Morgan, P. & Swanberg, C. A. On the Cenozoic uplift and tectonic stability of the Colorado Plateau. J. Geodyn. 3, 39–63

Article *in* Journal of Geodynamics · February 1985

DOI: 10.1016/0264-3707(85)90021-3 · Source: NTRS

ON THE CENOZOIC UPLIFT AND TECTONIC STABILITY OF THE COLORADO PLATEAU

P. MORGAN[1]* and C. A. SWANBERG[2]**

[1] *Lunar and Planetary Institute, 3303 NASA Road One, Houston, Texas 77058, USA.*
[2] *Departments of Physics and Earth Sciences, Box 3D, New Mexico State University, Las Cruces, New Mexico 88003, USA.*

(Accepted January 28, 1985)

ABSTRACT

Morgan, P. and Swanberg, C. A., 1985. On the Cenozoic uplift and tectonic stability of the Colorado Plateau. *Journal of Geodynamics*, 3: 39–63.

The Colorado Plateau is a major tectonic and physiographic province in the southwestern United States which has been relatively stable during the Phanerozoic, but has been epeirogenically uplifted about 2 km in the Cenozoic. Minor deformation occurred within the Plateau during Late Cretaceous-early Tertiary compressional orogeny, and three phases of Cenozoic igneous activity are represented on the Plateau. Constrains on timing of plateau uplift are poor, but there appears to have been at least two phases of uplift, the most recent being during the last 5 Ma. Numerous mechanisms for plateau uplift can be divided into three groups: 1. Thermal expansion, 2. crustal thickening, and 3. phase changes, and simple models indicate that each of these models seems physically capable of producing the required uplift. When the geologic and geophysical data for the Plateau are considered, however, it seems likely that more than one mechanism is required to explain the 2 km of uplift. Thermal expansion associated with lithospheric thinning and magmatic crustal thickening seem the most reasonable mechanisms to have caused Cenozoic uplift of the Plateau. Early Cenozoic stability of the Plateau was possibly related to the relatively cool geotherm and thus high strength in the Plateau lithosphere relative to the adjacent Basin and Range and Southern Rocky Mountain-Rio Grande rift provinces. Local topography could also have been a factor in maintaining the mechanical integrity of the Plateau. As the Plateau lithosphere has been heated and uplift has occurred, the stability of the Plateau appears to have decreased, and it may now be breaking up, especially in its margins.

INTRODUCTION

In an address to the Washington Philosophical Society in 1889 C. E. Dutton opened with a statement of what were, in his opinion, the three greatest

* Present address: Dept. Geosciences, Purdue University, West Lafayette, IN 47907, USA.
** Present address: GRI Operator Corporation, 545 Middlefield Road, Suite 200, Menlo Park, CA 94025, USA.

0264-3707/85/$3.00

problems of physical geology (Dutton, 1892). The second of these problems was, "What is the cause of the elevation and subsidence of restricted areas of the earth's surface?" Dutton's interest in this problem was undoubtably inspired by his geological studies of the Colorado Plateau during the previous ten years (Dutton, 1880, 1882a, b, 1885) which he discussed later in his address. He concluded his address with the statement:

> "Hence I infer that the cause which elevates the land involves an expansion of the underlying magmas, and the cause which depresses it is a shrinkage of the magmas. The nature of the process is, at present, a complete mystery."

Almost a hundred years later we are still studying the problem of uplift of the Colorado Plateau, and attempting to constrain and quantify the density changes in what we now call the lithosphere, alluded to as "expansion of the underlying magmas" by Dutton, that appear to have caused the uplift. In the present contribution to this debate we present models of plateau uplift applied to the Colorado Plateau and briefly examine the sequence and consequences of uplift of the Plateau and volcanism and rifting in the adjoining Basin and Range and Southern Rocky Mountain-Rio Grande rift provinces. In an accompanying paper (Swanberg and Morgan, this volume), we present new heat flow estimates and a new heat flow map of the Colorado Plateau adjoining areas.

