2003). Instead, it may be a local critical threshold phenomenon during a gradual change in climate conditions apparently associated, at least in the Asian case, with continuing orogenic plateau uplift (Quade and Cerling, 1995).

Role of Local Tectonism

Specific areas of incision occur adjacent to known local tectonic features in the Rocky Mountain orogenic plateau. The Yellowstone hotspot in northwest Wyoming, and the Rio Grande Rift, running north from New Mexico into Colorado have both been active in late Cenozoic time and may be responsible for some incision. Thermally driven surface uplift associated with the Yellowstone hotspot can be seen in the wake of its migration path across southern Idaho (Lowry et al., 2000). Pierce and Morgan (1992) argued that surface uplift associated with Yellowstone extends far east of the park boundary into north-central Wyoming. However, Lowry et al. (2000) argued that surface uplift to the east of Yellowstone is less than 200 m, even within 50 km of the present active center of the hotspot. Regardless, at the time of onset of regional incision, the Yellowstone hotspot was found far to the west of its present position (Pierce and Morgan, 1992), and thus likely had little to do with the early development of incision patterns seen in our results.

Normal faulting associated with the Rio Grande Rift can be traced along the crest of the Rocky Mountains into northern Colorado with decreasing influence as it approaches the Wyoming border (Tweto, 1979; Chapin and Cather, 1994; Naeser et al., 2002; Buffler, 2003). As noted earlier, the zones of most active faulting along the rift and its projection into Wyoming are associated with the regions of least incision and the rift flanks are associated with more incision. In addition, there is a northward decrease in both the magnitude of incision along each major river exiting the Rockies onto the Great Plains and in gradient along each interfluve between these rivers (Fig. 5; Leonard, 2002). These trends are consistent with the northward propagation of the Rio Grande Rift over time (Tweto, 1979; Chapin and Cather, 1994). A misfit exists, however, between the width of incision (>250 km) on either side of the Rio Grande Rift and the characteristic wavelength of thermal surface uplift associated with continental rifting (10–100 km; Wernicke and Axen, 1988). Thus, the Rio Grande Rift may be a surface manifestation of, or superimposed upon, some long-wavelength, possibly dynamic, uplift related to upper-mantle and/or thermal effects (Eaton, 1986, 1987; Heller et al., 2003; Roy et al., 2004).

Role of Epeirogeny

On a subcontinental scale, there is a broad correlation between the overall distribution of incision and apparent mantle velocity anomalies in the Rocky Mountain orogenic plateau. Large velocity anomalies seen in P- and/or S-wave traveltimes imaged independently by Dueker et al. (2001), Frederiksen et al. (2001), and Goes and van der Lee (2002) are centered beneath the southern Rocky Mountains and Yellowstone. The integrated isostatic effect of these anomalies, calculated by Goes and van der Lee (2002), results in a domal pattern of high topography (Fig. 4D) with 1–2 km of predicted relief between the Rocky Mountains and the central Great Plains. This zone of high isostatic topography follows the trend of our reconstructed fill surface (Fig. 4B) and results in a relief difference (1–2 km vertical change over >200 km length) between the Rockies and the Great Plains that matches both the gradient of the western edge of the reconstructed surface and calculations of regional tilting of the western Great Plains over the past ~5 m.y. (Leonard, 2002; McMillan et al., 2002). In addition, this feature is consistent with indications of broad subsurface upward-directed loading based on gravity modeling (Angevine and Flanagan, 1987), crustal thickness estimates from P-wave velocity interpretations (Sheehan et al., 1995), and flexural modeling of the southern flanks of the Rio Grande Rift (Roy et al., 1999). Although the pattern and location of incision (Fig. 4C) generally mimic the distribution of the mantle anomaly beneath the Rocky Mountain region, incision depths are greatest along the flanks of the anomaly and are low over the central part of the anomaly. This may result from the localization of zones of extension, where graben filling counteracts incision, over the center of the anomaly.

Proposed mechanisms that potentially could drive surface uplift of the Rocky Mountain orogenic plateau include isostatic and dynamic processes related to mantle flow, mid-crustal flow, anomalous compositional or thermal changes in the crust and mantle, or rebound related to the fate of the Farallon slab (Dickinson and Snyder, 1979; Morgan and Swanberg, 1985; Severinghaus and Atwater, 1990; Humphreys and Dueker, 1994; Humphreys, 1995; Spencer, 1996; Burgess et al., 1997; McQuarrie and Chase, 2000; Heller et al., 2003). Regardless of the process of origin, if the pattern and timing of incision described here are related to the formation of the velocity anomaly at depth, then our results may constrain when that anomaly began to develop.

Comparison with Other Studies

Other studies in recent years have addressed the relationship between elevation and late Cenozoic incision in the Rocky Mountain region. A compilation by Dethier (2001) of incision rates over approximately the same study area during the past 600 k.y. reveals a concentration of high rates in a very similar pattern to the location of maximum incision magnitudes seen in our results. This suggests that those factors driving long-term incision have been active in Quaternary time and are likely active today.

Within the Colorado Plateau, Pederson et al. (2002) reconstructed topography and calculated rock uplift and erosional exhumation using an Eocene-Oligocene datum. Overall, our net incision results are similar to their erosion map. Specific differences include divergence in the patterns within the Uinta basin and San Juan basin–San Juan Mountains areas and in our overall lower incision magnitudes. We attribute the differences to two factors. First, the lower magnitudes are due to our method of using preserved remnants versus reconvolved stratigraphy and thus returning the minimum reconstructed top and minimum incision magnitudes. Secondly, we use a late Miocene datum compared to their ca. 30 Ma datum. Their reconstruction predates most of the basin filling, and most of the constructional activity in the San Juan Mountains (35–15 Ma). The Uinta and San Juan basins were likely filled after 30 Ma and then re-exhumed with the onset of widespread incision (Pitman et al., 1982; Fassett, 1985). Thus, their net erosion for the Colorado Plateau may have been less over the past 30 m.y. than for the past ~8 m.y.

