COLORADO PLATEAU AND SOUTHERN ROCKY MOUNTAINS UPLIFT AND EROSION

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Abstract

The Paleoproterozoic lithosphere of the region that includes the Colorado Plateau (CP) and the Southern Rocky Mountains (SRM) does not appear to have had an inherent Proterozoic structural demarcation of the CP and SRM. Following its accretion to the Wyoming craton, it had relative periods of stability and sedimentation, followed by short periods of widespread faulting and disruption of the sedimentary rocks previously deposited. First demarcation of the SRM was during late Paleozoic deformation of the Ancestral Rocky Mountains in which deformation was focused in Texas, Oklahoma, and probably the SRM by plate boundary forces. When more regional stresses of the Cordilleran orogenies compressed the region, the SRM appear to have been reactivated because their lithosphere was weak rather than because the CP lithosphere was strong, and thus the CP and SRM were delineated by events rather than by inherited lithospheric properties. The late Mesozoic and Cenozoic uplift and erosion histories of the CP and SRM are difficult to constrain: Both were submerged or were close to sea level by a Late Cretaceous high sea level stand just prior to Laramide deformation, and we know their present elevations. There is little consensus on elevations between these two end points. The SRM were uplifted during Laramide deformation, but the elevation of a late Eocene erosion surface in the region is less certain. Evidence for a component of late Cenozoic uplift of the SRM is provided by geophysical data that indicate modern crustal and upper mantle anomalies consistent with the cause of recent uplift. Data concerning uplift age(s) for the CP are currently under debate, but xenolith data provide useful constraints on models of uplift. These data preclude any model that requires significant heating or replacement of the mantle lithosphere above a depth of 140 km by mechanisms such as low-angle subducting slabs, delamination, or

asthenospheric heating, prior to ~25 Ma, and any model of complete replacement of the mantle lithosphere beneath the CP prior to as recently as ~1 Ma. Mid-crustal flow models are precluded primarily on calculations of cooling with reasonable CP geotherms and arguments based on uniformitarianism. A new paleo-altimeter, based on basalt vesicle size, indicates about 1 km of uplift during the past 5 Ma, a number consistent with previous geological estimates. The preferred mechanism for uplift, consistent with the xenolith data, and the new basalt paleo-altimeter data, is a phase change from dense eclogite to less-dense garnet granulite in mafic rock close to the crust-mantle boundary. The model is consistent also with the relatively thick crust and gradational Moho measured by seismic experiments in the southwestern CP.

Introduction

The Colorado Plateau (CP) and Southern Rocky Mountains (SRM) are two major adjacent physiographic provinces in the Western United States (Figure 1). They contrast greatly in geology and physiography. The CP is a collection of plateaus of mostly flat lying sediments of Paleozoic and younger age sitting on horizontal and tilted Neoproterozoic sedimentary rocks and a Paleoproterozoic basement. The SRM are rugged mountains, including most of the highest peaks in the conterminous United States, comprising highly deformed sedimentary rocks, igneous intrusions and extrusions, Precambrian inliers, and rooted in the same Paleoproterozoic basement as the CP. Both provinces were last close to or below sea level during a high Late Cretaceous sea level stand, and although major deformation and initial uplift of the Southern Rocky Mountains was certainly Late Cretaceous to early Tertiary in age, evidence has also been interpreted for Neogene uplift and erosion in both provinces.

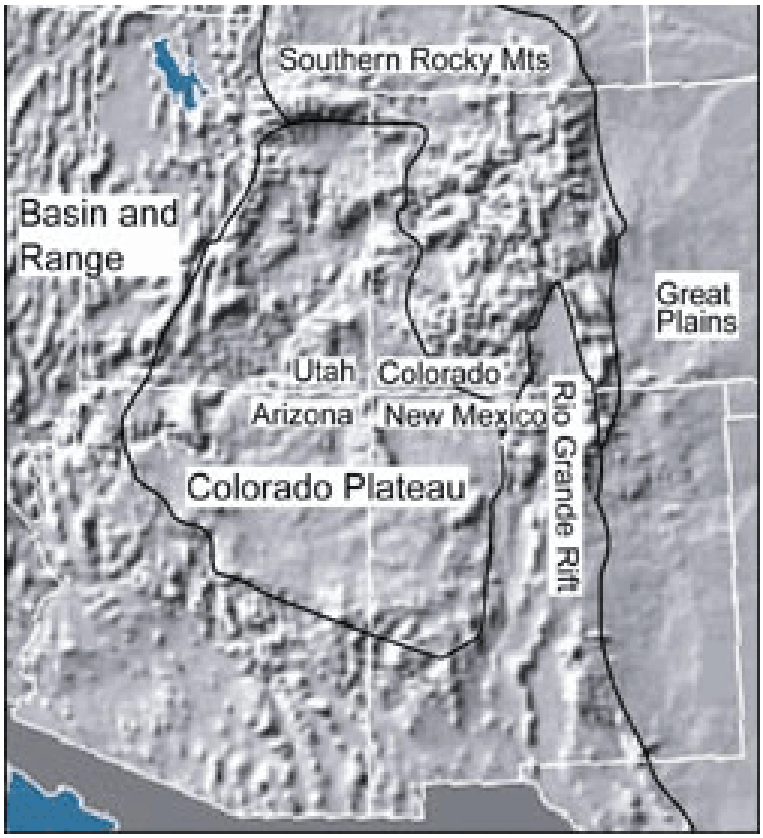


Figure 1. Location map of Colorado Plateau and Southern Rocky Mountains, and adjacent physiographic provinces with shaded relief map as a background. Structurally and tectonically the Southern Rocky Mountains extend south through the Rio Grande Rift into east Texas (Eaton,1986, 1987).

In this contribution an attempt is made to compile relevant data to understand the lithospheric evolutions of the Colorado Plateau and Southern Rocky Mountains from the Paleoproterozoic to the present, with particular emphasis on events relevant to Late Cretaceous though Cenozoic uplift and erosion. Models for Late Cretaceous-Cenozoic uplift are examined and an attempt is made to place constraints on the uplift histories of the Colorado Plateau and the Southern Rocky Mountains both with the available data and through consideration of likely processes responsible for uplift.

Episodic Activity and Stable Roots

The crust/lithosphere of the Colorado Plateau (CP) and Southern Rocky Mountains (SRM) is commonly referred to as cratonic through much of its geologic history. Examination of its geologic record, however, indicates that its longest period of "cratonic" behavior has been about 250 Ma, and that its times of geological quiescence have been interspersed with episodes of deformation, both extensional and compressional. (A period of about 450 Ma of relative quiescence may have existed in the Neoproterozoic further to the east and north). For most it its history, the record documents that its surface has been close to sea level. This record is briefly documented in order to examine its relevance to the Cretaceous-Cenozoic uplift and erosion history of the region.

Basement to the CP and SRM was accreted to the Archean Wyoming Province during the Paleoproterozoic with northeast trending belts of the Mojave (2.2 to 2.0 Ga; reworked Archean), Yavapai(1.8 to 1.7 Ga; juvenile), and Mazatzal (1.8 to 1.65; juvenile; deformed in Mazatzal orogeny 1.65 to 1.6 Ga) provinces (Figure 2; Karlstrom and Bowring, 1993; Van Schmus *et al.*, 1993). These rocks were intruded by 1.4 Ga granitoids and overlain by Middle and Late Proterozoic sedimentary rocks.

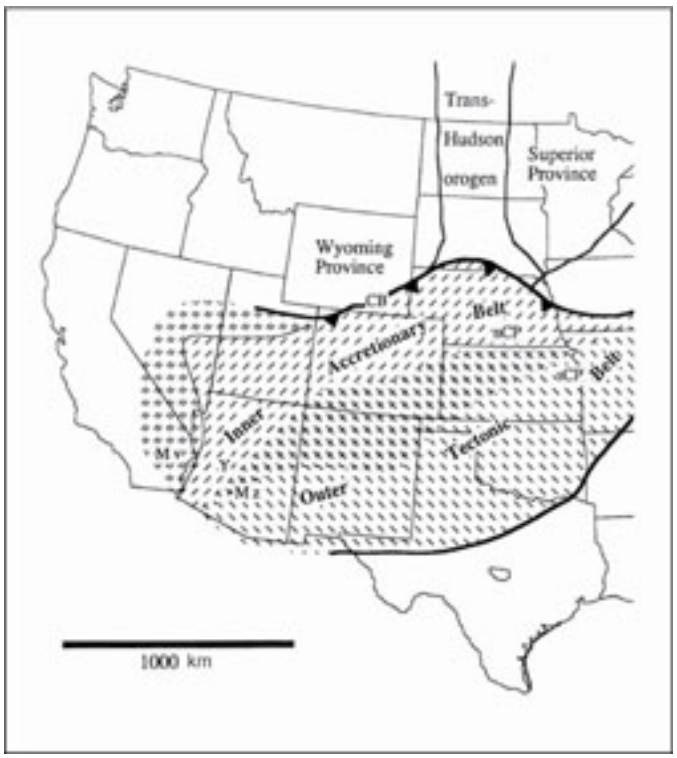


Figure 2. Proposed distribution of Paleoproterozoic crustal provinces that are basement to the Colorado Plateau and Southern Rocky Mountains. CB, Cheyenne belt; Mv, Mojave province; Y, Yavapai province; Mz, Mazatzal province; and nCP and sCP, northern and southern Central Plains orogens, respectively. Heavy line marking the southeastern edge of the Outer Tectonic belt is the inferred continental margin at ~1.6 to ~1.5 Ga and could have resulted from rifting, or could have been the outed limits of ~1.6 Ga continental growth. [modified from Van Schmus et al. (1993)].

These sediments are known as the Grand Canyon Supergroup in the Grand Canyon of Northern Arizona, (GCS), total 3,623 m in thickness, and range in age from ~1,250 to ~800 Ma (Elston, 1993). They rest nonconformably (the "Greatest Unconformity") on the Paleoproterozoic basement of the Vishnu Schist, and are overlain by an angular unconformity (the "Great Unconformity"). A number of disconformities divide the GCS, but all of its formations were deposited in shallow marine or coastal subaerial environments. These depositional environments indicate that at the site of deposition of the GCS, the newly accreted (Paleoproterozoic) crust of the proto-CP was eroded to sea level prior to the deposition of these sediments. Timmons *et al.* (2001) interpret extensional tectonics in these sediments from exposures in the Grand Canyon, suggesting that they were deposited in extensional basins formed during multistage extensional events from ~1.1 Ga to ~0.8Ga.

Older sedimentary rocks are also found in central and southern Arizona, probably resulting from tectonic activity associated with the episode of 1.4 Ga granitoid emplacement. A quartzite, the Troy Quartzite, is tentatively correlated with the lowest group of the GCS (Wrucke, 1993), indicating that the CGS, and the conditions that it represents, were laterally extensive.

Thick (>3,500 to >7,000 m) sections of sedimentary rocks and greenschist facies metamorphosed sedimentary rocks of Middle(?) and early Late Proterozoic age have been documented in the Uinta Mountains and Wasatch Mountains of northern Utah (Uinta Mountain Group and Big Cottonwood Fm., Link, 1993). These sections are dominated by quartz-rich rocks, and are interpreted to have been deposited primarily in shallow water. Based on paleomagnetic, paleontologic, and limited radiometric evidence, these sections most probably correlate with the Chuar Group, the upper group of the GCS (Table 1; Elston *et al*, 1993).

Table 1. Correlation of Late Proterozoic strata of Northern Utah and Southeastern Idaho with Northern Arizona. (Modified from Link et al., 1993, Figure 5)

Age Ma	Northern Utah and Southeastern Idaho		Northern Arizona
800		--?--------?---	----------
850	~~~~~~~~~~~~ Big Cottonwood Fm.	Uinta Mountain Group	Chuar Group
900	⌣⌣?⌣⌣⌣?⌣⌣⌣	~~~~~~~~~~	~~~~~~~~~~~~~~

Key: ⌣⌣ = thrust contact; ~~ = unconformity; ? = uncertain

The preserved sections of Middle and early-Late Proterozoic sedimentary rock in the CP and marginal regions indicate that, following Paleoproterozoic crustal accretion, this region was eroded to sea-level, then slowly subsided to receive shallow marine or subaerial sediments of a continental margin facies until around 800 Ma. No sedimentary rocks of this appropriate age are preserved in the SRM: uplift, tilting, and block faulting are inferred to have begun at about 800 Ma by Elston and Link (1993; see also Link et al., 1993).

Proterozoic sediment deposition in this region was brought to a halt by major crustal extension possibly associated with continental rifting to the west, and inland Basin and Range type half-graben extension across much of the CP region (Ford and Breed, 1973; Elston and McKee, 1982; Sears, 1990). The timing of this event (~800 Ma) suggests that it was associated with the breakup of the supercontinent Rodinia (Dalziel, 1997). Proterozoic sections were preserved in tilted down-dropped fault-block half-graben, and, following extension, the region was again eroded, setting the stage for a period of passive sedimentation.

The region that was to become the CP and SRM had been eroded to a broad gentle ridge, the Transcontinental Arch by the latest Proterozoic, extending generally northeast from New Mexico to Minnesota and the Lake Superior Region. Marine transgressions occurred during high sea-level stands from passive margins a few hundred kilometers to the west and southeast of the CP. Sedimentation and erosion across the region from the latest Proterozoic through the Mississippian Period were controlled by the Transcontinental Arch as epeiric seas transgressed and regressed in response to global eustatic sea level changes (see papers in Fouch and Magathan, 1980; Longman and Sonnenfeld, 1996). During the Late Devonian and Early Mississippian the sediment source to the west changed as an island arc collided with the passive margin to the west (the Antler orogeny, *e.g.*, Beus, 1980).