CENOZOIC GEOLOGIC HISTORY OF THE COLORADO PLATEAU

The Colorado Plateau is a major tectonic and physiographic province in the southwestern United States. It covers an area of approximately $3.5 \times 10^5\ km^2$ in northwestern New Mexico, western Colorado, and central, southern and eastern Utah (Fig. 1), which appears to have been relatively stable during most of the Paleozoic and Mesozoic, suggesting that its general plateau structure, which is of Cenozoic age, is an inherited feature (e.g., see Hunt, 1956). Widespread occurrence of the marine Mancos shale on the Plateau (e.g., Oetking and others, 1967) indicates that the Plateau was close to sea level in the Late Cretaceous, and from present mean altitude indicates approximately 2 km of uplift during Cenozoic. Although minor orogenic structures formed within the Plateau during early Tertiary time, the Plateau escaped most of the major Late Cretaceous-Tertiary deformation and magmatic events that were pervasive in the Basin and Range and Southern Rocky Mountain-Rio Grande rift provinces which border the Plateau to the west, south and east. Valuable reviews of the geology and tectonics of the Plateau region include those by Hess (1954), Kelley (1955), Hunt (1956), Eardley (1962), Gilluly (1963), Pakiser (1963), and

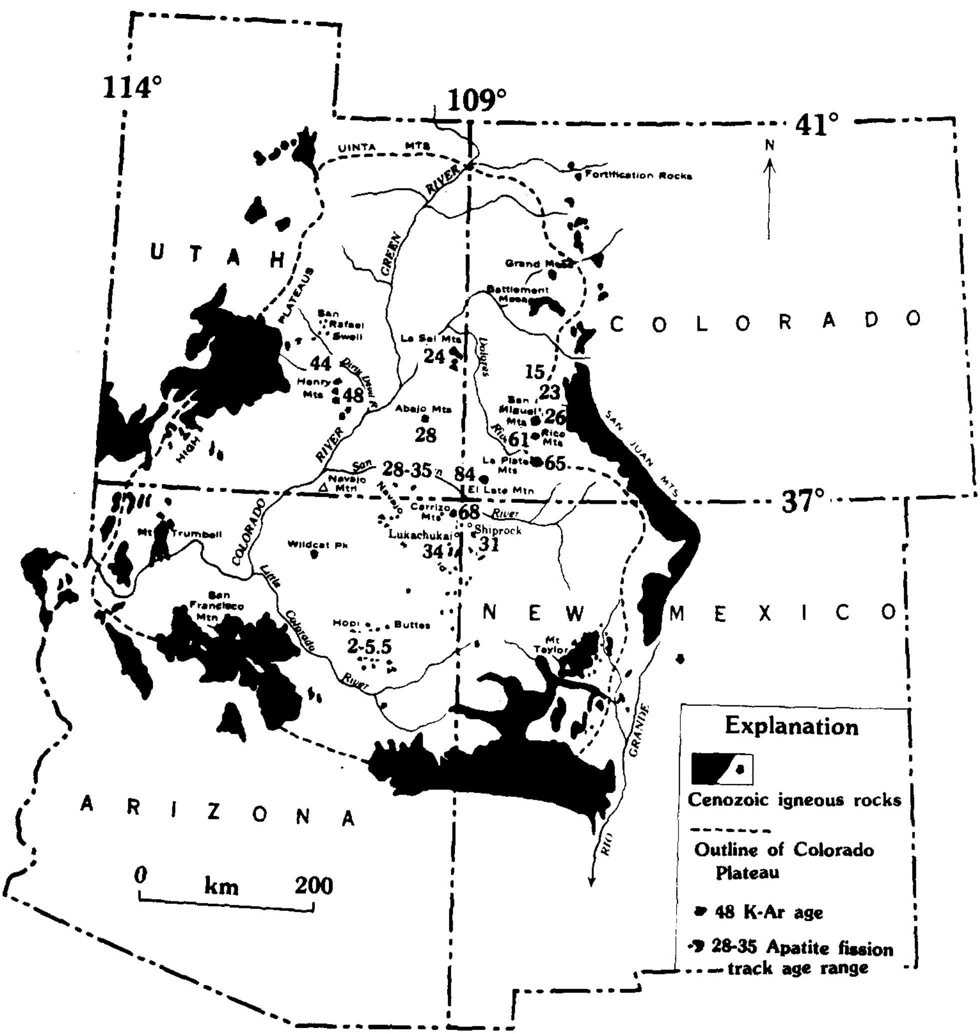

Fig. 1. Sketch map showing the location of the Colorado Plateau and the distribution of Cretaceous and Cenozoic igneous rocks on and immediately adjacent to the Plateau (adapted from Hunt, 1956). Age dates are given where available for igneous rocks primarily in the interior of the Plateau from Armstrong (1969) and Naeser (1971).