Although our results are generally consistent with those of Pederson et al. (2002), we draw different conclusions. They concluded that their results are most consistent with a scenario in which early Cenozoic events provided the majority of rock uplift necessary to get the Colorado Plateau to its current elevation, with late Cenozoic events limited to passive isostasy and incision driven by base-level fall along the lower Colorado River drainage. Our results, in contrast, suggest that after the Laramide orogeny, there was a period of slow regional subsidence during Oligocene-Miocene time, which was followed by regional rock uplift and incision over the past ~8 m.y. As such, incision is driven primarily by tectonic forces and is not localized along the Colorado River.

Late Cenozoic rock uplift of the Rocky Mountains seems to run counter to the estimates of paleoelevation based on analyses derived from fossil leaves. These data suggest the mean elevation of the Rockies has not changed relative to sea level since ca. 35 Ma (Gregory and Chase, 1992, 1994; Molnar and England, 1990; Wolfe et al., 1998). Hence there has been no net elevation gain in the Rocky Mountains since the end of Laramide time. These observations have been used to infer that climate change alone is the cause of regional incision of the Rocky Mountains (Zhang et al., 2001) and have been used to place constraints on models of regional tectonic processes (e.g., McQuarrie and Chase, 2000). Therefore, we need to examine the apparent conflict with our results and reevaluate the basis for elevation estimates derived from paleobotanical approaches.

The uncertainty in paleobotanical estimates of elevation is, in our view, large enough that it may not resolve the magnitude of net elevation change in the Rockies. Paleoelevation estimates derived from fossil plants require a series of steps in which errors can accumulate, a few of which we highlight here. The first step is to establish a correlation between the size and shape of dicotyledonous leaves in extant vegetation and the climate in which the plants grow. The strongest correlation is between the percent species with smooth (entire-margined) leaves and mean annual temperature (MAT) observed in a set of East Asian floras ($r^2 = 0.98$, standard error = ±0.8 °C; Wolfe, 1979; Wing and Greenwood, 1993). However, these correlations vary regionally and are less strong in other data sets (e.g., Wilf, 1997; Wolfe, 1993; Jacobs, 1999, 2002; Gregory-Wodzicki, 2000; Kennedy et al., 2002). These regional variations and imprecisions in the correlation of leaf traits with climate lead to uncertainties of 2–3 °C in estimating MAT.

The next step in determining paleoelevation is to estimate paleotemperature for sites of the same age and paleolatitude, one of which is known to have been at sea level at the time (usually a coastal site). In paleoelevation analyses, the differences in temperature between the two sites are assumed to be the result of elevation. If the two sites are precisely the same age, the assumption is justified, but if the sites are of different ages, the differences could be the result of regional or global climate change over time. For Cenozoic floras, it is difficult to establish contemporaneity within less than 1 m.y. If global or regional changes in climate exceed more than a few degrees per million years, this could mistakenly be read as elevational differences of at least 500 m between floras of the same elevation that are thought to be coeval. Other potential sources of uncertainty include the binomial error inherent in assessing the leaf traits of a flora from a finite (usually small) number of species (Wilf, 1997) and possible biases in the leaf characters of the species that are preserved when compared with the original vegetation (Burnham et al., 2001; Kowalski and Dilcher, 2003). Unless sample sizes are very large (e.g., >50 species), binomial error alone will generate uncertainties of ±2–3 °C in MAT estimates. With a lapse rate of 5 °C/km, this translates to an error of at least 500 m of elevation for each flora.

The next step in paleoelevation analysis assumes that differences in paleotemperature between the sea-level site and other sites (usually interior) are attributable to differences in elevation, and present-day lapse rates are typically used to calculate how large an elevational difference will explain the difference in paleoclimate. Modern lapse rates of MAT vary geographically from <4.0 to >8.0 °C, which has the potential to change paleoelevational estimates by a factor of two (Meyer, 1992). An additional step is required if fossil floras are not from similar paleolatitudes, because the temperature estimates have to be projected to the same paleolatitude using latitudinal gradients of MAT for the past, which are also subject to uncertainty (Greenwood and Wing, 1995; Fricke and Wing, 2004). An alternative to the lapse rate approach involves determining mean enthalpy from leaf features (Forest et al., 1999). However, the correlation between leaf features and precipitation is rather imprecise (Wing and Greenwood, 1993; Wilf et al., 1998), which leads to high regression error in estimation of paleoprecipitation, and therefore in mean enthalpy for fossil sites.

In summary, published uncertainties in the paleobotanical paleoelevation approach range from ±400 to ±1500 m (Gregory and Chase, 1992; Wilf et al., 1998). Our compilation of basin fill thickness (Fig. 4A) suggests that there was up to 850 m of regional subsidence across most or all of the Rockies and adjacent parts of the Great Plains with respect to their original position that took place between the end of the Laramide orogeny (ca. 40 Ma) and the beginning of regional incision (ca. 6–8 Ma). Subsequent incision of the same area and the Colorado Plateau was up to 1.6 km (Fig. 4C). In a simplistic view, if post-Laramide surface uplift is coincident with and equal to maximum incision amounts (1000–1600 km), then the Rockies have experienced less than 750 m (at most) net vertical motion since ca. 35 Ma. This value falls within the uncertainty associated with paleobotanic elevation reconstructions and resolves the apparent conflict between them.

CONCLUSION

By analyzing post-Laramide basin fill remnants, we determined the distribution of basin fill thickness, the minimum elevation of the reconstructed fill surface, and the pattern of subsequent minimum incision in the Rocky Mountain orogenic plateau during late Cenozoic time. Our results are most consistent with tectonic factors as the first-order influence on the history of relief. Thick accumulation of basin fill by through-flowing river systems suggests slow regional subsidence of at least the Rocky Mountains after the end of the Laramide orogeny until ca. 6–8 Ma in the center of the Rockies and ca. 3–4 Ma along the fringes of the region. The pattern of reconstructed topography coincides with calculated isostatic topography associated with upper-mantle velocity anomalies where hot mantle is supporting thin crust. The distribution of incision relative to the mantle anomalies follows a pattern of low incision over the maximum predicted topography and deep incision along the flanks. Thus, we suspect regional doming and incision related to the initiation of the upper-mantle anomaly. The timing of turnaround from aggradation to incision is not likely related to the ca. 3 Ma global climate change event, although incision rates certainly increased over the past 3–4 m.y. Our results suggest that, given inherent uncertainties, there is no fundamental disparity between estimates of paleoelevation at the end of the Laramide orogeny and the cumulative late Cenozoic tectonic history of the Rocky Mountain region. Finally, this type of analysis may be useful for providing age constraints on the emplacement of mantle anomalies.