"Cratonic" behavior of the CP-SRM lithosphere was curtailed in the Pennsylvanian with the development of the Ancestral Rocky Mountains and a number of other uplifts and basins developed in response to the development of the Ouachita Mobile Belt to the southeast (Miller *et al. 1992*a; Blakey, 1980). This deformation was the second time that a long period of sedimentation from epeiric seas was disrupted by tectonism, the first being the end of the Grand Canyon Supergroup by extension. There were large displacements (>2000 m) of high-angle reverse faults with additional strike-slip movements during Ancestral Rocky Mountain deformation and Miller *et al.* (1992a) suggest that these were probably reactivation of Precambrian normal faults. The uplifted blocks provided new sources for clastic sediments and enclosed basin sites for evaporite deposition. Access to open ocean to the east and southeast was cut off. Minor uplift and subsidence possibly continued into the Early Permian controlling erosion and deposition, but as the Permian progressed these topographic features became subdued, and by Late Permian many of the older structural highs and basins were unrecognizable. Further erosion and infilling of remnant Ancestral Rocky Mountain topography occurred during deposition of Triassic sediments, dominated by continental deposits, and to the west these deposits show a return to tectonic stability with transgressive-regressive sequences controlled by global eustatic sea level changes (Dubiel, 1994; Paull and Paull, 1994). Continental deposition continued to dominate the region in the Early Jurassic, but marine conditions returned in the Middle Jurassic when the opening of the Gulf of Mexico allowed flooding from the southeast by the Sundance Sea, and more influence of global eustatic sea level changes. This stratigraphic pattern is evidence of another period of stability following Ancestral

Rocky Mountain deformation.

This period of stability ended during a series of western seaboard plate margin events resulting in three orogenies at the end of the Mesozoic, the Nevadan, the Sevier, and the Laramide orogenies, collectively known as the Cordilleran orogeny. The Nevadan orogeny, defined by Wilmarth (1938) as Late Jurassic and earliest Cretaceous was accompanied by relatively minor thrusting to the west (Allmendinger, 1992). The term Nevadan orogeny in the California region was later restricted by Bateman and Clark's (1974) definition of a distinct, short-lived Late Jurassic deformation event, but geochronological and structural studies of the Sierra Nevada and Idaho batholiths and associated volcanics, and the western Klamath Mountains, demonstrate that Nevadan deformation extended into the Early Cretaceous time (Evernden and Kistler, 1970; Tobish *et al.*, 1989; Stern *et al.*, 1981; Harper et al., 1991; Saleeby and Busbe-Spera, 1992). The CP was tilted westward during early and middle Jurassic time, and several significant positive and negative structures were also active. Many of the structures on the CP that were strongly activated during the Laramide have now been shown to have had a complex pre-Laramide tectonic history, including activation during the Nevadan orogeny, and some of these movements are recorded in the sediments of the Morrison and Cedar Mountain Formations (Peterson, 1986; Currie, 1998). The topography on and off the CP formed sources for sediments for the CP.

The Sevier orogeny was defined by Armstrong (1968) as occurring approximately between the beginning of the Cretaceous and the Campanian (~145 to 75 Ma), and formed numerous overlapping thrusts mostly to the northwest of the CP. Although there is no evidence that the Sevier orogeny caused significant internal deformation in either the CP or the SRM, the Sevier orogeny is roughly coincident with the end of control of the sedimentary record in this region by global eustatic sea-level changes, but caused significant tilting of the region.

The eastern edge of Lower Cretaceous deformation extended from the southwestern corner of Utah, up through central Utah, north into extreme eastern Idaho and western Montana, forming a deep (3-5.5 km) north-south basin in central Utah and along the Idaho-Wyoming state line (the Utah-Idaho trough), and a slightly shallower (3-5 km) basin in south-central Wyoming. From roughly the 1.5 to 2 km contours of these basins a broad foreland basin shallowed to the east, feathering out to zero about 1,000 km east of the eastern edge of the deformation (Cross and Pilger, 1978a, b; Mitrovica *et al.*, 1989). These basins were filled with marine sediments, and the end of marine conditions over most of the region was the Late Cretaceous deposition of the Mancos/Pierre Shales and their equivalents (Elder and Kirkland, 1994). This deposition was during a very high sea level stand (Figure 3; Dickinson, 1992). Most of Arizona was missed in this tilting and deposition event, and only marginally affected by Cretaceous marine deposition. However, there is little evidence of significant topography in the Arizona CP at this time, and this is the last indication of the CP and SRM being at or close to sea level, although this sea level may have been some 300 m above present sea level (Figure 3).

The final pulse of the Cordilleran orogeny, the Laramide orogeny, occurred during the Late Cretaceous to Early Tertiary, and developed east and southeast of Sevier deformation. Most of the structural features of the modern Rocky Mountains developed during this orogeny, and it also caused monoclinal deformation throughout the CP (Figure 4). Significant crustal shortening within the SRM is clear in Laramide deformation, but CP deformation is minor. CP monoclines have been linked by many authors to deeper thrusts and reactivated basement faults

(Davis, 1978; Huntoon, 1981; Chapin and Cather, 1981, 1983; Miller *et al.*, 1992b).

Igneous activity accompanied Laramide deformation, and this activity also extends southwest from central Colorado into the Four Corners area of the CP (the common point of the states of Utah, Colorado, New Mexico and Arizona) and into southern Utah along a diffuse trend

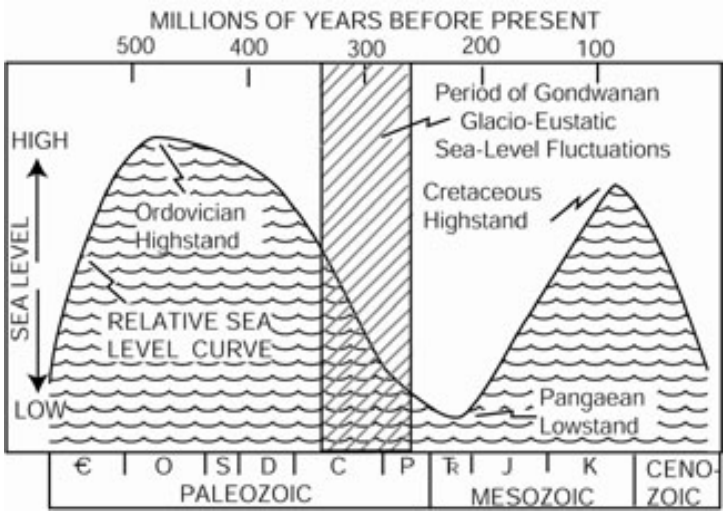


Figure 3. Smoothed global eustatic sea level changes during the Phanerozoic. [From Dickinson, 1992].

Figure 4. Paleogeographic map of the CP and SRM region of latest Cretaceous to middle Eocene time. Laramide thrust blocks (patterned where Precambrian basement involved, except in the Uinta Mountains, U, where only Precambrian sedimentary rocks are involved), thrusts, and monoclines are detailed. Laramide foredeep deposits are shown in dark grey (large areas). Low-grade metamorphic rocks are shown with oblique grey lines. Intrusive rocks are shown in black, and the Colorado Mineral Belt (CMB) is outlined in black dots. MFTB indicates the Maria fold and thrust belt. [Modified from Burchfiel *et al.*, 1992].

known as the Colorado Mineral Belt (Figure 4; Burchfiel *et al.*, 1992). The CP and SRM have never again been flooded by marine transgressions following Laramide deformation, but sea level has remained below the high stand that deposited the Mancos Shale, and how much of the modern elevations of these regions is due to Laramide tectonics, and how much is due to later events is unclear. Local topography in the SRM was subdued by erosion at the end of Laramide deformation leaving a Late Eocene erosion surface of low relief extending through south-central Colorado, which provides a post-Laramide, pre-Oligocene, regional structural datum (Epis and Chapin, 1975). This surface was originally thought to have developed at relatively low elevation on the basis of paleoflora data (<900 m; Epis and Chapin, 1975), but more recent paleobotanical studies place its development close to its modern elevation (Gregory and Chase, 1992).

Laramide compression was followed in the CP/SRM region by arc-type volcanism and extension, general patterns of which are shown from Late Eocene to 17 Ma in Figure 5. This activity was extensive in the SRM, but was minor in the CP, being mainly in the form of magmatic activity on the plateau margins. Extension was typically oriented WSW-ENE and much of it was accommodated along low angle faults, typical of metamorphic core complexes, many of which developed in the Southwestern US during this period (Crittenden *et al.* 1980; Morgan *et al.*, 1886; Burchfiel *et al.*, 1992). After 17 Ma both the style and orientation, and, to some degree, the location of deformation changed, and the foci of magmatic activity changed (Figure 6; papers in Lucchitta and Morgan, 1990; Burchfiel *et al.*, 1992). The stress field rotated to a more E-W extensional field, and extension was generally accommodated along more steeply-dipping normal faults producing the high relief basins and ranges of the modern Basin and Range and Rio Grande rift (Morgan *et al.*, 1986). Extension was limited to the south of the CP in the southern Basin and Range (Christiansen and Yeats, 1992). Magmatic activity was again extensive in the southern margin of the CP and along the SRM (Christiansen and Yeats, 1992; Lipman, 1992), and this magmatic activity continues into historical times (Sunset Crater, Arizona, 1066 A.D.). Current seismicity testifies to a low, but significant level of modern tectonic activity within and around the CP and SRM.

Post-Jurassic Uplift and Erosional History of the Colorado Plateau and Southern Rocky Mountains

Xenoliths. A constraint that is used in this and subsequent sections is data from xenolith samples, specifically xenoliths from the lithospheric mantle. These are samples of the uppermost mantle that are entrained in ascending magmas, typically basalts or magmas of deeper origin, such as kimberlites. From their measured densities (typically ~3.3 Mg m^{-3}) they are significantly more dense than the magmas that entrain them, and most estimates of their ascent times from the mantle are of the order of one day. Their *in situ* temperatures in the mantle are high, and from the reaction rims on their outer margins, they are typically altered very little during their ascent to the surface. Apart from being a lithologic sample of mantle at the time that they are carried to the surface, they can often be dated, to give an age for the mantle lithosphere beneath an area, yielding information on sub-crustal tectonics. If they have suitable mineralogies, by studying the partitioning of specific elements between minerals, they can also yield their temperature, and sometimes pressure (depth) of origin. These parameters can constrain lithospheric thickness, and the timing of heating events with respect to the eruption

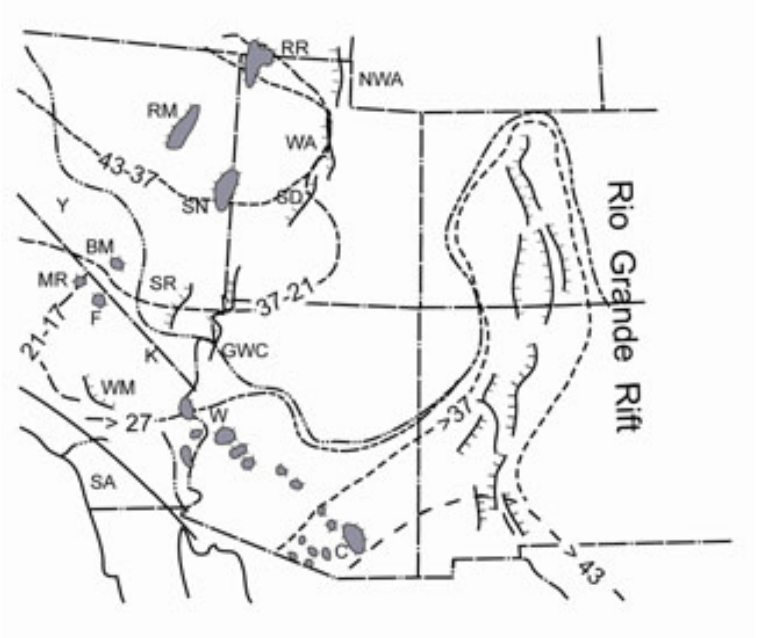


Figure 5. General patterns of arc-type volcanism and extension in the CP/SRM region from the Late Eocene to 17 Ma. Locations of major normal fault zones are shown by short sinuous lines with ticks on the hanging wall side. Metamorphic core complexes are indicated by grey pattern. Extent of volcanic rocks from west and south at different times in Ma are indicated by the sinuous broken lines. BM, Bare Mountain; C, Catalina Mountains; F, Funeral Mountains; GWC, Grand Wash Cliffs fault zone; K, Kingston Range; MR, Mineral Ridge; NWA, northern Wasatch fault; RM, Ruby Mountains; RR, Raft River/Albion Ranges; SA, Santa Rosa fault zone; SD, Sevier Desert fault zone; SN, Snake Range; SR, Sheep Range; W, Whipple Mountains; WA, Wasatch fault; WM, Western Mojave; Y, Yerrington. [Modified from Burchfiel et al., 1992].

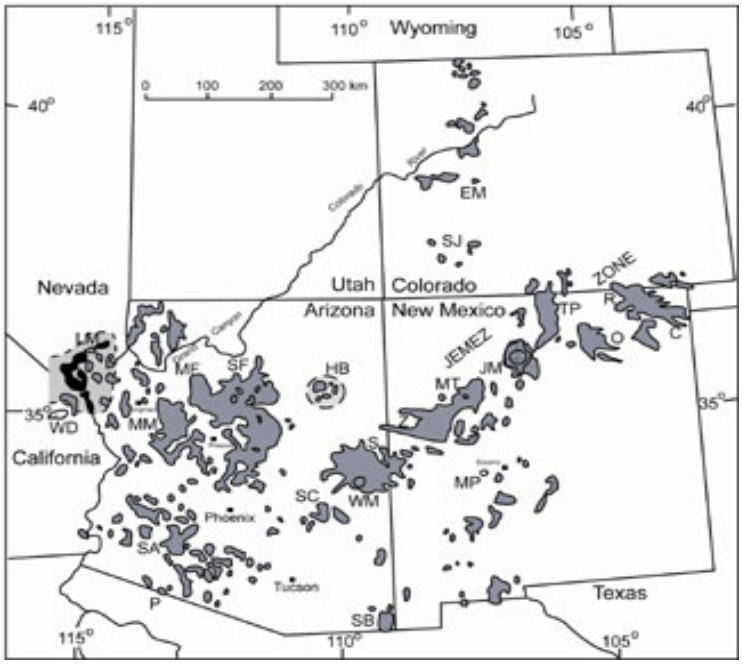


Figure 6. Distribution of igneous rocks in the CP/SRM and Rio Grande rift and Mojave regions younger than 17 Ma. Dark grey areas indicate approximate outcrop distributions; volcanism in southern Nevada is probably volcanic arc; light grey pattern indicates mainly rhyolitic bimodal volcanic suite or local intermediate centers, circle indicates caldera; other areas basaltic or manly basaltic bimodal volcanic suite. C, Clayton; EM, Elk Mountains; HB, Hopi Buttes (dashed line encloses volcanic field); JM, Jemez Mountains; LM, Lake Mead; MF, Mount Floyd; MM, Mahon Mountains; MP Magdalena Peak; MT, Mount Taylor; O, Ocate volcanic field; R, Raton; S, Springerville; SA, Sentinel Plain-Arlington volcanic field; SB San Bernardino (also known as Geronimo) volcanic field; SC San Carlos volcanic field; SF, San Francisco volcanic field; SJ San Juan Mountains; TP, Taos Plateau; WD Woods Mountains; WM, Woods Mountains; Z, Zuni. [From Christiansen and Yeats, 1992].

times of the magmas that entrain the xenoliths. These properties and deductions are used as constraints in the following discussions

Post-Jurassic uplift and erosion were briefly discussed in the previous section in similar detail to earlier uplift or non-depositional periods for the region. However, for is most recent uplift event we can add significant additional details in an attempt to constrain the uplift and erosion history of the plateau associated with this event.