McGetchin and others (1980), and the reader is referred to these studies for general details of the Cenozoic geologic history of the Plateau. For the present purposes of understanding the Cenozoic uplift and stability of the Plateau, a summary of the Cenozoic history of the Plateau will suffice, emphasizing only details directly relevant to the problems under investigation. The following summary of the Cenozoic geologic history of the Plateau, illustrated in Figure 2, was taken primarily from Hunt (1956).

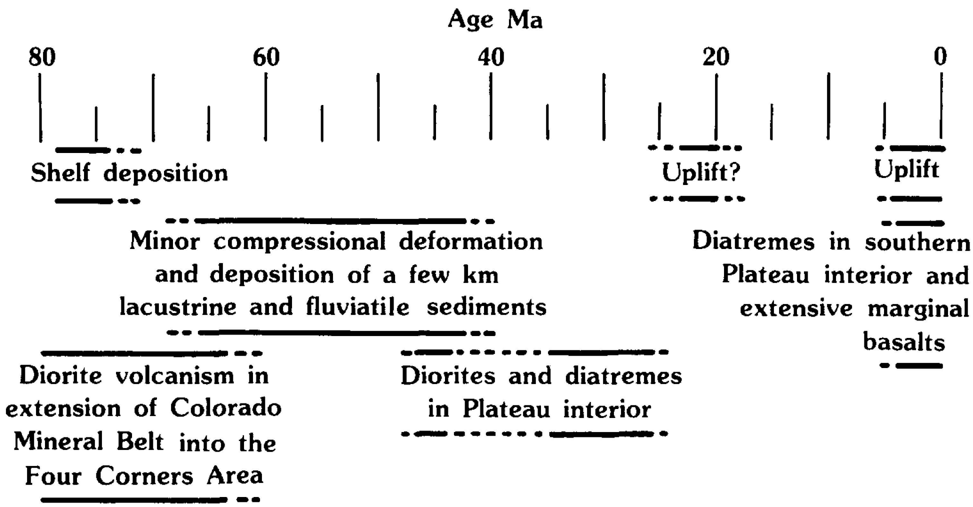

Fig. 2. Late Cretaceous and Cenozoic time chart for the main tectonic and volcanic events affecting the Colorado Plateau.

Prior to the Laramide (Late Cretaceous-early Tertiary) orogeny in the western U.S. the Plateau was a shelf area. It was transformed into a basin or trough, surrounded by newly formed mountains during Laramide folding, and the deposition of several thousand feet of lacustrine and fluviatile sediments in the lower parts of this depression indicates that it was still close to sea level during the early Tertiary. After the Eocene, general aggradation ceased in the Plateau, and general degradation began. Igneous activity, volcanism and intrusion, became common, although much less extensive than that in the neighboring provinces, and faults developed, especially along the west and south edges of the Plateau; uplift began. Uplift and erosion have continued to the present, with volcanism and faulting being mostly confined to the Plateau margins. The general erosion-sedimentation history of the adjacent Basin and Range province is opposite to that of the Plateau, with highlands undergoing erosion during the Paleocene, Eocene and Oligocene, and extensive aggradation during Late Miocene and Early Pliocene in the Basin and Range.