ACKNOWLEDGMENTS

This work was supported by National Science Foundation grant EAR-0106029 to Heller and by Geological Society of America and Colorado Scientific Society student grants to McMillan. We have benefited from discussions with many colleagues, especially P. Beland, R.T. Buffler, D.P. Dethier, K. Dueker, N.F. Humphrey, and faculty and students at the University of Wyoming. Detailed reviews by P. Molnar, J. Stock, C. Chapin, and P. Chamberlain are greatly appreciated. McMillan is indebted to F.J. Pazzaglia and his students at Lehigh University for guidance in applying geographic information systems (GIS) data to landscape evolution.

REFERENCES CITED

Angevine, C.L., and Flanagan, K.M., 1987, Buoyant subsurface loading of the lithosphere in the Great Plains foreland basin: Nature, v. 327, p. 137–139, doi: 10.1038/327137a0.

Axelrod, D.I., and Bailey, H.P., 1976, Tertiary vegetation, climate, and altitude of the Rio Grande depression, New Mexico–Colorado: Paleobiology, v. 2, p. 235–254.

Best, M.G., Christiansen, E.H., Deino, A.L., Gromme, C.S., McKee, E.H., and Noble, D.C., 1989, Excursion 3A: Eocene through Miocene volcanism in the Great Basin of the western United States, *in* Chapin, C.E., and Zidek, J., eds., Field excursions to volcanic terranes in the western United States, Volume II: Cascades and the intermountain West: New Mexico Bureau of Mines and Mineral Resources Memoir 47, p. 91–133.

Bond, W.A., 1984, Application of Lopatin's method to determine burial history, evolution of the geothermal gradient, and timing of hydrocarbon generation in Cretaceous source rocks in the San Juan Basin, northwestern New Mexico and southwestern Colorado, *in* Woodward, J., et al., eds., Hydrocarbon source rocks of the Greater Rocky Mountain region: Denver, Rocky Mountain Association of Geologists, p. 433–447.

Buffler, R.T., 2003, The Browns Park Formation in the Elkhead region, northwestern Colorado–south central Wyoming: Implications for Late Cenozoic sedimentation, *in* Raynolds, R.G., and Flores, R.M., eds., Cenozoic systems of the Rocky Mountain region: Denver, Rocky Mountain Society for Economic Paleontologists and Mineralogists, p. 183–212.

Burgess, P.M., Gurnis, M., and Moresi, L., 1997, Formation of sequences in the cratonic interior of North America by interaction between mantle, eustatic, and stratigraphic processes: Geological Society of America Bulletin, v. 109, no. 12, p. 1515–1535, doi: 10.1130/0016-7606(1997)109<1515:FOSITC>2.3.CO;2.

Burnham, R.J., Pitman, N.C.A., Johnson, K.R., and Wilf, P., 2001, Habitat-related error in estimating temperatures from leaf margins in a humid tropical forest: American Journal of Botany, v. 88, p. 1096–1102.

Cerling, T.E., Yang, W., and Quade, J., 1993, Expansion of C_4 ecosystems as an indicator of global ecological change in the late Miocene: Nature, v. 361, p. 344–354, doi: 10.1038/361344a0.

Cerling, T.E., Harris, J.M., MacFadden, B.J., Leakey, M.G., Quade, J., Eisenmann, V., and Ehleringer, J.R., 1997, Global vegetation change through the Miocene/Pliocene boundary: Nature, v. 389, p. 153–158, doi: 10.1038/38229.

Chapin, C.E., and Cather, S.M., 1994, Tectonic setting of the axial basins of the northern and central Rio Grande Rift, *in* Keller, G.R., and Cather, S.M., eds., Basins of the Rio Grande Rift: Structure, stratigraphy, and tectonic setting: Geological Society of America Special Paper 291, p. 5–25.

Chapin, C.E., and Kelley, S.A., 1997, The Rocky Mountain erosion surface in the Front Range of Colorado, *in* Bolyard, D.W., and Sonnenberg, S.A., eds., Colorado Front Range guidebook: Denver, Rocky Mountain Association of Geologists, p. 101–114.

Cross, T.A., and Pilger, R.H., Jr., 1978, Tectonic controls of Late Cretaceous sedimentation, western interior, USA: Nature, v. 274, p. 653–657, doi: 10.1038/274653a0.

DeCelles, P.G., Gray, M.B., Ridgway, K.D., Cole, R.B., Srivastava, P., Pequera, N., and Pivnik, D.A., 1991, Kinematic history of a foreland uplift from Paleocene synorogenic conglomerate, Beartooth Range, Wyoming and Montana: Geological Society of America Bulletin, v. 103, p. 1458–1475, doi: 10.1130/0016-7606(1991)103<1458:KHOAFU>2.3.CO;2.

Dethier, D.P., 2001, Pleistocene incision rates in the western United States calibrated using Lava Creek B tephra: Geology, v. 29, p. 783–786, doi: 10.1130/0091-7613(2001)029<0783:PIRITW>2.0.CO;2.

Dickinson, W.R., and Snyder, W.S., 1979, Geometry of subducted slabs related to San Andreas transform: Journal of Geology, v. 87, p. 609–627.

Dickinson, W.R., Klute, M.A., Hayes, M.J., Janecke, S.U., Lundin, E.R., McKittrick, M.A., and Olivares, M.D., 1988, Paleogeographic and paleotectonic setting of Laramide sedimentary basins in the central Rocky Mountain region: Geological Society of America Bulletin, v. 100, p. 1023–1039, doi: 10.1130/0016-7606(1988)100<1023:PAPSOL>2.3.CO;2.

Dueker, K., Yuan, H., and Zurek, B., 2001, Thick-structured Proterozoic lithosphere of the Rocky Mountain region: GSA Today, v. 11, no. 12, p. 4–9, doi: 10.1130/1052-5173(2001)011<0004:TSPLOT>2.0.CO;2.

Eaton, G.P., 1986, A tectonic redefinition of the southern Rocky Mountains: Tectonophysics, v. 132, p. 163–193, doi: 10.1016/0040-1951(86)90031-4.