There are two fundamental points that we may place on the uplift curve, the point when the region was last at sea level (a time), and where the region stands today (an elevation). As discussed above, the last time that the region recorded a sedimentary event that responded to the global eustatic sea level curve was during the Jurassic to Early Cretaceous, a time of average mean sea level (Figure 3). Based on the non-deposition of marine deposits despite a global rise in sea level, the region appears to have been uplifted during the time of the Sevier orogeny. Increases in sea level of at least 135 m have been deduced from the phytoplankton response in epicontinental settings (Prauss, 2001), and, as this is probably a minimum estimate of the amount that sea level may rise, we may assume that the region was no more than a few hundred meters above mean sea level when it was flooded for the deposition of the Mancos/Pierre Shales and their equivalents immediately prior to Laramide deformation. Therefore, the first point on the uplift history is mean sea level at the beginning of the Cretaceous, ~145 Ma. Further, we may estimate that the region was no more than a few hundred meters above mean sea level immediately prior to Laramide deformation, ~85 Ma.

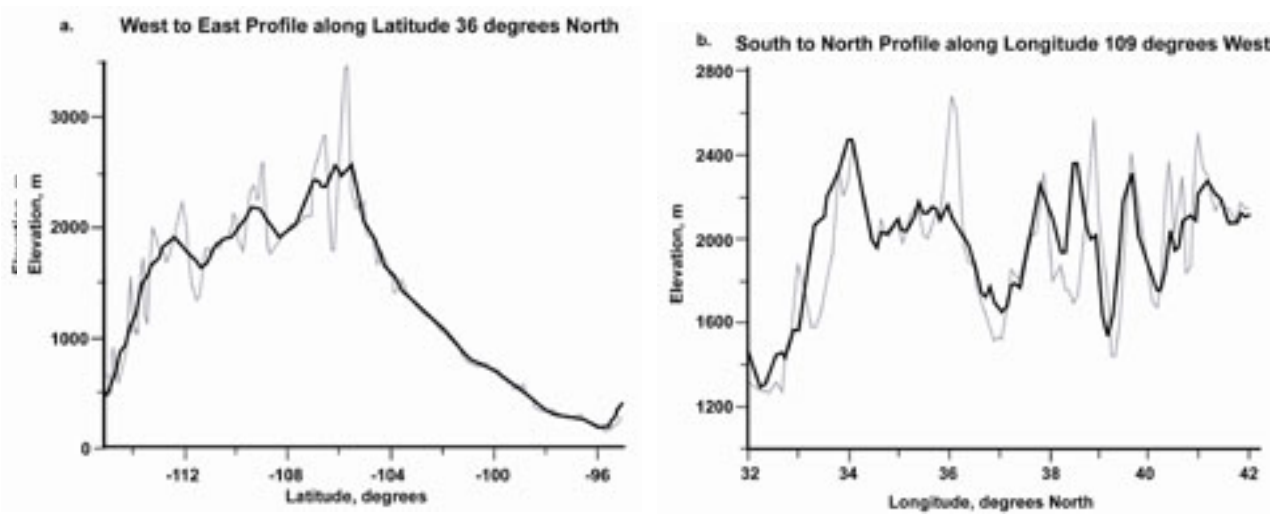


Figure 7. **a.** West to east elevation profiles along latitude 36° North across the CP and SRM. **b**. South to north elevation profiles along longitude 109° West across the CP. Heavy lines show elevations for 5 minute increments averaged half a degree either side of the profiles. Light lines show elevations for 5 minute increments for 5 minute areas along the profiles. [Data from U. S. Geological Survey through EROS Data Center.]

The second firm point on the elevation history is current elevation. Figure 7 shows profiles through the CP and SRM and indicate that the modern elevation of the CP is about 2,000 m and

that the peaks and valleys of the SRM average to an elevation of 2,500 to 3,000 m.

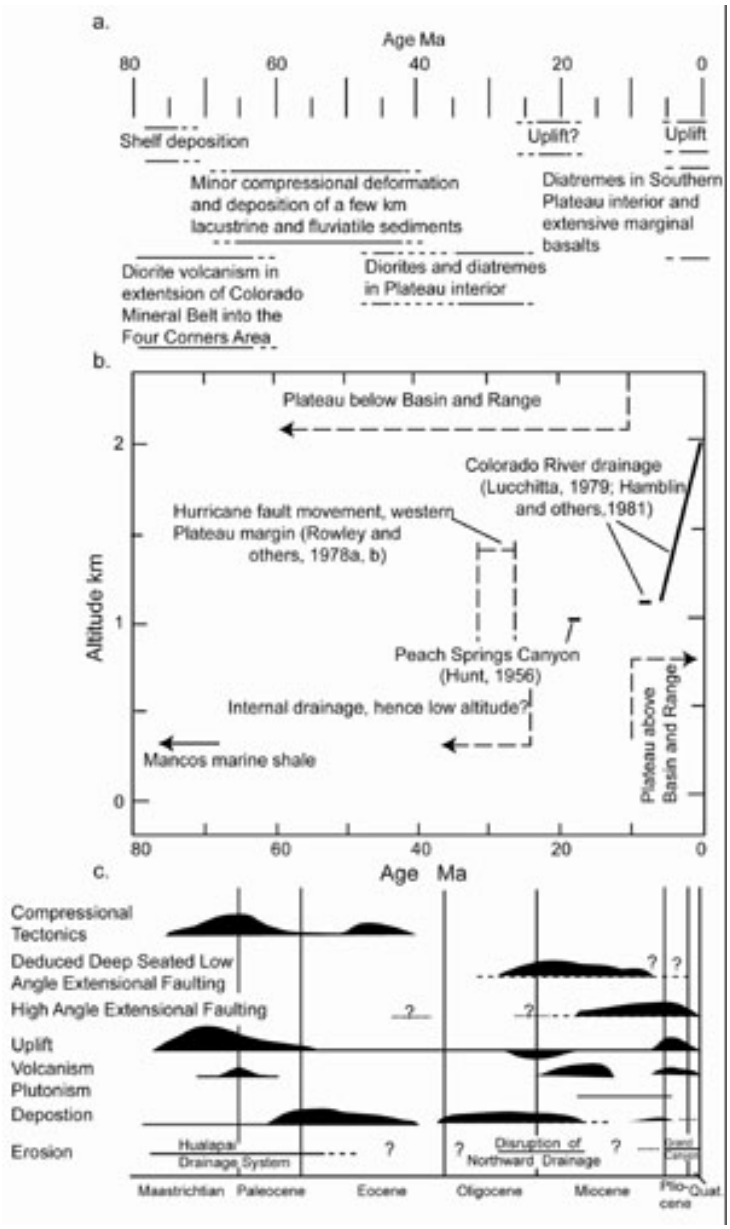


Figure 8. Reconstructions of the geologic history of the Colorado Plateau from Late Cretaceous to the present. **a**. Main tectonic and volcanic events, and **b**. Compilation of uplift and altitude data. [After Morgan and Swanberg, 1985]. **c**. Temporal relationships among tectonics, sedimentation, and erosion in the Grand Canyon region, Arizona. [After Huntoon, 1990].

Other elevation points and significant geological events relevant to uplift history have been compiled by Morgan and Swanberg (1985) and Huntoon (1990), and are summarized in Figure 8. Huntoon (1990) suggests two periods of uplift, a Laramide event associated with Laramide compressional tectonics, and very late Miocene to Recent event, temporally correlated with volcanism and plutonism and at the waning of extensional faulting. Morgan and Swanberg found documentation only for this second event, although they acknowledge that this was the second of two events, the timing of the first being unknown.

Late Cenozoic Uplift. Evidence for late Cenozoic uplift of the CP comes from interpretation of the Miocene Bouse Formation to the west of the Plateau being of deltaic origin and having been uplifted with the CP since its deposition (Lucchitta, 1979), and from strain studies along the western margins of the CP (Hamblin *et al.,* 1981). More recent studies of strontium isotopes in the Bouse Formation have been interpreted to indicate that it is of lacustrine origin and may have been deposited at its present elevation, removing the Bouse Formation evidence for young uplift of the CP (Spencer and Patchett, 1997). The strain studies measure only relative motion: thus, the evidence for late Cenozoic uplift of the CP must be examined more closely.

The strontium isotope signature (initial $^{87}Sr/^{86}Sr$) measured in the Bouse Formation is very close to the signature of the modern Colorado River and significantly different from the Late Miocene seawater strontium isotopic ratio, and this Bouse strontium signature is thought to be primary, not a feature of diagenesis or later alteration (Spencer and Patchett, 1997). Sedimentological studies with limited paleontological evidence, however, interpret the Bouse Formation as marine deltaic (Metzger, 1968; Buising, 1990, 1993; Buising and Beratan, 1993), and reconstructions of movement of the San Andreas fault system link it with the Imperial Valley to the west. Paleontological evidence from the Imperial Valley is very strong for deposition at intertidal to outer neritic depths (McDougall *et al*, 1999) and similar preliminary studies of the Bouse Formation are indicating a similar marine environment (K. McDougall, personal communication, 2001). Thus, there is a conflict among the interpretations of what appear to be solid data relating to the depositional environment of the Bouse Formation.

An attempt has been made to reconcile these data with a deltaic depositional environment for the Bouse Formation by Lucchitta *et al*. (in press) in recognizing that the delta in which the Bouse would have been deposited would have had very limited access to open ocean waters as it was deposited during the formation of the Gulf of California when the Sea of Cortes was little more than an elongate "Basin and Range" province. High strontium ratio waters could have been provided by radiogenic Proterozoic basement granites that are know to exist beneath and around the western CP, and evaporation could have concentrated these waters giving a continental strontium signature to marine deltaic waters. Conversely, Spencer and Patchett (1997) have suggested that marine fauna could have been transported into freshwater lakes by birds, a modern documented process, although perhaps not for all the types of fauna found in the Bouse (K. McDougall, personal communication, 2001).

New data have recently been added to the debate in the form of a new paleo-altimeter based on the size of vesicles in the chill margins at the tops and bottoms of thick (3-4 m) distal, basaltic lava flows (Sahagian and Maus, 1994; Sahagian and Proussevitch, 2002a). These vesicle sizes respond to air pressure, and hence to elevation, and the method has been tested by using the technique to estimate the known solidification elevations of modern lava flows from different elevations on Hawaii. The test results have all been within the experimental errors of $\pm$ 300 m. By using the technique to estimate the solidification elevations of ancient basalts, and using radiometric techniques to date the lavas, the elevation history of an area can be determined if the eruption of the lavas provide suitable samples throughout any moderately large elevation changes in the area. The CP is a suitable area for application of this technique, and preliminary data from the CP are shown in Figure 9.

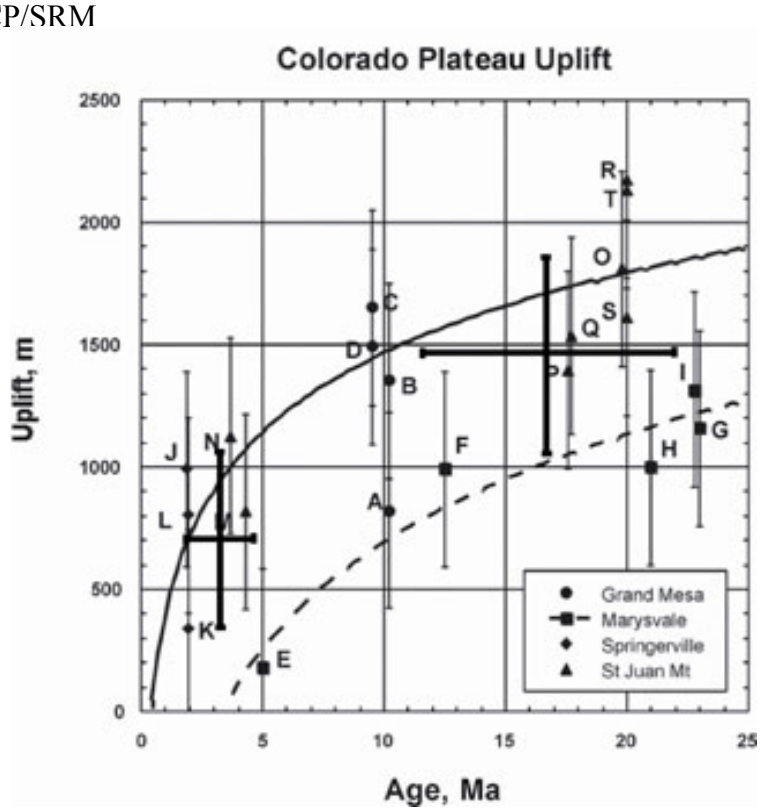


Figure 9. Uplift history of the Colorado Plateau based on vesicular basalt paleo-altimetry. For each sample plotted on the diagram, the uplift is the indicated uplift for that sample since the time of formation of that sample (Age). The solid curve is a best fit logarithmic curve to the data with the highest r-value for the curve relative to other curve fits. This curve indicates slow uplift of 40 m Ma^{-1} at 25 Ma, increasing to 220 m Ma^{-1} for the past 5 Ma. The dashed line is a fit to the Marysvale samples only, which were collected from down-faulted blocks of the transition zone to the Basin and Range. The two heavy symbols with error bars for both Age and Uplift are means and standard deviations for the subsets of data with ages less than 7 Ma and greater than 7 Ma. [Modified from Sahagian *et al.,* 2002].