The general distribution of Cenozoic igneous rocks on the Colorado Plateau and approximate age dates for igneous rocks in the central portion of the Plateau are shown in Figure 1. Sporadic and sometimes extensive igneous activity has occurred adjacent to and sometimes overlapping the western, eastern and southern margins of the Plateau throughout the Cenozoic (e.g., Snyder and others, 1976), and young (Late Miocene to Holocene) volcanic rocks are common on the Plateau margins (e.g., Luedke and Smith, 1978a, b). Hunt (1956) considered most of the Plateau igneous activity to be late Cenozoic in age, primarily from geomorphic relationships. From their K-Ar isotopic age and fission-track ages, however, igneous rocks in the central portion of the Plateau can be divided into three age groups: Laramide (80 to 50 Ma), mid-Tertiary (46 to 24 Ma), and late Tertiary (less than 15 Ma), and these periods of activity roughly correspond to increases in igneous activity around the Plateau.

Laramide igneous activity in the central Plateau was primarily represented by dioritic laccoliths which extend from the Four Corners area of the Plateau (the junction of the four states of Utah, Colorado, New Mexico and Arizona) northeast into Colorado along the projection of a trend of Laramide mineralization which crosses central Colorado (the Colorado Mineral Belt) (Fig. 1; and Armstrong, 1969). Dioritic laccoliths are also the dominant form of mid-Tertiary igneous activity in the central portion of the Plateau, but these intrusions do not follow any discernable spatial trend. During the same period, contrasting kimberlite, carbonate and minette (alkali-lamprophyre) dikes and diatremes of the Navajo field were also intruded, primarily in the Four Corners area of the Plateau (Fig. 1, Williams, 1936; Watson, 1967; McGetchin and Silver, 1970a, b, 1972; Naeser, 1971; McGetchin and others, 1973). Late Cenozoic igneous activity in the central portion of the Plateau is primarily represented by monchiquite diatremes of the Hopi Buttes volcanic field in northeastern Arizona with ages around 2 to 5.5 Ma (Fig. 1; and Williams, 1936; Hack, 1942; Hunt, 1956; Naeser, 1971), and late Tertiary-Quaternary, primarily basaltic volcanic rocks are common on the Plateau margins (Hunt, 1956; Snyder and others, 1976; Luedke and Smith, 1978a, b).

The timing of uplift of the Colorado Plateau from below the sea level in Late Cretaceous to about 2 km now, is crucial for interpreting the mechanism of uplift, but unfortunately details of this timing are elusive. The following summary is taken primarily from McGretchin and others (1980), based upon studies by Damon (1979), Kelly (1979), Lucchitta (1979), Mayer (1979), Otton and Brooks (1978), Pierce and others (1979), Shoemaker (1978), and Young (1979). Prior to 24 Ma the Plateau was topographically low with internal drainage, although some relatively high spots may have existed as a result of Laramide (Late Cretaceous-early Tertiary) deformation and Eocene and Oligocene magmatic activity. Hunt (1976) reports 1000 m of downcutting below the present top of the Plateau by the Colorado River in Peach Springs Canyon at 18 Ma, indicating a Plateau elevation of at least 1 km at 18 Ma. Prior to about 10 Ma the Plateau was topographically low with respect to the adjacent Basin and Range province, but this topographic relationship was reversed at about 10 Ma. At the end of Basin and Range faulting (about 8 Ma), the Plateau was 1.1 km above adjacent basin floors in western Arizona, which were