Eaton, G.P., 1987, Topography and origin of the southern Rocky Mountains and Alvarado Ridge, *in* Coward, M.P., et al., eds., Continental extensional tectonics: Geological Society [London] Special Paper 28, p. 355–369.

Epis, R.C., and Chapin, C.E., 1975, Geomorphic and tectonic implications of the post-Laramide late Eocene erosion surface in the southern Rocky Mountains, *in* Curtis, B.F., ed., Cenozoic history of the southern Rocky Mountains: Geological Society of America Memoir 144, p. 45–74.

Erslev, E.A., 1993, Thrusts, backthrusts, and detachment of Rocky Mountain foreland arches, *in* Schmidt, C.J., et al., eds., Laramide basement deformation in the Rocky Mountain foreland of the western United States: Geological Society of America Special Paper 280, p. 339–358.

Evanoff, E., 1990, Early Oligocene paleovalleys in southern and central Wyoming: Evidence of high local relief on the late Eocene unconformity: Geology, v. 18, p. 443–446, doi: 10.1130/0091-7613(1990)018<0443:EOPISA>2.3.CO;2.

Fassett, J.E., 1985, Early Tertiary paleogeography and paleotectonics of the San Juan basin area, NM and CO, *in* Flores, R.M., and Kaplan, S.S., eds., Cenozoic paleogeography of the west-central United States: Rocky Mountain Paleogeography Symposium 3: Society of Economic Paleontologists and Mineralogists, Rocky Mountain Section, p. 317–334.

Fenton, C.R., Webb, R.H., Pearthree, P.A., Cerling, T.E., and Poreda, R.J., 2001, Displacement on the Toroweap and Hurricane faults: Implications for Quaternary downcutting in the Grand Canyon, Arizona: Geology, v. 29, p. 10,035–10,038, doi: 10.1130/0091-7613(2001)029<1035:DROTTA>2.0.CO;2

Flanagan, K.M., 1990, Late Cenozoic geology of the Pathfinder region, central Wyoming, with tectonic implications for adjacent area [Ph.D. thesis]: Laramie, University of Wyoming, 186 p.

Flanagan, K.M., and Montagne, J., 1993, Neogene stratigraphy and tectonics of Wyoming, *in* Snoke, A.W., et al., eds., Geology of Wyoming: Laramie, Geological Survey of Wyoming Memoir 5, p. 572–607.

Flower, B.P., and Kennett, J.P., 1993, Middle Miocene ocean-climate transition: High-resolution oxygen and carbon isotopic records from Deep Sea Drilling Project Site 588A, southwest Pacific: Paleoceanography, v. 8, p. 811–843.

Forest, C.E., Wolfe, J.A., Molnar, P., and Emanuel, K.A., 1999, Paleoaltimetry incorporating atmospheric physics and botanical estimates of paleoclimate: Geological Society of America Bulletin, v. 111, p. 497–511, doi: 10.1130/0016-7606(1999)111<0497:PIAPAB>2.3.CO;2.

Fox, D.L., and Koch, P.L., 2003, Tertiary history of C_4 biomass in the Great Plains, USA: Geology, v. 31, p. 809–812, doi: 10.1130/G19580.1.

Frederiksen, A.W., Bostock, M.G., and Cassidy, J.F., 2001, S-wave velocity structure of the Canadian upper mantle: Physics of the Earth and Planetary Interiors, v. 124, p. 175–191, doi: 10.1016/S0031-9201(01)00194-7.

Fricke, H.C., 2003, Investigation of early Eocene water-vapor transport and paleoelevation using oxygen isotope data from geographically widespread mammal remains: Geological Society of America Bulletin, v. 115, p. 1088–1096, doi: 10.1130/B25249.1.

Fricke, H.C., and Wing, S.L., 2004, Oxygen isotope and paleobotanical estimates of temperature and $\delta^{18}O$-latitude gradients over North America during the early Eocene: American Journal of Science, v. 304, p. 612–635.

Goes, S., and van der Lee, S., 2002, Thermal structure of the North American uppermost mantle inferred from seismic tomography: Journal of Geophysical Research, ser. B, Solid Earth and Planets, v. 107, p. ETG 2-1–ETG 2-13.

Green, G.N., 1992, The Digital Geologic Map of Colorado in ARC/INFO Format: U.S. Geological Survey Open-File Report 92-507, scale 1:500,000.

Green, G.N. and Drouillard, P.H., 1994, The Digital Geologic Map of Wyoming in ARC/INFO Format: U.S. Geological Survey Open-File Report 94-0425, scale 1:500,000.

Green, G. N., and Jones, G.E., 1997, The Digital Geologic Map of New Mexico in ARC/INFO Format: U.S. Geological Survey Open-File Report 97-0052, scale 1:500,000.

Greenwood, D.R., and Wing, S.L., 1995, Eocene continental climates and latitudinal temperature gradients: Geology, v. 23, p. 1044–1048.

Gregory, K.M., and Chase, C.G., 1992, Tectonic significance of paleobotanically estimated climate and altitude of the late Eocene erosion surface, Colorado: Geology, v. 20, p. 581–585, doi: 10.1130/0091-7613(1992)020<0581:TSOPEC>2.3.CO;2.

Gregory, K.M., and Chase, C.G., 1994, Tectonic and climatic significance of a late Eocene low-relief geomorphic surface, Colorado: Journal of Geophysical Research, v. 99, p. 20,141–20,160, doi: 10.1029/94JB00132.

Gregory-Wodzicki, K.M., 2000, Relationships between leaf morphology and climate, Bolivia: Implications for estimating paleoclimate from fossil floras: Paleobiology, v. 26, p. 668–688.

Gross, E.L., Patchett, P.J., Dallegge, T.A., and Spencer, J.E., 2001, The Colorado River and Neogene sedimentary formations along its course: Apparent Sr isotopic connections: Journal of Geology, v. 109, p. 449–461, doi: 10.1086/320793.

Hagen, E.S., and Surdam, R.C., 1984, Maturation history and thermal evolution of Cretaceous source rocks of the Bighorn Basin, Wyoming and Montana, *in* Woodward, J., et al., eds., Hydrocarbon source rocks of the greater Rocky Mountain region: Denver, Rocky Mountain Association of Geologists, p. 321–338.