Two observations are clear from these preliminary data. 1) The data are internally consistent. Average uplift from sites older than ~5 Ma is significantly larger than average uplift from sites younger than ~5 Ma. Scatter in the data is to be expected from known experimental error in the technique, and because all points on the CP are not predicted to be uplifted equally from variations in modern elevation of ancient marine horizons, such as the Permian Kaibab Limestone which now caps the plateau at elevations ranging from elevations ranging from <1,500 m to >2,500 m in Arizona. 2) An uplift of the order of 1 km is indicated by these data since about 5 Ma. The timing and magnitude of this uplift is consistent with interpretations of the Bouse formation as a marine estuarine deposit, and that the strontium isotopes in this formation are strongly controlled by the unusual restricted circulation conditions under which the Bouse was deposited. The evidence for at least 900 m of Late Cenozoic uplift of the CP is therefore considered to be very strong.

Early Cenozoic Uplift. Crustal thickening by compressional tectonics was the major mechanism of uplift of the SRM in the Early Cenozoic (Laramide, see *Uplift by Physical Crustal Thickening* below). This deformation was primarily restricted to the SRM with minor deformation in the CP, and mechanisms for restricting the deformation to the SRM are discussed below. Huntoon (1990) proposed a phase of very Late Cretaceous-early Tertiary uplift associated with Laramide compressional tectonics for the CP (Figure 8). In the CP, although monoclines offset the upper strata, Laramide deformation cannot account for more than 5-10% crustal shortening, which would result in minor crustal thickening and isostatic uplift.

Mechanisms have been proposed by which the CP was somehow underthrust by oceanic lithosphere associated with subduction under the western margin of the North American Plate resulting in cryptic crustal thickening (thickening out of proportion to visible crustal strain; *e.g.,* Bird, 1988). However, isotopic ratios in mantle xenoliths erupted from beneath the CP suggest that the CP is underlain by a Proterozoic mantle root, not one replaced by underthrusting and thickening by younger mantle from the west (Liviccari and Perry, 1993; Riter and Smith, 1993, 1996; Alibert, 1994). Laramide compression in the SRM of a thick-skinned tectonic style (Miller et al., 1992) must have thickened the crust and would have resulted in some isostatic uplift (see *Uplift Through Physical Crustal Thickening* below) However, significant crustal shortening and uplift of the CP cannot be documented. Although it is commonly assumed that there was significant uplift of the CP associated with Laramide tectonism, there is no direct evidence to support either high or low elevation for the CP following Laramide compression. The data do, however, preclude significant crustal thickening by compressional shortening or crustal underthrusting. If the CP was at an elevation of about 300 m following basin sedimentation and prior to Laramide deformation, with a generous interpretation of the crustal shortening, a final elevation in excess of 600 m <u>by</u> <u>this</u> <u>mechanism</u> is unlikely (see *Uplift Through Physical Crustal Thickening* below).

An alternative mechanism that has been proposed for uplift associated with Late Cretaceous-Early Tertiary tectonics is associated with rebound associated with subduction. Many calculations have been published to suggest that because subducting slabs are negatively buoyant and sink, they should have a tendency to pull down the overriding plate by viscous coupling through the mantle wedge between the subducting plate and the plate above (*e.g*., Tovish *et al.*, 1978; Gurnis, 1992; Burgess *et al*., 1997). The deep basins east of the Lower Cretaceous edge of deformation, and the west-tilted foreland basin in western North America at this time have been cited as evidence for this viscous coupling (*e.g.,* Mitrovica *et al*., 1989). When subduction ceased or its geometry changed so that the viscous coupling in a region was no longer effective in the region, the (former) overriding plate was predicted to rebound, resulting in uplift, and this was proposed as a mechanism for early Cenozoic uplift of the CP (Gurnis, 1992).

If there is significant viscous coupling between a subducting slab and its overriding plate causing depression of the overriding plate, such coupling and depression should be visible in modern subduction zones. Modern subduction zones fall into two basic categories, extensional in which a back-arc basin is formed, as in the Mariana trench, the Tonga trench, and the Lesser Antilles island arc system, and compressional in which large earthquakes occur and no back-arc basin is formed, such as the Chile trench. In the extensional systems, there is depression in the overriding plate in the back-arc basins, but these are the result of extension, not of viscous coupling with the sinking slab. In the compressional systems, there is no clear modern evidence of depression, for example, of the overriding Andean plate; gravity data (a strong positive free air gravity anomaly; Grow and Bowin, 1975) suggest a mass excess beneath the overriding plate, and this mass excess appears to reside in the sinking subducting slab. If the slab were viscously coupled to the buoyant overriding plate, depression of this buoyant plate would result in a negative free air gravity anomaly would cancel much of the positive anomaly from the sinking slab. Thus, the strong positive free air gravity anomaly and lack of evidence for surface depression indicate that viscous coupling between the subducting slab and the overriding plate is insignificant. Billen and Gurnis (2001) suggest that the viscous coupling may be much less than

previously modeled because there may be a high concentration of water in the region above a subducting slab that could decrease the viscosity of the mantle locally by several orders of magnitude.

If the CP-SRM region was not depressed by viscous coupling to a subducting slab, some explanation for the thick, eastward-thinning deposits of Cretaceous deposits in the region prior to Laramide deformation is still required. Prior to these deposits, sediment thicknesses were controlled primarily by relatively minor upwarps and downwarps, and proximity to the shoreline. Thicknesses of Cretaceous sedimentary rocks were clearly controlled by the structure of a major sedimentary basin, however. The asymmetric form of this basin and the proximity of its deepest portions to the Lower Cretaceous edge of deformation on the west of the basin are strongly suggestive of a foreland basin resulting from downward flexure of a free end of a section of lithosphere (Turcotte and Schubert, 1982, p. 131). The basin is loaded on its western margin and the flexure reduces to zero to the east. The basin has been modeled by this mechanism by Jordan (1981), and the deformation is amplified by loading of the infilling sediments. If temperatures were sufficiently low at the crust-mantle boundary, the flexural downwarp of the crust could be stabilized by the increased pressure experienced by mafic material in this region converting it from the garnet granulite phase to eclogite. There is no evidence that the region was significantly above the level of the high Cretaceous sea-level stand prior to Laramide deformation.

Middle Cenozoic Uplift. Middle Cenozoic uplift of the CP has been suggested, based on secondary geological features likely to accompany reasonably mechanisms for uplift. Bird (1979) and Bodell and Chapman (1982) suggested that the primary mechanism for uplift of the CP was lithospheric thinning resulting in increased buoyancy of the plateau lithosphere. Bird (1979) proposed delamination of the mantle lithosphere from the crust at about 30 Ma with a second delamination event at about 5 Ma, the two delamination events resulting in uplift and the eruption of mantle-derived magmas in the form of the Navajo Diatremes and the Hopi Buttes, respectively. Bodell and Chapman (1982) modeled a non-specified thinning of the lithosphere from 120 to 80 km starting at 30 Ma, coinciding with the eruption of the Navajo Diatremes. Once again, evidence from mantle xenoliths argues against these hypotheses. Delamination would replace the sub-crustal mantle with young mantle and is inconsistent with the Proterozoic mantle isotopic signature of the xenoliths. The thinning model, as proposed by Bodell and Chapman (1982), and a similar model proposed by Morgan and Swanberg (1985), suggests that as the lithosphere thins from 120 to 80 km it is heated: these models are inconsistent with xenolith geothermometry and geobarometry data that indicate cool mantle temperatures to a depth of 140 km consequent to igneous events at ~25 Ma (Riter and Smith, 1996). Riter and Smith (1996) argue that the temperature record from xenoliths is consistent with a low elevation of the CP until uplift consequent to igneous events at ~25 Ma. There is evidence for relative movement of the western margin of the plateau with respect to the adjacent Basin and Range province along the Hurricane fault, but no direct evidence for absolute uplift or for the absolute elevation of the CP at this time. Thus, in the middle Cenozoic, there are data indicating a cool Precambrian root to the CP, but little constraint on its absolute elevation.

Erosion of the CP and SRM. Across much of the CP in Arizona its surface is capped by Permian Kaibab Limestone, with the occasional outcrops of the overlying Triassic Moenkopi Formation preserved. In general, the Mesozoic section is missing, except in the Black Mesa

area, where it is fairly complete, and elsewhere over much of northeastern Arizona where it is buried (Peterson, 1986). This section is exposed to the north in Utah in the Zion Canyon Area (Triassic-Jurassic) and in the Bryce Canyon and Cedar Breaks Area (Cretaceous, also Eocene and Miocene). Similarly the Mesozoic section is present in varying amounts to the east in New Mexico and to the northeast in Colorado. The exact amount of erosion of Mesozoic strata in Arizona is difficult to extrapolate as depositional conditions indicate thinning of beds to the southeast (*e.g.*, Lawton, 1994). The section is approximately 2,100 m thick in southern Utah, and perhaps a conservative estimate may be that approximately 1,000 m of Mesozoic sedimentary rocks have been eroded from the CP in Arizona. As more of the Mesozoic section remains in other areas of the CP, regional erosion in other areas of the CP is probably less.

Epis and Chapin (1975) presented evidence for a post-Laramide, pre-Oligocene, late Eocene erosion surface in the SRM that extended laterally into adjacent provinces. They postulate that this surface was of low relief, and was formed at low elevation, less than 900 m, on the basis of floral evidence and regional environmental trends. Gregory and Chase (1992), however, using paleobiological data, have argued that the surface was formed at essentially its present elevation of 2,300 to 3,800 m. There is therefore no dispute that an erosion surface was formed in the SRM following Laramide tectonics, but interpretations of the elevation of this surface range from close to sea level to a few thousand meters.

Eaton (1986, 1987) compiled data for east-west profiles across the CP, SRM, and Great Plains, and remarked on the general modern shape of these profiles. He deduced a late Pliocene to Recent age for uplift based on the deposition and dissection age of the late Miocene to late Pliocene Ogallala Formation on the Great Plains, assuming that the Ogallala was deposited near sea level. However, Smith (1940) and many subsequent workers have interpreted that the Ogallala Formation was deposited initially in stream channels as a warped and dissected Piedmont alluvial plain deposit, gradually building up, and with the addition of eolian deposits (Hawley, 1984). The post-Late Pliocene uplift of the CP-SRM-Great Plains system deduced by Eaton is therefore not required.

Final quantitative evidence for erosion rates in the SRM comes from studies of exposure ages of summit flats based on the concentrations of cosmogenic radionuclides in these surfaces (Small and Anderson, 1998). These data indicate a modern erosion rate of these summit flats of ~10 m/Ma and that erosion rates in the intervening valleys are an order of magnitude faster because of different rock types. As mean valley erosion depth from a smooth surface is 280 to 340 m, Small and Anderson (1993) estimate that relief production began at about 3 Ma, and they hypothesize that it was climatically driven and associated with the onset or enhancement of alpine glaciation in these ranges. This climate change coincides with global climate change so there is good temporal correlation for cause and effect. However, if this dissection represents the first major incision of the Eocene erosion surface, and that erosion surface has remained at high elevations since is pre-Oligocene formation, we are left with the conclusion that the Eocene erosion surface was at roughly its present elevation for about 30 Ma without significant erosion and dissection. The CP had internal drainage and lacustrine deposition for much of this time (e.g., Cather and Johnson, 1986), so this interpretation may be reasonable. Alternatively, the Eocene erosion surface was at lower elevations, and pre-3 Ma uplift contributed to the modern erosional regime by lowering erosional base level.

Stability of the Colorado Plateau-Southern Rocky Mountains Lithosphere Through Time

Many discussions of the CP-SRM refer to the "cratonic" behavior of this region with respect to extended periods of its sedimentary history in the Late Proterozoic, Paleozoic, and Mesozoic. These extended periods of stability were punctuated with episodes of mild to moderate deformation, however. Apparent stability or deformation of the region appears to relate more to the regional tectonic setting of "western North America" through time than to any temporally changing inherent properties of the lithosphere of this region. However, that does not preclude spatial heterogeneities in lithospheric properties. The use of the term "craton" is well understood by workers in this region, but in a more global perspective, the region has been relatively active in comparison with, for example, the West African Craton. The proximity of open ocean, the general history of sedimentation, and the relatively minor tectonic episodes suggest that the region would be more appropriately classified as platform rather than "cratonic."

The first major Phanerozoic deformation of the CP-SRM lithosphere was the late Paleozoic formation of the Ancestral Rocky Mountains, major elements of which are shown in Figure 10. As discussed earlier, the location of these mountains is thought to be linked to the formation of the Ouachita Mountains along the southeastern boundary of this region, as was recognized by Burchfiel (1979), Kluth and Coney (1981), and Kluth (1986). The lack of major concurrent deformation in other areas of the CP may be explained if the Ouachita orogeny was the primary source of stresses. Thus, the evidence suggests that the locations of the Ancestral Rocky Mountains were controlled not by existing lithospheric structure but by stress concentration from the Ouichita orogeny.

There is no obvious concentration of stresses responsible for the localization of deformation in formation of the SRM. Jones at el. (1998) have suggested that gravitational potential energy and past orogenesis were closely linked to the localization of Basin and Range and Laramide deformation. They demonstrate that high topography from the Lower Cretaceous deformation to the west of the future CP should have extensional body forces associated with its high gravitational potential energy, and that the deep Cretaceous sedimentary basins in the CP and SRM region should have compressional body forces associated with their gravitational potential energy. There is no question that these forces exist if there are topographic changes and significant lateral density variations (Crough, 1983). There is clearly a spatial relation between the Lower Cretaceous deformation to the west of the region and significantly later Basin and Range extension. However, the deepest Cretaceous basins, Central Utah northward to the Idaho-Wyoming state line, and south-central Wyoming (Cross and Pilger, 1978a, b; Mitrovica *et al.*, 1989), do not show a good spatial relation with the SRM. This mechanism is therefore not considered to be a good candidate for localization of Laramide SRM deformation.