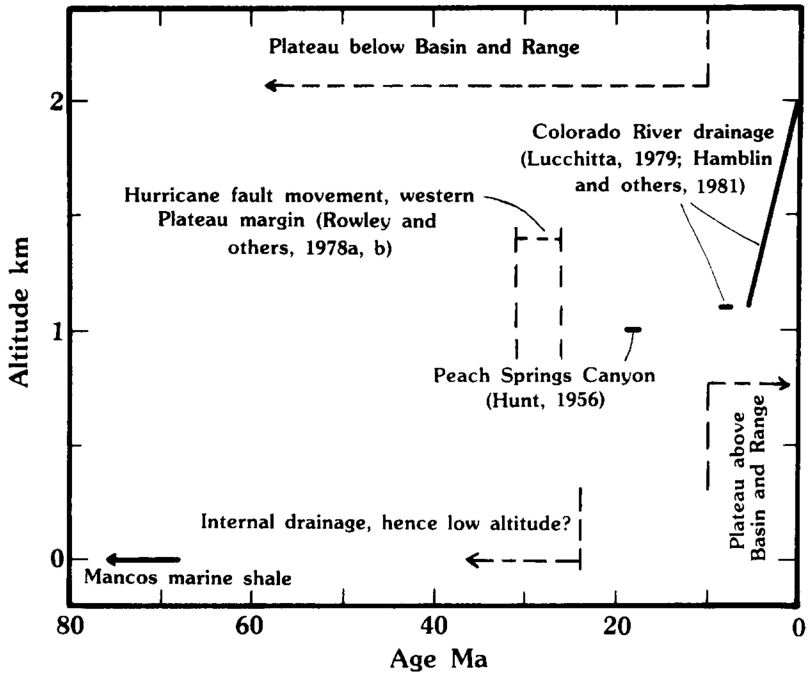

Fig. 3. Compilation of uplift and altitude data for the Colorado Plateau.

thought to be close to sea level (Lucchitta, 1979). Deposits from the Colorado River draining the Plateau indicate that during the last 5.5 Ma the Basin and Range adjacent to the Plateau has been uplifted at least 550 m, and the Plateau has risen at least 880 m in western Arizona relative to sea level (see also Hamblin and others, 1981). Thus, there is evidence that Plateau uplift has occurred in at least two stages, and that the later stage was very recent. Uplift data for the Plateau are summarized in Figure 3.

GEOPHYSICAL STUDIES OF THE COLORADO PLATEAU

Geophysical data provide information about the subsurface structure of the lithosphere, and thus provide constraints on modern lithospheric structure that any models for uplift of the Colorado Plateau must meet. Geophysical reviews of the Colorado Plateau have recently been published by Thompson and Zoback (1979) and Keller and others (1979), and the salient points of these reviews and of later relevant publications are summarized here.

Seismic data indicate that the crust of the Colorado Plateau is about 45 km thick, which is significantly thicker than the adjacent Basin and

Range province to the west and south (about 30 km thick) and the Rio Grande rift to the east (30 to 35 km thick). No significant change in crustal thickness is associated with the northern boundary of the Plateau. Heat flow in the interior of the Plateau is probably in the range 55 to 65 mW m^{-2} (Keller and others, 1979; Bodell and Chapman, 1982; Reiter and Clarkson, 1983a; Swanberg and Morgan, this volume), but low upper mantle compressional wave velocities ($P_n = 7.8$ km s^{-1}; Keller and others, 1979) and high mantle electrical conductivities (e.g., Pedersen and Hermance, 1981) suggest relatively high upper mantle temperatures, higher than those suggested by simple extrapolation of the surface of heat flow. Gravity data indicate that the Plateau is regionally close to isostatic equilibrium. Very low upper mantle seismic velocities, belts of active seismicity, anomalously thin crust, and late Cenozoic volcanism and faulting are associated with the western, southern and eastern margins of the Plateau, and these anomalous zones extend as much as 100 km into the Plateau. Thus, geophysical data indicate that the Plateau is approximately in isostatic equilibrium, and although heat flow in the interior of the Plateau is not particularly high, upper mantle temperatures beneath the Plateau are high, and conditions in the lithosphere of the Plateau margins are more similar to conditions in the adjacent provinces than in the Plateau interior.

ISOSTATIC UPLIFT MECHANISMS

Numerous mechanisms have been suggested for uplift to the Colorado Plateau (e.g., see McGetchin and others, 1980, Table III), many of which are genetically related to the Cenozoic plate tectonic evolution of the western United States. The physical concepts of uplift behind many of these mechanisms are similar, however. The Plateau is in approximate isostatic equilibrium at present, and if it is assumed that it was also in approximate isostatic equilibrium prior to uplift, its present altitude must be supported by a relative mass deficiency at depth. The suggested causes of this mass deficiency can be divided into three groups: 1. Thermal expansion (heating and/or thinning of the lithosphere); 2. Crustal thickening; and 3. Phase changes. Suggested uplift mechanisms divided into these three types of mass deficiency are listed in Table I.