Hay, W.W., Shaw, C.A., and Wold, C.N., 1989, Mass-balanced paleogeographic reconstructions: Geologische Rundschau, v. 78, p. 207–242, doi: 10.1007/BF01988362.

Hay, W.W., Soeding, E., DeConto, R.M., and Wold, C.N., 2002, The late Cenozoic climate change paradox: International Journal of Earth Sciences, v. 91, p. 746–774, doi: 10.1007/s00531-002-0263-1.

Heller, P.L., Dueker, K., and McMillan, M.E., 2003, Post-Paleozoic alluvial gravel transport as evidence of continental tilting in the U.S. Cordillera: Geological Society of America Bulletin, v. 115, p. 1122–1132, doi: 10.1130/B25219.1.

Hintze, L.F., Willis, G.C., Laes, D.Y.M., Sprinkel, D.A., and Brown, K.D., 2000, Digital geologic map of Utah: Utah Geological Survey Map 179DM, scale 1:500,000.

Hirschberg, D.M., and Pitts, G.S., 2000, Digital geologic map of Arizona: A digital database derived from the 1983 printing of the Wilson, Moore, and Cooper: U.S. Geological Survey Open-File Report 00-409, scale 1:500,000.

Hoy, R.G., and Ridgway, K.D., 1997, Structural and sedimentological development of footwall growth synclines along an intraforeland uplift, east-central Bighorn Mountains, Wyoming: Geological Society of America Bulletin, v. 109, p. 915–935, doi: 10.1130/0016-7606(1997)109<0915:SASDOF>2.3.CO;2.

Humphreys, E.D., 1995, Post-Laramide removal of the Farallon slab, western United States: Geology, v. 23, p. 987–990, doi: 10.1130/0091-7613(1995)023<0987:PLROTF>2.3.CO;2.

Humphreys, E.D., and Dueker, K.G., 1994, Physical state of the western U.S. upper mantle: Journal of Geophysical Research, v. 99, p. 9635–9650, doi: 10.1029/93JB02640.

Hunt, C.B., 1956, Cenozoic geology of the Colorado Plateau: U.S. Geological Survey Professional Paper P0279, 99 p.

Izett, G.A., 1975, Late Cenozoic sedimentation and deformation in northern Colorado and adjoining areas, *in* Curtis, B.F., ed., Cenozoic history of the southern Rocky Mountains: Geological Society of America Memoir 144, p. 179–209.

Jacobs, B.F., 1999, Estimation of rainfall variables from leaf characters in tropical Africa: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 145, p. 231–250, doi: 10.1016/S0031-0182(98)00102-3.

Jacobs, B.F., 2002, Estimation of low-latitude paleoclimates using fossil angiosperm leaves: Examples from the Miocene Tugen Hills, Kenya: Paleobiology, v. 28, p. 399–421.

Jenson, S.K., and Domingue, J.O., 1988, Extracting topographic structure from digital elevation data for geographic information system analysis: Photogrammetric Engineering and Remote Sensing, v. 54, p. 1593–1600.

Kennedy, E.M., Spicer, R.A., and Rees, P.M., 2002, Quantitative palaeoclimate estimates from Late Cretaceous and Paleocene leaf floras in the northwest of the South Island, New Zealand: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 184, p. 321–345, doi: 10.1016/S0031-0182(02)00261-4.

Kirkham, R.M., and Scott, R.B., 2002, Introduction to late Cenozoic evaporite tectonism and volcanism in west-central Colorado, *in* Kirkham, R.M., et al., eds., Late Cenozoic evaporite tectonism and volcanism in west-central Colorado: Geological Society of America Special Paper 366, p. 1–14.

Kowalski, E.A., and Dilcher, D.L., 2003, Warmer paleotemperatures for terrestrial ecosystems: Proceedings of the National Academy of Sciences of the United States of America, v. 100, p. 167–170, doi: 10.1073/pnas.232693599.

Kunk, M.J., Budahn, J.R., Unruh, D.M., Stanley, J.O., Kirkham, R.M., Bryant, B., Scott, R.B., Lidke, D.J., and Streufert, R.K., 2002, ^{40}Ar/^{39}Ar ages of late Cenozoic volcanic rocks within and around the Carbondale and Eagle collapse centers, Colorado; constraints on the timing of evaporite-related collapse and incision of the Colorado River, *in* Kirkham, R.M., et al., eds., Late Cenozoic evaporite tectonism and volcanism in west-central Colorado: Geological Society of America Special Paper 366, p. 213–234.

Lange, R.A., Carmichael, I.S.E., and Hall, C.M., 2000, ^{40}Ar/^{39}Ar chronology of the Leucite Hills, Wyoming: Eruption rates, erosion rates, and an evolving temperature structure of the underlying mantle: Earth and Planetary Science Letters, v. 174, p. 329–340, doi: 10.1016/S0012-821X(99)00267-8.

Larson, E.E., and Evanoff, E., 1998, Tephrostratigraphy and source of the tuffs of the White River sequences, *in* Terry, D.O., LaGarry, H.E., and Hunt, R.M., Jr., eds., Depositional environments, lithostratigraphy, and biostratigraphy of the White River and Arikaree Groups (late Eocene to Early Miocene, North America): Geological Society of America Special Paper 325, p. 1–14.

Larson, E.E., Ozima, M., and Bradley, W.C., 1975, Late Cenozoic volcanism in northwestern Colorado and its implications concerning tectonism and the origin of the Colorado River system, *in* Curtis, B.F., ed., Cenozoic history of the southern Rocky Mountains: Geological Society of America Memoir 144, p. 155–178.

Leat, P.T., Thompson, R.N., Morrison, M.A., Hendry, G.L., and Dickin, A.P., 1991, Alkaline hybrid mafic magmas of the Yampa area, NW Colorado, and their relationship to the Yellowstone mantle plume and lithospheric mantle domains: Contributions to Mineralogy and Petrology, v. 107, p. 310–327, doi: 10.1007/BF00325101.

Leonard, E.M., 2002, Geomorphic and tectonic forcing of late Cenozoic warping of the Colorado piedmont: Geology, v. 30, p. 595–598, doi: 10.1130/0091-7613(2002)030<0595:GATFOL>2.0.CO;2.