The close spatial association of the SRM to the Ancestral Rocky Mountains is recognized in the naming of the Ancestral Rocky Mountains. Structural uplift in the Ancestral Rocky Mountains, exceeded 6 km in places, and indicates significant local thickening of the crust and

Figure 10. Paleogeographic and tectonic map of western North America of late Paleozoic time showing Cordilleran elements, intracratonic Ancestral Rocky Mountains, and Ouachita orogenic belt. Rock units: ophiolites, black; miogeoclinal rocks, light gray; offshelf rocks, gray; arc-volcanic rocks, red; CA, Chilliwack arc (red); CC, Cache Creek assemblage (pink), ELAF, eastern limit of Antler foredeep; EP, El Paso Mountains; H, Hermosillo; IN, Independence Mountains; NA, Nicola arc (red); NC, northern Cascade Mountains; OB, Oquirrh basin; OG, Osgood Mountains; PR, Peninsular Ranges; SB, San Bernardina Mountains; SJ, San Juan Islands; SM, Slide Mountain; UU Uinta uplift; UB, Uinta basin; WB, Williston basin; WISZ, future location of West Idaho suture zone (gray); W-SD, Wallowa-Seven Devils arc (red). ADD Transcontinental Arch, Defiance Uplift, Uncompahgre Uplift, Tooele Arch and Uinta Uplift from Poole *et al.* To Burchfiel *et al.* Figure. [Modified from Burchfiel *et al.*, 1992 and Poole *et al.*, 1992].

modification of the lithosphere. There is no apparent stress concentration that appears to have caused Laramide deformation to have focused deformation in the SRM. What caused the SRM to be formed in the same region as the Ancestral Rocky Mountains? As there is no evidence of a stress concentration, the cause must be a lithospheric weakness.

Obvious magmatic activity associated with Ancestral Rocky Mountain development is absent: no volcanic or hot spring deposits are known from that synorogenic section. However, there is good evidence for crustal thickening, and an analogy was been drawn between the formation of the Ancestral Rocky Mountains and the deformation in Asia in response to the Cenozoic collision with India (Kluth and Coney, 1981). One of the primary similarities in this analogy is that crustal thickening developed in both areas developed in essentially non-magmatic

(non-arc) settings. Although arc-type igneous rocks are absent in the Himalaya, however, syn-tectonic granites intrude the mountains (Molnar and Tapponier, 1975). The primary source of these granites is thought to be melting in the lower portion of the over-thickened crust as the crust reheats following the depression of the geotherm during thickening. Presumably a similar process would have occurred during the formation of the Ancestral Rocky Mountains, but where is the evidence of this granitic plutonic activity. If the source areas for the granites were relatively rich in water, producing granite magmas with more than a few percent water in the melts, the solidus for these granites has the unusual property that its gradient is negative with respect to depth, *i.e.*, the melting temperature decreases with depth (Tuttle and Bowen, 1958). This property gives the result that ascending "wet" granitic magmas inevitably solidify upon ascent and are unlikely to result in volcanism or shallow plutonism. Thus, by analogy with the Cenozoic India-Asia collision, as suggested by Kluth and Coney (1981), crustal magmatism may be expected in the Ancestral Rocky Mountains, but if the resulting magmas were of "wet" granitic composition, they are unlikely to have produced any shallow plutonism or volcanism.

The speculation of cryptic magmatism in the Ancestral Rocky Mountains suggests that at the end of their formation, their lithosphere was hot relative to the surrounding lithosphere. This lithosphere would then be expected to thermally relax as the heat at depth cooled to the surface. The time for this cooling would depend on the thickness of the lithosphere and depth of the thermal anomalies, but for a 100-km thick lithosphere the thermal relaxation time would be of the order of 70 Ma, and for 200 km the relaxation time would be of the order of 250 Ma (Lachenbruch and Sass, 1977; Morgan and Sass, 1984). The topography of the Ancestral Rocky Mountains would also be expected to be eroded, however, and the isostatic response to erosion is for the lithosphere to rise, advecting heat toward the surface, prolonging the thermal relaxation time for any existing thermal anomalies, and producing a thermal anomaly even if one was not present before erosion. For relatively slow erosion rates Morgan (1983a) has demonstrated mathematically that erosion can prolong thermal relaxation of 100, or even 150 Ma. Thus, it is plausible that temperatures in the Ancestral Rocky Mountain lithosphere could have remained thermally anomalous with respect to surrounding lithosphere for up to 300 Ma after the end of Ancestral Rocky Mountain deformation.

If a residual lithospheric thermal anomaly from Ancestral Rocky Mountain deformation made a zone of lithosphere more susceptible to deformation at the time of SRM deformation than other cooler lithosphere, then the combination of an east-west compressional stress field together with relatively strong lithosphere in Utah and Arizona could have caused deformation to be concentrated where a favorably-oriented north-south zone of relatively weak lithosphere existed in central Colorado and New Mexico. Once deformation started in this zone, it would enhance the residual thermal anomaly, further increasing the contrast in lithospheric strengths, and further focusing deformation the SRM zone. Minor deformation occurred in the cooler lithosphere through which the stresses were transmitted, in the form of the monoclines of what is now the Colorado Plateau. However, this lithosphere probably remained relatively thermally unperturbed, as indicated by the xenolith evidence (see below; Riter and Smith, 1996).

Post-Laramide behavior of the CP, SRM, and adjacent areas with respect to volcanism and deformation may also be explained in terms of prior thermal weaknesses in the lithosphere. All margins of the CP, including the SRM experienced some form of pre-Cenozoic deformation or igneous activity (the southwestern margin was the location of a Late Cretaceous and early Tertiary magmatic arc, Miller *et al.*, 1992) and presumably had thermally perturbed lithospheres relative to the CP. Increasing temperatures in the lithosphere (increasing the geotherm) is one of

the major factors in decreasing lithospheric strength (Lynch and Morgan, 1987). We may consider this strength both on a regional scale, in terms of lithospheric deformation, and on a local scale, in terms of magma ascent. Cenozoic regional stresses have resulted in strain in the weaker lithosphere adjacent to the CP. Volcanism is more prevalent in the weaker lithosphere. Observations show a general association of greater magmatic activity with greater deformation.

Models of Uplift

Dating back to the mid-eighteenth century, isostasy is one of the oldest geophysical principles with direct application to geology and one of the most enduring. It is the basis for all models of uplift of the CP and SRM. The two main hypotheses to isostasy were both put forward in 1855. Pratt's hypothesis assumes that the density within the shell of the earth above the depth of compensation varies laterally in inverse proportion to the elevation of the overlying topography. Airy's hypothesis assumes that a variable-thickness, low-density layer overlies a higher-density substratum, and that the upper layer varies in thickness in proportion to the elevation of the overlying topography. The variable-density shell in Pratt's hypothesis and the variable-thickness, low-density upper layer in Airy's hypothesis are commonly equated with the crust, but when these hypotheses were introduced 1855, the discovery of the Mohorovičić discontinuity (1909) and the formal definition of the crust as a chemical boundary layer had not been made. Equating the constant-density upper layer to the crust in Airy isostasy is a reasonable approximation as crustal density variations are generally secondary to crustal thickness variations. However, lateral density variations are generally more significant in the lithosphere/asthenosphere system in which some form of Pratt isostasy is applicable.

For convenience, both Airy and Pratt isostasy may be combined in a reformulation of isostasy into columns of equal mass, as shown in Figure 11. For continental topography we need consider only columns that extend above sea level, *i.e.*, column a in Figure 11. For the column to be in isostatic equilibrium the mass of the lithospheric column must equal the mass of asthenosphere that it displaces:

$$\rho_l L = (L - H)\rho_a \qquad (1a)$$

or

$$\rho_a H = (\rho_a - \rho_l)L \qquad (1b)$$

Letting ε represent elevation above sea level, we can write:

$$\varepsilon = H - H_O \qquad \varepsilon > 0 \qquad (2)$$

where H is the buoyant height of the continental surface, H_O is the buoyant height of sea level, both measured from the free asthenosphere surface (see column b, Figure 11), and *l* and *a* are subscripts for the lithosphere and asthenopshere, respectively. All symbols are given in Table 2. Combining equations 1 and 2, surface elevation ε can be expressed as:

$$\varepsilon = [(\rho_a - \rho_l)/\rho_a]L - H_O \qquad \varepsilon > 0 \qquad (3)$$

Using a modified form of equation 3 for $\varepsilon < 0$, Lachenbruch and Morgan (1990) calculated a value of ~2.4 km for H_O (column b, $\varepsilon = 0$), assuming the following parameters: $\rho_a = 3.2$ Mg m^{-3}; $\rho_l = 2.8$ Mg m^{-3}; $\rho_w = 1.0$ Mg m^{-3}; and $\varepsilon = -2.5$ km for $L = 5.5$ km. Figure 12 shows the crustal and mantle lithosphere contributions to surface elevation, ε_C and H_m, respectively, for different crustal and mantle lithosphere thicknesses and densities based on these initial calculations and equation 3.

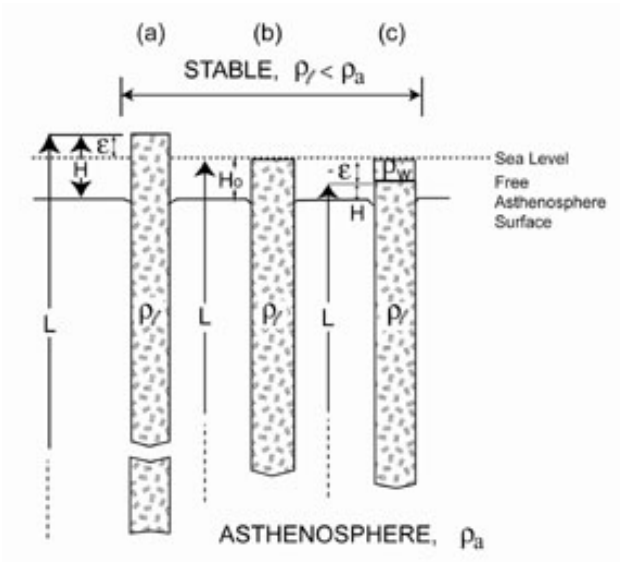


Figure 11. Mass balance relations and notation for lithospheric columns of mean density ρ_l and length L suspended in asthenosphere of density ρ_a with a free (unloaded) surface. H_o is the height of sea level above that surface. Density of sea water is ρ_w. See text and Table 2 for other symbols. [Modified from Lachenbruch and Morgan, 1990].

Table 2. Definitions of symbols used in Equations 1 though 7, and Figure 11.

Symbol	Definition
ρ_l	Density of the lithosphere
ρ_a	Density of the asthenosphere
ρ_c	Density of the crust
ρ_w	Density of water
L	Thickness of the lithospheric column
H	Buoyant height of the continental surface above the free surface of the asthenosphere*
H_o	Buoyant height of sea level above the free surface of the asthenosphere*
ε	Elevation above sea level
δc	Change in crustal thickness required to produce a predetermined change in elevation
δl	Change in thickness of the mantle lithosphere required to produce a predetermined change in elevation
β	Deformation factor defined as ratio of deformed length to undeformed length

* Free surface of the asthenosphere is the hypothetical level to which the asthenosphere would rise if there were no crust.

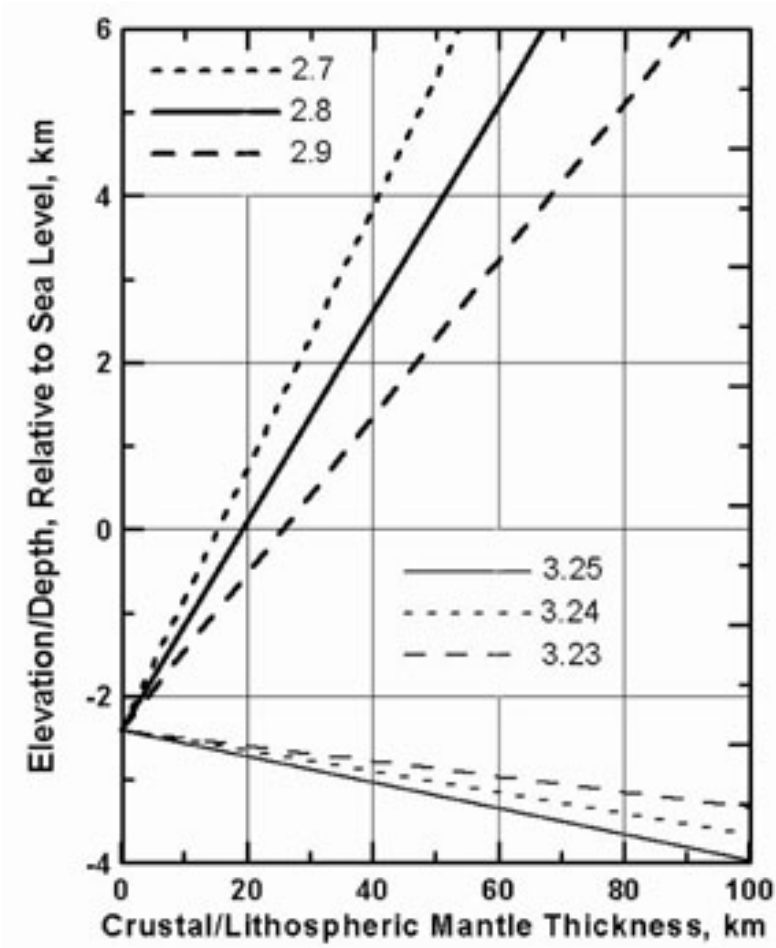


Figure 12. Individual contributions of crust and lithospheric mantle to the elevation of the continental surface. Elevation for the crustal contribution is given by taking its thickness and using the curve appropriate to its mean density (2.7-2.9 Mg^{-3}). Elevation for the lithospheric mantle contribution is given by taking its thickness and using the curve appropriate to its mean density: 3.25 Mg^{-3} is appropriate for cool mantle lithosphere; 3.24 Mg^{-3} for warm Basin and Range mantle lithosphere; and 3.23 Mg^{-3} for hot Basin and Range mantle lithosphere. Final elevation is calculated by adding the elevations for the crustal and lithospheric mantle contributions. [Modified from Lachenbruch and Morgan, 1990].