Thermal expansion in various forms is the most commonly quoted mechanism for uplift of the Colorado Plateau (Table I). It basically depends upon the mean coefficient of thermal expansion of the lithosphere, and the integrated difference between the geotherm prior to uplift and the modern geotherm down to the depth of isostatic compensation. Unfortunately, although we now have reasonable estimates of the modern surface heat flow from the Colorado Plateau (e.g., Swanberg and Morgan, this volume), as it is likely that the Plateau lithosphere is not in a state of thermal equilibrium (e.g., Bird, 1979; Bodell and Chapman, 1982) and as at present we have no estimates of the contribution of crustal radiogenic heat production to the Plateau heat flow, even the modern geotherm is subject to great uncertainties. We have even less constraints upon the pre-uplift

TABLE I

Summary and grouping of Uplift Mechanisms suggested for the Colorado Plateau.

Physical Process	Mechanism	Reference
Thermal Expansion	Heating of lithosphere by deep mantle plume, hot spot or unspecified thermal event.	1,2
	Heating due to overriding and subduction of an oceanic ridge.	3,4,5
	Heating due to viscous shear heating between lithosphere and asthenosphere (not suggested for Colorado Plateau).	6,7
	Heating following cessation of subduction and resulting thermal equilibration of subducted slab.	8,9
	Heating due to delamination of mantle portion of lithosphere and resultant rise of asthenosphere.	10
Crustal Thickening	Horizontal transfer of mass in lower crust by compression or unspecified.	11,12
	Underplating or subduction at very shallow angle.	13,14
Phase Change	Expansion accompanying partial melting.	15
	Hydration reactions such as serpentinization.	16
	Introduction of volatiles resulting in deep-seated hydration.	17,18
	Temperature dependant solid-state phase changes such as eclogite-basalt or spinel-olivine (not suggested forColorado Plateau).	19,20

Table and references are intended to illustrate uplift mechanisms suggested for the Colorado Plateau and are not intended to be comprehensive. Some mechanisms and references incorporate features of other mechanisms and/or physical processes. Key to references: 1-Crough (1979); 2-Bodell and Chapman (1982); 3-Lipman and others (1972); 4-Damon (1979); 5-Damon (1983); 6-Melosh and Edel (1979); 7-Shaw and Jackson (1973); 8-Christiansen and Lipman (1972); 9-Thompson and Zoback (1979); 10-Bird (1979); 11-Hess (1962); 12-Gilluly (1963); 13-Helmstaedt and Schulze (1977); 14-McKenzie (1984); 15-McGetchin and others (1980); 16-Hess (1954); 17-McGetchin and Silver (1972); Silver and McGetchin (1978); 19-O'Connell and Wasserburg (1967); 20-Lovering (1985); 21-Green and others (1980, Table III).

geotherm. It is easy to demonstrate that thermal expansion is a viable mechanism for uplift of the Colorado Plateau, but less easy to place reliable constraints upon this mechanism.

If we assume that the Plateau was in thermal equilibrium prior to uplift it is possible to use a simple model to estimate the maximum uplift possible by thermal expansion alone. As shown by Morgan (1983), maximum uplift is predicted for thermal expansion if the heated (or thinned) lithosphere is in a state of thermal equilibrium, and thus we compare hypothetical pre- and post-uplift equilibrium geotherms (Fig. 4). Maximum temperatures in the lithosphere are controlled by temperature of the asthenosphere, T_a, which we assume to be constant as a first approximation. We limit lithospheric thinning to the level of the Moho because any thinning of the buoyant crust