Lillegraven, J.A., 1993, Correlation of Paleogene strata across Wyoming—A user's guide, *in* Snoke, A.W., et al., eds., Geology of Wyoming: Laramie, Geological Survey of Wyoming Memoir 5, p. 414–477.

Lillegraven, J.A., and Ostresh, L.M., 1988, Evolution of Wyoming's early Cenozoic topography and drainage patterns: National Geographic Research, v. 4, p. 303–327.

Liu, S., and Nummedal, D., 2004, Late Cretaceous subsidence in Wyoming: Quantifying the dynamic component: Geology, v. 32, p. 397–400, doi: 10.1130/G20318.1.

Love, J.D., 1970, Cenozoic geology of the Granite Mountains area, central Wyoming: U.S. Geological Survey Professional Paper 495-C, 154 p.

Lowry, A.R., Ribe, N.M., and Smith, R.B., 2000, Dynamic elevation of the Cordillera, western United States: Journal of Geophysical Research, v. 105, p. 23,371–23,390, doi: 10.1029/2000JB900182.

Lucchitta, I., 1989, History of the Grand Canyon and of the Colorado River in Arizona, *in* Jenney, J.P., and Reynolds, S.J., eds., Geologic evolution of Arizona: Arizona Geological Society Digest 17, p. 710–715.

McKee, E.D., and McKee, E.H., 1972, Pliocene uplift of the Grand Canyon region/time of drainage adjustment: Geological Society of America Bulletin, v. 83, p. 1923–1932.

McKenna, M.C., and Love, J.D., 1972, High-level strata containing early Miocene mammals in the Bighorn Mountains, Wyoming: American Museum Novitates, no. 2490, 31 p.

McMillan, M.E., Angevine, C.L., and Heller, P.L., 2002, Postdepositional tilt of the Miocene-Pliocene Ogallala Group on the western Great Plains: Evidence of late Cenozoic uplift of the Rocky Mountains: Geology, v. 30, p. 63–66, doi: 10.1130/0091-7613(2002)030<0063:PTOTMP>2.0.CO;2.

McQuarrie, N., and Chase, C.G., 2000, Raising the Colorado Plateau: Geology, v. 28, p. 91–94, doi: 10.1130/0091-7613(2000)028<0091:RTCP>2.3.CO;2.

Mears, B., Jr., 1993, Geomorphic history of Wyoming and high-level erosion surfaces, *in* Snoke, A.W., et al., eds., Geology of Wyoming: Laramie, Geological Survey of Wyoming Memoir 5, p. 608–626.

Mears, B., Jr., 1998, Neogene normal faulting superposed on a Laramide uplift: Medicine Bow Mountains, Sierra Madre, and intervening Saratoga Valley, Wyoming and Colorado: Contributions to Geology, v. 32, p. 181–185.

Meyer, H.W., 1992, Lapse rates and other variables applied to estimating paleoaltitudes from fossil floras: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 99, p. 71–99, doi: 10.1016/0031-0182(92)90008-S.

Mitrovica, J.X., Beaumont, C., and Jarvis, G.T., 1989, Tilting of continental interiors by the dynamical effects of subduction: Tectonics, v. 8, p. 1079–1094.

Molnar, P., and England, P., 1990, Late Cenozoic uplift of mountain ranges and global climate change: Chicken or egg?: Nature, v. 346, p. 29–34, doi: 10.1038/346029a0.

Montagne, J., 1991, Cenozoic history of the Saratoga Valley area, Wyoming and Colorado: University of Wyoming Contributions to Geology, v. 29, p. 13–70.

Montgomery, D.R., 1994, Valley incision and the uplift of mountain peaks: Journal of Geophysical Research, v. 99, p. 13,913–13,921, doi: 10.1029/94JB00122.

Moore, T.C., Loutit, T.S., and Grenlee, S.M., 1987, Estimating short term changes in eustatic sea level: Palaeoceanography, v. 2, p. 625–637.

Morgan, P., and Swanberg, C.A., 1985, On the Cenozoic uplift and tectonic stability of the Colorado Plateau: Journal of Geodynamics, v. 3, p. 39–63, doi: 10.1016/0264-3707(85)90021-3.

Naeser, N.D., 1986, Neogene thermal history of the northern Green River Basin, Wyoming—Evidence from fission-track dating, *in* Gantier, D.L., ed., Roles of organic matter in sediment diagenesis: Society of Economic Paleontologists and Mineralogists Special Publication 38, p. 65–72.

Naeser, N.D., 1992, Miocene cooling in the southwestern Powder River Basin, Wyoming—Preliminary evidence from apatite fission-track analysis: U.S. Geological Survey Bulletin 1917-0, p. 1–17.

Naeser, C.W., Izett, G.A., and Obradovich, J.D., 1980, Fission-track and K-Ar ages of natural glasses: U.S. Geological Survey Bulletin 1489, 31 p.

Naeser, C.W., Bryant, B., Kunk, M.J., Kellogg, K.S., Donelick, R.A., and Perry, W.J., Jr., 2002, Tertiary cooling and tectonic history of the White River uplift, Gore Range, and western Front Range, central Colorado: Evidence from fission-track and ^{39}Ar/^{40}Ar ages, *in* Kirkham, R.M., Scott, R.B., and Judkins, T.W., eds., Late Cenozoic evaporite tectonism and volcanism in west-central Colorado: Geological Society of America Special Paper 366, p. 31–53.

Nelson, S.T., Davidson, J.P., and Sullivan, K.R., 1992, New age determinations of central Colorado Plateau laccoliths, Utah: Recognizing disturbed K-Ar systematics and re-evaluating tectonomagnetic relationships: Geological Society of America Bulletin, v. 104, p. 1547–1560, doi: 10.1130/0016-7606(1992)104<1547:NADOCC>2.3.CO;2.

Nuccio, V.F., and Finn, T.M., 1994, Structural and thermal history of the Paleocene Fort Union Formation, central and eastern Wind River Basin, with emphasis on petroleum potential of the Waltman Shale Member, *in* Flores, R.M., et al., eds., Organics and the Rockies, field guide: Laramie, Wyoming, Geological Survey of Wyoming, p. 53–68.