A summary of the different classes of published models applicable to uplift of the CP is given in Table 3. Primary uplift of compressional mountain belts, including the SRM, is generally assumed to be by physical crustal thickening, in the case of the SRM by horizontal mass transfer associated with Laramide compressional tectonics. However, one or more other mechanisms may be also responsible for additional and/or subsequent uplift. A discussion of most of these mechanisms with the information available at the time was given by Morgan and Swanberg (1985). The reader is directed to this reference for a more full discussion of individual mechanisms than is given below. As much as possible, the discussion below is limited by inclusion of all available constraints on the models.

Discussion

Subsidence Came Before Uplift. A relevant observation to the starting condition for models of uplift is that the surface of the CP-SRM region was close to sea level for much of the Paleozoic. Similarly, there are sedimentary rocks of Proterozoic age and Mesozoic age representing significant periods of time during which the surface was close to sea level. However, during these time periods, the lithosphere was slowly subsiding at the rate of sedimentation, always maintaining the approximate surface elevation at mean sea level. Over such a large area, modern gravity measurements indicate that, outside active subduction zones,

isostatic balance is generally maintained, and thus the relations among changes in crustal and/or

Table 3. Summary of uplift mechanisms suggested for the Colorado Plateau.

Physical Process	Mechanism	Reference
Thermal Expansion/ Lithospheric Thinning	Heating of lithosphere by deep mantle plume, hot spot, or unspecified thermal event.	1,2,3
	Heating due to overriding and subduction and subduction of an oceanic ridge.	4,5,6
	Viscous shear heating between lithosphere and asthenosphere.	7*,8*,9
	Heating following cessation of subduction and/or lithospheric thinning resulting from subduction.	10,11,12
	Heating due to delamination of mantle lithosphere and resultant replacement by hot asthenosphere.	13
Physical Crustal Thickening	Horizontal transfer of mass in lower crust by compression or unspecified mechanism.	14*,15,16
	Underplating or subduction at a very shallow angle.	17,18*
	Crustal inflation by material addition in a mid-crustal channel.	19
Phase Change	Expansion accompanying partial melting.	20
	Hydration reactions, such as serpentinization.	21*
	Introduction of volatiles, resulting in deep-seated hydration.	22,23
	Temperature/pressure-dependent solid-state phase changes, such as garnet granulite to eclogite	24*,25*,26*, 27

Table and references are intended to illustrate uplift mechanisms suggested for the CP and are not claimed to be bibliographical. Some mechanisms and references incorporate features of other mechanisms and/or physical processes, and references marked with an asterisk do not make specific reference to the CP. Key to references: 1 - Crough (1979); 2 - Bodell and Chapman (1982); 3 - Morgan and Swanberg (1985); 4 - Lipman *et al.* (1972); 5,6 - Damon (1979, 1983); 7 - Melosh and Edel (1979); 8 - Shaw and Jackson (1973); 9 - van Hunen *et al.* (2000); 10 - Chritiansen and Lipman (1972), 11 - Thompson and Zoback (1979); 12 - Spencer (1996); 13 - Bird (1979); 14 - Hess (1962); 15 - Gilluly (1963); 16 - Bird (1988); 17 - Helmstaedt and Schulze (1979); 18 - McKenzie (1984); 19 - McQuarrie and Chase (2000); 20 - McGetchin *et al.* (1980); 21 - Hess (1954); 22 - McGetchin and Silver (1972); 23 - Silver and McGetchin (1978); 24 - O'Connell and Wasserburg (1967); 25 - Lovering (1958); 26 - Green and Ringwood (1967); 27 - This study.

mantle lithosphere thickness indicated in Figure 12 are expected to apply as new sedimentary rocks were added to the crust. For example, in Arizona, the modern Paleozoic section is approximately 1,200 m in thickness with an average density of about 2.4 Mg m^{-3}. Assuming an average crustal density of 2.8 Mg m^{-3}, and an average mantle lithosphere density of 3.25 Mg m^{-3}, this thickness of sedimentary rocks could be added to the crust and the surface returned to sea level by thinning the crust by

$$\delta c = [(2.8 - 2.4)/(3.25 - 2.8)] \times 1{,}200 \text{ m} \simeq 1{,}070 \text{ m}, \quad (5)$$

or by thickening the mantle lithosphere by

$$\delta l = [(2.8 - 2.4)/(3.25 - 3.2)] \times 1{,}200 \text{ m} \simeq 9{,}600 \text{ m}, \quad (6)$$

or by some combination of thinning the crust and thickening the lithosphere. Complexities in the sedimentary record indicated by unconformities are not taken into account in these simplified calculations, but the changes in lithospheric structure required to maintain isostatic equilibrium are valid relative to overall sedimentary thickness changes relative to mean sea level.

Sedimentary basins in extensional settings and passive continental margins have two distinct phases of subsidence and sedimentation, a syn-tectonic phase, dominated by crustal thinning during extension (which also causes subordinate effect of extensional thinning of the mantle lithosphere). Syn-tectonic subsidence driven by crustal thinning is followed by post-tectonic subsidence that is consistent with thickening and cooling of the mantle lithosphere (thermal relaxation) back to its pre-extensional state (McKenzie, 1978). The post-tectonic subsidence is related to thermal contraction of the lithosphere as it cools, similar to the cooling and increasing depth of oceanic lithosphere away from a mid-ocean ridge. Extensional tectonics accompanied Proterozoic sedimentation in Arizona (Timmons *et al.*, 2001), so this sedimentation was probably associated with extension. For Phanerozoic sedimentation and subsidence of the CP-SRM lithosphere, however, there is no evidence of extensional tectonics contemporaneous with the initiation of the extended periods of sedimentation. There is also no characteristic slowing through time of sedimentation and subsidence that would suggest thermal relaxation associated with cooling and thickening of the lithosphere to drive subsidence. Some sedimentation seems to have been driven by external compressional events, but other sedimentation was associated only with global eustatic sea-level changes. Most of these sedimentary records have been preserved, however, suggesting that the lithosphere has adjusted to accommodate the sedimentary loads.

All of the mechanisms in Table 3 listed for uplift can be reversed to produce subsidence, but only one group can be considered as a likely candidate for internal compensation in response to external loading, and that group is phase changes. The first three types of phase changes in the table, partial melting, hydration reactions, and hydration involving the introduction of volatiles are unlikely to occur spontaneously because all involve the input of external components, such as heat, water, or other volatile components. However, temperature- and pressure-dependent phase changes, such as the solid-state transition from garnet granulite to eclogite could occur at any time in the lower crust/upper mantle in response to pressure/temperature changes. Loading by sedimentation from above, whether induced by a global eustatic rise in sea level, or a flexural downwarping caused by compressional stresses, would increase the pressure of rocks in the lower crust/upper mantle without significantly increasing their temperatures. Under these conditions, mafic granulite may be converted to eclogite (Green and Ringwood, 1967, 1972). This phase change causes a dramatic increase in density from about 2.8 Mg m^{-3} for garnet granulite to as high as 3.6 Mg m^{-3} for eclogite, although the net effect may be less if rocks of this composition (basaltic) comprise less than

100% at the depth at which the phase change occurs (Griffin and O'Reilly, 1987). Strictly speaking, the phase change could occur in either the lower crust or the upper mantle and there would be no formal changes in layer thicknesses, but changes in layer densities (and seismic properties). However, as discussed below, based on xenolith data, Griffin and O'Reilly (1978) have proposed that the phase change may dominate the seismic Moho in contrast to the chemical crust-mantle boundary.

Thus, a requirement of the long periods that the surface CP-SRM region was close to sea level, in a state of apparent stability, and slowly accumulating sediments, is that its lithosphere was dynamically responding to accommodate these sediments and maintain the surface at sea level. A viable mechanism for this dynamic response is the garnet granulite to eclogite phase change in the lower crust and/or upper mantle in response to loading of the sediments. Both garnet granulite and eclogite xenoliths have been reported from volcanic rocks of mid-Tertiary age from the central CP (*e.g.*, Wendlandt *et al.*, 1996; see also Helmstaedt and Schulz, 1979).

Uplift Through Thermal Expansion/Lithospheric Thinning. The primary mechanism controlling the elevation of the ocean floor is cooling and thermal contraction in the lithosphere/asthenosphere system as newly-formed oceanic lithosphere moves away from oceanic ridges (Turcotte and Schubert, 1982, p. 163). Cooling and subsidence appears to be reversible, as shown by the passage of oceanic lithosphere over a hot spot, creating anomalously shallow oceanic crust, such as the Hawaiian Swell, and similar mechanisms have been proposed for continental uplift (*e.g.*, Crough, 1979). In effect, any mechanism by which the mean temperature of the lithosphere is raised, causing thermal expansion and a decrease in mean density, or in which the negatively buoyant mantle portion of the lithosphere is thinned or removed will result in isostatic uplift (Table 3).

Most of the models involving thermal expansion/lithospheric thinning for the CP associate the heating to mid-Tertiary magmatic activity and/or thinning to Cretaceous to Paleogene low-angle subduction beneath the CP. Complete removal and replacement of the mantle lithosphere by asthenosphere beneath the CP by this subduction is precluded by mantle xenolith data that indicate that at least as late as 30 Ma, and in places (Grand Canyon Field, San Francisco Volcanic field) much later, the CP was underlain by mantle lithosphere of Proterozoic age (Liverccari and Perry, 1993; Riter and Smith, 1993, 1996; Alibert, 1994, Chen and Arculus, 1995; Smith *et al.,* 1999; Mattie *et al.*, 2000; Roden and Shimizu, 2000). Temperature and pressures deduced for some of the xenoliths from the center of the CP indicate that the lithosphere was cool and thick ($\geq$ 140 km) up to ~25 Ma, and thus, any thermal or thinning effects on the lithosphere must be constrained to below these depths. With reference to equation 3 and Figure 11, and, as concluded by Riter and Smith (1996), we must therefore conclude that uplift associated with subduction-related heating or thinning of the lithospheric mantle is not supported by the data.

Uplift Through Physical Crustal Thickening. Crustal thickening has clearly been an important mechanism in uplift of the SRM at some stage in their history, and is the primary mechanism for uplift in fold mountains. Crustal shortening evident in their structure would translate into crustal thickening and an increase in lithospheric buoyancy resulting in isostatic uplift. Lachenbruch and Morgan (1990) considered the reverse of this situation in which the lithosphere was stretched, and demonstrated that if the lithosphere was assumed to deform by pure shear, *i.e.*, lateral strains were equal at all vertical levels, then changes in surface elevation were independent of details in the vertical density structure of the lithosphere, depending only on the net buoyancy of the lithosphere. If pure-shear deformation is assumed for compressional

deformation of the CP and SRM, then the same result applies and change in elevation during lithospheric compression is described by:

$$\varepsilon + H_o = (\varepsilon_o + H_o)/(\beta) \qquad \varepsilon_o, \varepsilon > 0 \qquad (7)$$

where ε_o is the initial surface elevation, ε is the post-deformation surface elevation, H_o is the height of sea level above the free asthenosphere surface (see Figure 11), and β is the deformation factor given as the ratio of deformed length to undeformed length (from Lachenbruch and Morgan, 1990, equation 25, assuming only adiabatic lithospheric strain). For lithospheric extension, $\beta > 1$, and is usually called the stretching factor (McKenzie, 1978; Lachenbruch and Morgan, 1990). For compression, or lithospheric shortening, $\beta < 1$. Other factors can be included in equation 7 to include the effects of crustal erosion or sedimentation, magmatic thickening of the crust, mean temperature changes in the lithosphere, and thinning or accretion to the mantle lithosphere (Lachenbruch and Morgan, 1990). However, as presumably physical thickening of the CP-SRM crust would be restricted to times of Laramide compression, these additional factors are not considered as they would either decrease uplift of the CP, or are not supported by geologic evidence (widespread magmatism concurrent with compression).

There is little evidence in upper crustal rocks of the CP for significant crustal shortening. Deformation is dominated by widely spaced monoclines from which it is difficult to account for more than a few per cent of shortening across the CP (Bird, 1979). As discussed above, with mantle xenolith samples from as far west as the Grand Canyon and San Francisco volcanic fields in Arizona, with ages of $\leq$ 1 Ma, indicating a Proterozoic upper mantle, there is little room to cryptically thicken the CP crust by underthrusting from the west. In addition, lower crustal xenoliths from the San Francisco volcanic field indicate Proterozoic crustal growth ages and Cenozoic magmatism, inconsistent with cryptic crustal addition from the west. Thus, crustal shortening and thickening in the CP appears to be limited to deformation visible on the surface.

If a very liberal interpretation of surface shortening is taken for the CP of 10%, $\beta = 0.9$, if we assume that the plateau elevation was at sea level prior to deformation, $\varepsilon_o = 0$, and taking $H_o = 2.4$ km (from Lachenbruch and Morgan, 1990), then, using equation 7, the post deformation elevation would be $\varepsilon = 270$ m. This shortening and thickening includes thickening of the mantle portion of the lithosphere which may slowly reheat and thin back to an equilibrium thermal state, but this would account for no more than a few tens of meters of additional uplift. Thus, physical thickening of the crust is likely to produce no more than about 300 m of uplift of the surface associated with compression during the Cordilleran orogenies, probably less. Interestingly the amount of uplift predicted by these calculations is consistent with the estimated elevation of the surface of the CP <u>prior</u> to Laramide deformation when it experienced its last marine inundation during the high Late Cretaceous sea-level stand responsible for deposition of the Mancos/Pierre Shales and associated deposits.

A creative cryptic crustal thickening mechanism has been proposed by McQuarrie and Chase (2000a) who suggest that the crust of the CP has been inflated by about 14 km by intracrustal flow. Specifically they suggest that the Cretaceous crust of the CP-SRM region was ~30 km and isostatically balanced at sea level, basically in agreement with conclusions reached above, and that the addition of 14.5 km of mobile crust (density 2.8 Mg m^{-3}) reproduces the present structure of the CP and increases its elevation by 2 km. This mobile crust is added by middle crustal flow driven from the west by thick crust of the Sevier Plateau. The reader is referred to Macquarie and Chase (2000a) for geological and mechanical details of this proposal. Significant to CP evolution is their calculation that the flow would require a time duration of the order of 35 Ma, a west to east topographic slope across the CP of 1 to 5 km, and temperatures in

the middle crustal unit to maintain flow of 600 to 900°C (McQuarrie and Chase, 2000a). Kilty (2000) raised some questions concerning the mechanics of the model, to which McQuarrie and Chase (2000b) have responded. The model is examined here with respect to a few details specific to the CP.

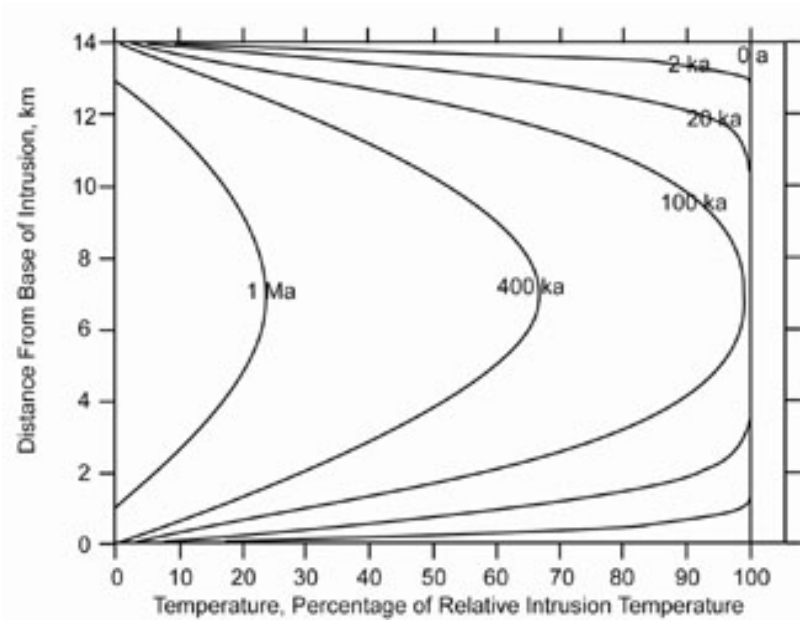
The topographic gradient associated with intrusion of the middle crustal layer proposed by McQuarrie and Chase (2000a) would not be expected to disappear immediately after flow ceased, as the high viscosity of the layer would be able to support a significant lateral stress. The driving head from the West, although reduced from its initial level, would also not disappear until the regional stress field changed in the middle Cenozoic, documented by the regional change from compressional to extensional tectonics. Therefore, the west to east topographic gradient across the CP would be expected to persist at least through the Paleogene. However, The Laramide formation of the SRM would have reversed the topographic gradient on the eastern side of the CP, and there is no evidence of a regional west to east Paleogene topographic gradient in the formation of the Eocene erosion surface nor in Paleogene lacustrine and fluvial deposits on the CP. In fact, there is no evidence for any gradient in the formation of the erosion surface, and arguments based on such a gradient are thus, unreliable.

The subduction events did not appear to affect the lithosphere underlying the CP and xenoliths have been interpreted to indicate that the lithosphere was cool and thick ($\geq$140 km) up to at least ~25 Ma (Riter and Smith 1993, 1996). Thus, the mobile layer proposed by McQuarrie and Chase (2000a) was intruded into cool crust. Assuming an average lithospheric geothermal gradient of 12°C km^{-1}, consistent with the xenolith data, the temperature at about 18 km, the proposed intrusion depth for the mobile middle crust layer would have been about 230°C if the mean surface temperature was about 15°C. This temperature is significantly lower than the range of temperatures proposed by McQuarrie and Chase (2000a) for the mobile middle crustal layer, and suggests that the mobile layer would lose heat to the host crust. As an order of magnitude calculation of how fast this heat loss would be, I have assumed that flow within the cooling layer was laminar and assumed that the flow moved at constant velocity at all levels within the layer. The laminar assumption is probably reasonable, but obviously the flow will be fastest in the center of the channel. However, This calculation gives a best-case estimate because it assumes that the flow continues at all levels until the center of the layer is too cool.

The results of this calculation are shown in Figure 13 and show that cooling of the intruding layer is a very significant factor that needs to be taken into account in this model. Even if the intrusion takes place at maximum thickness, the intrusion temperature drops to less than 25% of its original temperature excess after 1 Ma. If the initial temperature was 900°C and the host rock temperature was 230°C, the center of the intrusion would have dropped to less than 400°C. Similarly, with an initial temperature of 600°C, the center temperature after 1 Ma would be less than 325°C. With the highly nonlinear temperature behavior of viscosity, this loss of heat and drop in temperature is likely to prevent middle crustal flow it has flowed into the middle crust of the CP for about 1 Ma. However, the model of McQuarrie and Chase (2000a) requires flow for much longer time periods. Slower flow in a thinner initial channel would cause more rapid cooling, so the model presented here is a maximum cooling time for the mid0crustal channel in the CP, and heat loss appears to make this model non-viable.

A third test of the middle crustal flow model is uniformitarianism. The San Francisco volcanic field sits on the southwestern margin of the CP and is a dormant volcanic field. Its

Figure 13. Relative temperatures within a 14 km thick intrusion with margins fixed at zero



temperature at time equals zero. Assumed thermal diffusivity 1 $mm^2 s^{-1}$ (32 $km^2 Ma^{-1}$). Model represents a 14 km thick layer of middle crust intruding cool crust by uniform laminar flow so that its sides are continuously encountering new cool host rock. To calculate actual temperature, multiply difference between intrusion temperature and host rock temperature by the percentage of relative intrusion temperature and add to the host rock temperature. [Calculations modified from Lachenbruch *et al.* (1976)].

products include voluminous basaltic eruptions and more silicic products including the major andesitic constructs of the 3,850 m, Late Pliocene to Quaternary San Francisco Peaks. Regional elevation drops to around 1,000 m in the Verde Valle approximately 80 to the south, and to around 1,500 m in the broad drainage of the Little Colorado River to the northeast. The silicic volcanic products in the San Francisco volcanic field followed 5 Ma of basaltic volcanism, and indicate that the silicic volcanism resulted from crustal melting driven by heating from mafic magmas. Thus, we may infer that the crust in the region of the San Francisco volcanic field is very hot. However, despite a very hot crust, and large lateral topographic gradients, there is no indication of vertical movements in the region indicative of flow in the middle crust. Hence the middle crustal flow model fails the test of uniformitarianism, and this process should be regarded as unlikely until some supporting field evidence can be found.

Uplift Associated With Phase Changes. The final mechanism for crustal thickening and uplift is through phase changes. Phase changes associated with partial melting are considered unlikely because of the large area involved to produce partial melt. Phase changes associated with hydration reactions are possible, and these will be grouped together with the garnet-granulite to eclogite phase change. Spencer (1996) made calculations of uplift due to hydration associated with low-angle subduction and found that hydration could not produce enough uplift to be a serious contender for the cause of uplift. However, other phase changes, with or without hydration, remain a viable mechanism. This mechanism has two factors immediately in its favor: i) the mechanism also explains subsidence to allow the sections of sediments to slowly deposit on the crust in the region; and ii) xenoliths demonstrate that both garnet granulite and eclogite were in the lower crust/uppermost mantle during the middle Tertiary, as discussed above.

Heat flow in the interior of the CP is generally characterized as in the range 55 to 65 mW m^{-2} (Keller *et al.*, 1979, Reiter *et al.*, 1979; Thompson and Zobach, 1979; Bodell and Chapman, 1982; Reiter and Clarkson, 1983; Swanberg and Morgan, 1985) and if the higher heat flow of its marginal regions are included, its heat flow rises to 68 $\pm$ 26 mW m^{-2} (Morgan and

Gosnold, 1989). Even taking the lower interior heat flow range, however, this heat flow is significantly higher than that suggested by the cool lithosphere suggested by xenolith data as recently as ~25 Ma. (Riter and Smith, 1993, 1996). These data suggest significant heating of the CP lithosphere during the past 25 Ma, which would drive the phase change to convert eclogite to garnet granulite, resulting in a reduction in density and uplift.

One component that is commonly seen in regions of young volcanism and moderately high heat flow, such as the margin of the CP, and that is missing in this region is hot springs and other geothermal manifestations. Some thermal springs may be found around the Valles Caldera in New Mexico (e.g., Trainer, 1984), but, in general, these features are rare. The paucity of these features is thought to be related to the general aridity of the CP and the deep water table over much of the CP. Downward percolation of groundwater is thought to advect heat from the near surface in many areas of the CP, including the dormant San Francisco volcanic field where both hot and cold springs do not exist from any deep sources (J. H. Sass, personal communication, 2002).

Seismic data also suggest that the CP lithosphere is currently in a thermal state more similar to hot active tectonic regions than cool stable regions. Plots of P-wave velocity as a function of depth for the CP are shown in Figure 14 using data from Wolf and Cipar (1993). In general, these data show a gradational seismic Moho (see also Parsons *et al.*, 1996). The Moho is only abrupt on the Chinle and Hanksville columns which are reinterpretations of old data (Roller, 1965) with widely spaced recording stations and may be an artifact of the station spacing. If the CP seismic velocity profiles are compared with seismic velocity profiles from southeastern and western Australia, also shown in Figure 14, the CP profiles are similar to the southeastern Australia profiles, a region with a high geotherm and Cenozoic tectonothermal activity. By contrast, the cool, stable lithosphere of western Australia is characterized by a relatively thin crust, and an abrupt velocity change at its Moho. Western and southeastern Australia are characterized today by low and moderately high geotherms which qualitatively correlate with the CP geotherms prior to ~25 Ma and today, respectively (Figure 14).

Uplift associated with heating/thinning of the lithosphere post-25 Ma may produce as much as 1 km of uplift but is unlikely to produce the full complement of uplift required to raise the CP from its pre-Laramide elevation to its modern elevation (equation 3, Figure 12, see also references in Table 3 and elsewhere). However, as xenoliths have provided evidence for eclogite in the lower crust/uppermost mantle of the CP, it is possible that the conversion of eclogite to garnet granulite in the CP lithosphere may provide the additional component of uplift.

Griffin and O'Reilly (1987) proposed a model for the crust-mantle boundary in which this boundary was marked by a dominant mafic component at the boundary (100%) with a decreasing mafic component in felsic material above the boundary, and a decreasing mafic component in ultramafic material below the boundary (Figure 14). The model was based on the compositions of xenoliths collected primarily from southeastern Australia. Based also on xenolith data, they observed that the mineral phases of the mafic component changed according to the ambient geotherm, garnet granulites being more common with high geotherms, and eclogite being more common with cool geotherms. They then collected seismic velocity data on the xenoliths at appropriate pressures and compiled synthetic velocity-depth profiles using these data for comparison with field data from sites close to the xenolith collection locations. The results showed excellent agreement among synthetic and field velocity-depth profiles. The high

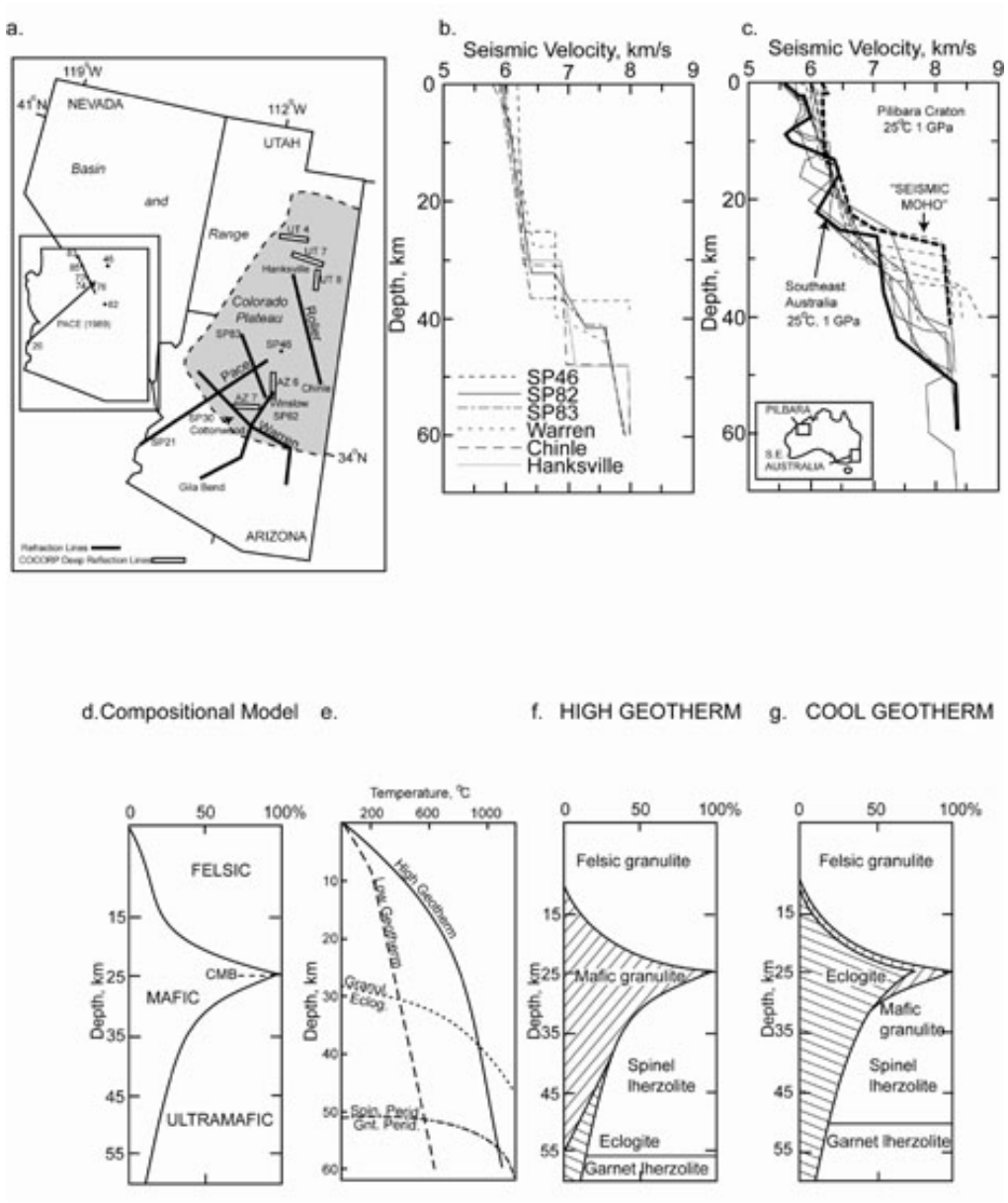


Figure 14. **a**. Map showing locations of seismic profiles shown in **b**. [From Wolf and Cipar, 1993]. **b**. P-wave velocity as a function of depth for seismic data on and close to the western Colorado Plateau. See **a**. for column locations: 1 - Winslow; 2 - SP 46; 3 - SP 82; 4 - SP 83; 5 - Chinle; 6 - Hanksville. [Data plotted from Wolf and Cipar 1993, Table 3 and Figure 14]. **c**. Vp-depth models for Western Australia (Pilbara) and for southeastern Australia. [After Drummond, 1982, from Griffin and O'Reilly, 1987]. **d**. Compositional model for lower crust and upper mantle. [Derived from Griffin and O'Reilly, 1987]. **e**. Geotherms for southeastern Australia (xenolith derived; From O'Reilly and Griffin, 1985] and Western Australia [(Extrapolated from surface heat flow; Sass and Lachenbruch, 1979]. The granulite-to-eclogite boundary is a conservatively high estimate. [From Griffin and O'Reilly, 1987]. **f**. and **g**. Rock types in the crust and upper mantle based on the composition model in **d**. based on the warm and cool geotherms from **e**., respectively. [Modified from Griffin and O'Reilly, 1987].