Paola, C., Parker, G., Seal, R., Sinha, S.K., Southard, J.B., and Wilcock, P.R., 1992, Downstream fining by selective deposition in a laboratory flume: Science, v. 258, p. 1757–1760.

Pederson, J.L., Mackley, R.D., and Eddleman, J.L., 2002, Colorado Plateau uplift and erosion-amounts and causes evaluated with GIS: GSA Today, v. 12, no. 8, p. 4–10, doi: 10.1130/1052-5173(2002)012<0004:CPUAEE>2.0.CO;2.

Pierce, K.L., and Morgan, L.A., 1992, The track of the Yellowstone hot spot; volcanism, faulting, and uplift, *in* Link, P.K., et al., eds., Regional geology of eastern Idaho and western Wyoming: Geological Society of America Memoir 179, p. 1–53.

Pitman, J.K., Fouch, T.D., and Goldhaber, M.B., 1982, Depositional setting and diagenetic evolution of some Tertiary unconventional reservoir rocks, Uinta Basin, Utah: American Association of Petroleum Geologists (AAPG) Bulletin, v. 66, p. 1581–1596.

Poulson, S.R., and John, B.E., 2003, Stable isotope and trace element geochemistry of the basal Bouse Formation carbonate, southwestern United States; implications for the Pliocene uplift history of the Colorado Plateau: Geological Society of America Bulletin, v. 115, p. 434–444, doi: 10.1130/0016-7606(2003)115<0434:SIATEG>2.0.CO;2.

Quade, J., and Cerling, T.E., 1995, Expansion of C$_4$ grasses in the late Miocene of northern Pakistan: Evidence from stable isotopes in paleosols, *in* Badglev, C., and Behrensmeyer, A.K., eds., Special issue long records of continental ecosystems: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 115, p. 91–116.

Raines, G.L., and Johnson, B.R., 1996, Digital representation of the Montana state geologic map in ARC/INFO export format: U.S. Geological Survey Open-File Report 95-691, scale 1:500,000.

Roy, M., Karlstrom, K., Kelley, S., Pazzaglia, F.J., and Cather, S., 1999, Topographic setting of the Rio Grande Rift, New Mexico: Assessing the role of flexural "rift-flank uplift" in the Sandia Mountains: New Mexico Geological Society Guidebook, 50th Field Conference, Albuquerque Geology, p. 167–174.

Roy, M., Kelley, S., Pazzaglia, F.J., Cather, S., and House, M.A., 2004, Middle Tertiary buoyancy modification and its relationship to rock exhumation, cooling, and subsequent extension at the eastern margin of the Colorado Plateau: Geology, v. 32, p. 925–928, doi: 10.1130/G20561.1.

Rye, R.O., Bethke, P.M., Lanphere, M.A., and Steven, T.A., 2000, Neogene geomorphic and climatic evolution of the central San Juan Mountains, Colorado; K/Ar age and stable isotope data on supergene alunite and jarosite from the Creede mining district, *in* Bethke, P.M., and Hay, R.L., eds., Ancient Lake Creede; its volcano-tectonic setting, history of sedimentation, and relation to mineralization in the Creede mining district: Geological Society of America Special Paper 346, p. 95–103.

Schruben, P.G., Arndt, R.E., and Bawiec, W.J., 1994, Geology of the Conterminous United States: A Digital Representation of the 1974 P.B., King and H.M. Beikman Map: U.S. Geological Survey Digital Data Series DDS-11, scale 1:2,500,000.

Sahagian, D., Proussevitch, A., and Carlson, W., 2002, Timing of Colorado Plateau uplift: Initial constraints from vesicular basalt-derived paleoelevations: Geology, v. 30, p. 807–810, doi: 10.1130/0091-7613(2002)030<0807:TOCPUI>2.0.CO;2.

Scott, J.W., and Chamberlain, K., 2002, The Upper Miocene Moonstone Formation of central Wyoming: Linking vertebrate biostratigraphy and U-Pb geochronology with post-Laramide tectonism: Geological Society of America Abstracts with Programs, v. 34, no. 6, p. A-408.

Seeland, D.A., 1985, Oligocene paleogeography of the northern Great Plains and adjacent mountains, *in* Flores, R.M., and Kaplan, S.S., eds., Cenozoic paleogeography of the west-central United States: Rocky Mountain Paleogeography Symposium 3: Denver, Society of Economic Paleontologists and Mineralogists, Rocky Mountain Section, p. 187–205.

Segerstrom, K., and Young, E.J., 1972, General geology of the Hahns Peak and Farwell quadrangles, Routt County, Colorado: U.S. Geological Survey Bulletin B1349, 63 p.

Severinghaus, J., and Atwater, T., 1990, Cenozoic geometry and thermal state of the subducting slabs beneath western North America, *in* Wernicke, B.P., ed., Basin and Range extensional tectonics near the latitude of Las Vegas, Nevada: Geological Society of America Memoir 176, p. 1–22.

Sheehan, A.F., Abers, G.A., Jones, C.H., and Lerner-Lam, A.L., 1995, Crustal thickness variations across the Colorado Rocky Mountains from teleseismic receiver functions: Journal of Geophysical Research, v. 100, p. 20,391–20,404, doi: 10.1029/95JB01966.

Small, E.E., and Anderson, R.S., 1998, Pleistocene relief production in Laramide mountain ranges, western United States: Geology, v. 26, p. 123–126, doi: 10.1130/0091-7613(1998)026<0123:PRPILM>2.3.CO;2.

Small, E.E., Anderson, R.S., Repka, J.L., and Finkel, R., 1997, Erosion rates of alpine bedrock summit surfaces deduced from in situ ^{10}Be and ^{26}Al: Earth and Planetary Science Letters, v. 150, p. 413–425, doi: 10.1016/S0012-821X(97)00092-7.

Snyder, G.L., 1980, Geologic map of the northernmost Park Range and southernmost Sierra Madre, Jackson and Routt counties, Colorado: U.S. Geological Survey Miscellaneous Investigations Series, Report I-1113, scale 1:48,000.

Spencer, J.E., 1996, Uplift of the Colorado Plateau due to lithospheric attenuation during Laramide low-angle subduction: Journal of Geophysical Research, v. 101, p. 13,595–13,609, doi: 10.1029/96JB00818.