geotherm causes most of the mafic material to be in the garnet granulite phase, resulting in a gradational velocity transition at the Moho (velocity increase to 7.9-8.0 km s^{-1}), interpreted as a "thick" crust. The same chemical model for the lithopshere results in the mafic material in the eclogite phase for a cool geotherm with an abrupt velocity increase to 8.0 km s^{-1} and a distinct seismic Moho close to the crust-mantle boundary (Figure 14). The transition from eclogite to garnet granulite is accompanied by a decrease in density from ~3.6 Mg m^{-3} to ~2.9 Mg m^{-3}.

From the available data, the Griffin and O'Reilly (1987) model fits what is know about the CP. The CP had a cool geotherm prior to 25 Ma and eclogite xenoliths (and some garnet granulite xenoliths) were brought to the surface. Direct information on the elevation of the plateau is lacking. Paleobiological information from the SRM Eocene erosion surface have been interpreted to indicate that surface was at its present elevation, but even if these data are being interpreted correctly, their extrapolation to the CP is uncertain. Thus, the surface of the CP could have been at a relatively low elevation (~500? m) prior to 25 Ma, with much of the mafic component in its lithosphere in the eclogite phase giving a shallow Moho, and a crustal thickness of 30-35 km and a lithospheric thickness of >150 km.

Extensive Neogene bimodal magmatic activity has been manifested by volcanism and exposed by extensional tectonism in the Basin and Range and Rio Grande rift provinces on three sides of the CP. This activity was enhanced by extension in these provinces. The current heat flow in the interior of the CP and elevated heat flow and young volcanism along its margins indicate that the CP has been affected by the same thermal event that was responsible for the major Neogene magmatism in the western US. From the slow response of the surface heat flow to this event (e.g., see Morgan, 1983b), magmatic advection of heat into the CP lithosphere in association with this event is likely. This heat has raised the geotherm and possibly driven the eclogite to garnet granulite phase change so that most of the mafic component in the lithosphere is now in the garnet granulite phase. This has resulted in a thicker crust with a gradational Moho, and, together with heating, and perhaps thinning of the lithosphere, uplift of >1.5 km post ~25 Ma. No eclogitic xenoliths have been erupted on the CP post 25 Ma. Thus, available evidence supports the phase change/late heating model for uplift of the CP. Less igneous activity is evident on the CP because its lack of extensional tectonics has, in general, hindered the ascent of magmas through this province, and because more heat has been used heating the lithosphere and driving phases in the lithosphere than in the adjacent provinces.

Uplift of the Southern Rocky Mountains. The discussion has focused on uplift of the CP because an obvious mechanism for uplift in this region is not readily apparent. However, some questions about the Cenozoic elevation history of the SRM also remain. Crustal shortening and thickening associated with Laramide compression require Early Tertiary isostatic uplift of the SRM. Fission-track data provide quantitative evidence that suggests a second pulse of late Cenozoic uplift (Kelley and Duncan, 1984, 1986; Winkler *et al.*, 1999; see also papers referenced in Winkler *et al*., 1999). These data reveal distinct periods of Paleogene and Neogene denudation, the latter being consistent with a pulse of uplift interpreted by many workers from geological and other evidence (*e.g.*, Epis and Chapin, 1975; Seager *et al*., 1984; Decker, 1995; Pazzaglia *et al*., 1999) and the recent erosion rates (Small and Anderson, 1998).

The fission-track data may be interpreted in terms of denudation through tectonics associated with the Neogene development of the Rio Grande rift and/or the delay in erosion of Paleogene topography associated with climate change (Winkler *et al*., 1999). As with the CP, dating of the final development of the regional elevation of the SRM is not completely resolved

(*e.g.*, Chapin, 1999). However, geophysical data indicate that the Rio Grande rift and the SRM extending up to the Wyoming state line are underlain by very hot crust (Decker *et al*., 1980, 1984, 1988; Decker, 1995). The heat flow in this region implies unrealistically high equilibrium temperatures near the crust-mantle boundary if steady-state conditions are assumed, and the preferred interpretation is that the geotherm is not in equilibrium but that gravity lows in the area are caused by young (10 - 1 Ma) intrusions in a late Tertiary rhyolitic complex in the upper crust (Decker, 1995). The source for these upper crustal intrusions is likely to be magmatic thickening of the crust, which would be a source of late Cenozoic uplift of the northern Rio Grande rift and SRM. Thus, in this region, there is good evidence for a source mechanism for late Cenozoic uplift, even though the direct evidence for uplift may be questioned.

The heat flow data raise interesting questions about late Cenozoic uplift of the SRM. By analogy with the CP, a significant pulse of late Cenozoic SRM uplift may be inferred driven by thermal lithospheric thinning, thermal expansion, magmatic crustal thickening, and possibly a thermally induced phase change. The current elevation of the SRM and the continuity of their elevation with adjacent provinces (Eaton, 1986, 1987) suggests that perhaps their elevation is young and not just a result of structural crustal thickening. There are not sufficient data to answer these fascinating questions at present. With new tools, however, such as the vesicular basalt paleo-altimeter, we may attack these questions in the near future.

Conclusions

The Colorado Plateau-Southern Rocky Mountain region has not been a distinct geologic region until relatively recently. There was probably nothing to distinguish these portions of Proterozoic lithosphere from adjacent lithospheric sections added to North America in the Paleoproterozoic until the formation of the Ancestral Rocky Mountains. Although commonly designated as “cratonic” because of their long periods of stability and flat sedimentary sections, these sections of lithosphere were probably stable primarily through an absence of deformational stresses rather than inherent lithospheric strength. As the periods of stability and sedimentation were punctuated by widespread, if gentle deformation, platform may be a more appropriate province tectonic description. However, the long period of Paleozoic sedimentation, and possibly other periods of sedimentation in the region, although undoubtedly in some manner related to plate margin events, do not follow subsidence patterns expected from flexural loading or extension and thermal subsidence related to plate margin interactions. Long periods of continual sedimentation (brief periods of about 10 Ma over a few 100 Ma) at or close to sea level closely responding to the global eustatic sea-level curve indicate amplified subsidence responding simply to sedimentary loading. The mechanism responsible for amplifying the subsidence is likely to have been a phase change in the lithosphere, probably the garnet granulite to eclogite phase change at or near the crust-mantle boundary.

The first localized deformation in the region of the CP-SRM was the formation of the Ancestral Rocky Mountains and this localization is believed to have been caused by a focusing of stresses by the plate boundary forces responsible for this deformation. Stresses at the plate boundary associated with the Ouachita orogeny were not uniform along a straight boundary, but heterogeneous along a sinuous boundary, and internal plate heterogeneities and deformation between the plate boundary and nascent Ancestral Rocky Mountains were probably responsible for the final focusing of stresses that formed the Ancestral Rockies. The Pennsylvanian deformation included areas in Texas and Oklahoma, and importantly for this study, the region that was to become the SRM. When more regional west to east compressional stresses returned

with the Cordilleran orogenies 80 to 150Ma later in the Late Jurassic/Cretaceous, either the thermal perturbation in the SRM lithosphere from Ancestral Rocky Mountain deformation was not fully relaxed (Morgan, 1984; Morgan and Sass, 1984), and/or a structural weakness remained in the SRM lithosphere from Ancestral Rocky Mountain deformation, and deformation was reactivated in the SRM, particularly in the last of the Cordilleran orogenies, the Laramide orogeny. Other regions exist where stresses appear to have been transmitted across an undeformed area to a zone of deformation, and the undeformed area is commonly characterized as particularly rigid (*e.g.*, the Tarim basin between Tibet and the Tien Shan, Kao *et al.*, 2001). However, as prior to the Cordilleran orogenies the CP showed no special resistance to deformation relative to surrounding regions, the interpretation that the lithosphere of the SRM deformed preferentially through weakness is preferred to invoking special properties of rigidity for the CP lithosphere.

Prior to the Cordilleran orogenies the sedimentary record indicates that the surface of most of the Paleoproterozoic lithosphere of North America was commonly close to sea level, and this is especially valid for the Phanerozoic. Some regional uplift may have occurred during the Sevier orogeny, followed by the formation of a structural basin and deposition thick Cretaceous marine sediments ending with deposition of the Late Cretaceous Mancos Shale and associated units, a time of particularly high sea level, pegging the regional surface elevation at most at a few hundred meters above present day sea level. We know the present elevations of the CP and SRM. However, constraining the elevation histories of the CP and SRM between their last marine inundation and the present is difficult and controversial. Most geological elevation indicators yield information concerning relative elevation, and there has been an elevation inversion between the CP and the adjacent Basin and Range province which complicates these interpretations. Erosion data related to elevation also depend on relative topography and also on climate which is known to have varied in this region during the Cenozoic. Fossil data related to elevation depends not only on an understanding of the climatic interactions of slope angle and azimuth to elevation, but also the climatic changes and the effects of microclimates. Fission-track data, although commonly interpreted in terms of uplift actually measure erosion or denudation, and may similarly be affected by climate and tectonics. However, a new technique for measuring paleo-elevations, based on the vesicularity of basaltic lava flows measures absolute elevations relative to paleo-sea level, indicates rapid uplift of about 1 km of the CP since 5 Ma.

More information is available for the evolution of the lithospheric mantle of the CP than for its surface elevation through the Cenozoic. This information comes from mantle xenolith samples, some of which relates to their depth of origin through geothermometry and geobarometry studies. These results indicate that the lithospheric mantle beneath the CP has remained attached to the overlying crust in the sense that it yields Proterozoic chemical and isotopic signatures and Paleoproterozoic ages. This lithospheric mantle remained cool to depths of at least 140 km until at least as recently as ~25 Ma ago in the center of the CP, and mantle xenolith samples brought to the surface as recently as 1 Ma and as far west as the Grand Canyon volcanic field continued to yield unaltered Proterozoic signatures. Mid-Tertiary xenoliths included eclogites, but younger xenoliths have included no eclogites. These results essentially preclude any models of uplift for the CP that require replacement of the CP mantle lithosphere by underthrusting slabs, delamination, or significant heating of the CP lithosphere, any of which affect the lithosphere above 140 km at any time prior to or at ~25 Ma, and any mechanism that requires complete replacement of the mantle lithosphere at any time prior to at least as recently

as ~1 Ma. Geophysical data constraining the modern structure of the CP lithosphere indicate a relatively thick crust and an intermediate thickness lithosphere, consistent with recent and ongoing heating and uplift. These data may be explained by crustal thickening primarily by a phase change, again most probably the garnet granulite to eclogite phase change, and the eclogite to garnet granulite phase change in mafic rocks around the chemical crust-mantle boundary is consistent with all available data. Seismic structure of the lower crust-upper mantle in the southeastern CP is also consistent with this phase-change interpretation.

Laramide deformation caused contemporaneous uplift in the SRM, and there is evidence of a well-developed late Eocene erosion surface in this region. However, the elevation of this erosion surface at and subsequent to the time of its formation is not well established. Fission-track and cosmogenic-radionucleide data indicate a late Neogene pulse of erosion in the SRM which is consistent with a second late Cenozoic phase of SRM uplift, although these data could also be explained by other mechanisms. However, thermal evidence for very high crustal temperatures in the northern Rio Grande rift and the nonextended SRM up to the Wyoming state line indicate Late Miocene to Holocene magmatic activity in the upper crust with associated magmatic thickening of the crust in this region, which would cause uplift. Hence, uplift is the preferred dominant mechanism to explain the fission-track and cosmogenic-radionucleide data, and evidence for a late Cenozoic episode of SRM uplift is considered to be good.

Problems abound in unraveling even the Cenozoic history of uplift and erosion of the Colorado Plateau and Southern Rocky Mountains. There does not seem to be a technique that is golden bullet for elucidating the tectonic history of this region, but a multi-disciplinary approach may be the most productive. Xenolith studies have provided important constraints on uplift models and fission-track and cosmogenic-radionucleide data are beginning to provide important real constraints on erosion and denudation ages. Results from the new basalt-vesicule paleo-altimeter have provided important preliminary Neogene elevation constraints and the technique promises further useful constraints. However, more detailed studies of the sedimentary history, crustal and upper mantle structure, dates, styles and amounts of deformation, and timing and implications of magmatism will all make contributions to the solution to this Gordian knot.

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