Spencer, J.E., and Patchett, P.J., 1997, Sr isotope evidence for a lacustrine origin for the upper Miocene to Pliocene Bouse Formation, lower Colorado River trough, and implications for timing of Colorado Plateau uplift: Geological Society of America Bulletin, v. 109, p. 767–778, doi: 10.1130/0016-7606(1997)109<0767:SIEFAL>2.3.CO;2.

Steven, T.A., Evanoff, E., and Yuhas, R.H., 1997, Middle and late Cenozoic tectonic and geomorphic development of the Front Range of Colorado, *in* Bolyard, D.W., and Sonnenberg, S.A., eds., Geologic history of the Colorado Front Range: Rocky Mountain Association of Geologists, Rocky Mountain Section, and American Association of Petroleum Geologists field trip guidebook: Denver, Rocky Mountain Association of Geologists, p. 115–124.

Stormer, J.C., Jr., 1972, Ages and nature of volcanic activity on the Southern High Plains, New Mexico and Colorado: Geological Society of America Bulletin, v. 83, p. 2443–2448.

Stroud, J.R., 1997, The geochronology of the Raton-Clayton volcanic field, with implications for volcanic history and landscape evolution [M.S. thesis]: Socorro, New Mexico Institute of Mining and Technology, 116 p.

Swinehart, J.B., Souders, V.L., DeGraw, H.M., and Diffendal, R.F., Jr., 1985, Cenozoic paleogeography of western Nebraska, *in* Flores, R.M., and Kaplan, S.S., eds., Cenozoic paleogeography of the west-central United States: Rocky Mountain Paleogeography Symposium 3: Denver, Society of Economic Paleontologists and Mineralogists, Rocky Mountain Section, p. 209–229.

Tedford, R.H., Skinner, M.F., Fields, R.W., Rensberger, J.M., Whistler, D.P., Galusha, T., Taylor, B.E., MacDonald, J.R., and Webb, S.D., 1987, Faunal succession and biochronology of the Arikareean through Hemphillian interval (late Oligocene through earliest Pliocene epochs) in North America, *in* Woodburne, M.O., ed., Cenozoic mammals of North America, geochronology and biostratigraphy: Berkeley, University of California Press, p. 153–210.

Trimble, D.E., 1980, Cenozoic history of the Great Plains contrasted with that of the southern Rocky Mountains: A synthesis: The Mountain Geologist, v. 17, p. 56–69.

Tucker, G.E., and Slingerland, R., 1997, Drainage basin response to climate change: Water Resources Research, v. 33, p. 2031–2047, doi: 10.1029/97WR00409.

Tweto, O., 1979, The Rio Grande Rift system in Colorado, *in* Riecker, R.E., ed., Rio Grande Rift: Tectonics and magmatism: American Geophysical Union, p. 33–56.

U.S. Geological Survey, 1993, Digital elevation models, data user guide, 5: Reston, Virginia, U.S. Geological Survey, p. 1–50.

Wayne, W.J., Aber, J.S., Agard, S.S., Bergantino, R.N., Bluemle, J.P., Coates, D.A., Cooley, M.E., Madole, R.F., Martin, J.E., Mears, B., Jr., Morrison, R.B., and Sutherland, W.M., 1991, Quaternary geology of the northern Great Plains, *in* Quaternary nonglacial geology of the conterminous U.S.: Boulder, Colorado, Geological Society of America, The Geology of North America, v. K-2, p. 441–476.

Wernicke, B., and Axen, G.J., 1988, On the role of isostasy and the evolution of normal fault systems: Geology, v. 16, p. 848–851, doi: 10.1130/0091-7613(1988)016<0848:OTROII>2.3.CO;2.

Wilf, P., 1997, When are leaves good thermometers? A new case for leaf margin analysis: Paleobiology, v. 23, p. 373–390.

Wilf, P., Wing, S.L., Greenwood, D.R., and Greenwood, C.L., 1998, Using fossil leaves as paleoprecipitation indicators; an Eocene example: Geology, v. 26, p. 203–206, doi: 10.1130/0091-7613(1998)026<0203:UFLAPI>2.3.CO;2.

Wing, S.L., and Greenwood, D.R., 1993, Fossils and fossil climate: The case for equable continental interiors in the Eocene, *in* Allen, J.R.L., et al., eds., Palaeoclimates and their modelling with special reference to the Mesozoic Era: Philosophical Transactions of the Royal Society, London, ser. B, Biological Sciences, v. 341, p. 243–252.

Wolfe, J.A., 1979, Temperature parameters of humid to mesic forests of eastern Asia and relation to forests of other regions of the Northern Hemisphere and Australasia: U.S. Geological Survey Professional Paper P1106, 37 p.

Wolfe, J.A., 1993, A method of obtaining climatic parameters from leaf assemblages: U.S. Geological Survey Bulletin 2040, p. 1–71.

Wolfe, J.A., Forest, C.E., and Molnar, P., 1998, Paleobotanical evidence of Eocene and Oligocene paleoaltitudes in midlatitude western North America: Geological Society of America Bulletin, v. 110, p. 664–678, doi: 10.1130/0016-7606(1998)110<0664:PEOEAO>2.3.CO;2.

Zachos, J.C., Shackleton, N.J., Revenaugh, J.S., Palike, H., and Flower, B.P., 2001, Climate response to orbital forcing across the Oligocene-Miocene boundary: Science, v. 292, p. 274–278, doi: 10.1126/science.1058288.

Zaprowski, B.J., Evenson, E.B., Pazzaglia, F.J., and Epstein, J.B., 2001, Knickzone propagation in the Black Hills and northern High Plains: A different perspective on the late Cenozoic exhumation of the Laramide Rocky Mountains: Geology, v. 29, p. 547–550, doi: 10.1130/0091-7613(2001)029<0547:KPITBH>2.0.CO;2.

Zhang, P., Molnar, P., and Downs, W.R., 2001, Increased sedimentation rates and grain sizes 2–4 Myr ago due to the influence of climate change on erosion rates: Nature, v. 410, p. 891–897, doi: 10.1038/35069099.

Manuscript Received by the Society 29 October 2004
Revised Manuscript Received 28 May 2005
Manuscript Accepted 31 August 2005

Printed in the USA