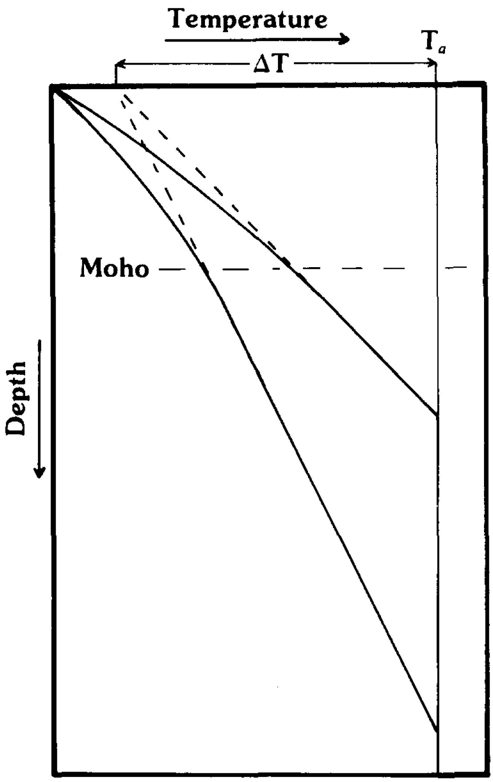

Fig. 4. Schematic equilibrium geotherms to illustrate the change in geotherm associated with lithospheric thinning and plateau uplift by thermal expansion. T_a is the temperature of the base of the lithosphere (assumed to be constant here as a first approximation), and ΔT is the effective temperature drop across the lithosphere, defined as the temperature drop across the lithosphere minus the temperature increment in the crust due to crustal radiogenic heat production and due to the thermal conductivity contrast between the crust and mantle (geotherms for effective temperature drop illustrated by dashed curves).

would cause subsidence, and this limitation allows us to ignore the crust in our models as a first approximation. The geotherms are adjusted for the effects of crustal heat production and conductivity variations as these effects remain essentially unchanged during heating (dashed lines, Fig. 4). Using these series of simplifying assumptions, we can calculate the mean density of the lithosphere, ρ_L, relative to the asthenosphere density, ρ_A, as

$$\rho_L = \rho_A(1 + \alpha\, \Delta T/2) \qquad (1)$$

where α is the mean coefficient of volume expansion, and ΔT is the effective temperature drop across the lithosphere (asthenosphere minus surface temperature, corrected for crustal heat production and conductivity). Assuming conditions of isostasy, it is easily shown that the ratio of surface uplift, U, to the lithospheric thinning (or that portion of the lithosphere heated to the asthenosphere temperature), ΔL, is given by

$$U/\Delta L = (\rho_L - \rho_A)/\rho_A = \alpha\, \Delta T/2 \qquad (2)$$

and plots of uplift as a function of lithospheric thinning are shown in Fig. 5. Comparison of this simplified analysis with a more complete treatment of the problem (e.g., Crough and Thompson, 1976; Bird, 1979; Mareschal, 1981; Morgan, 1983; Morgan and Phillips, 1983) indicates that the error involved with ignoring the crust is only a few percent, much less than the uncertainties in α and ΔT. As the Plateau is probably not in thermal equilibrium, its geotherm will not conform to the equilibrium heat flow assumptions used for the geotherms in the uplift calculations here. However, these simple calculations are useful for constraining the amount of heating and/or lithospheric thinning required to uplift the Plateau by thermal expansion.

Data recently compiled by Roy and others (1981) indicate that the volume coefficient of expansion for olivine from about 0 to 1000°C is in the range 3.0 to 3.8×10^{-5}°C^{-1}, and that for pyroxene and hornblende it is in the ranges 1.65 to 7.8×10^{-5} and 2.4 to 3.6×10^{-5}°C^{-1}, respectively. Thus, assuming an upper mantle mineralogy dominated by olivine, α is probably in the range of 3.0 to 4.0×10^{-5}°C^{-1}. To estimate ΔT both an estimate of the asthenosphere temperature and the crustal temperature increment is required. Lachenbruch and Sass (1978) suggest that the asthenosphere temperature can be estimated from the basalt dry solidus (BDS, Fig. 4) which they approximate by $1050 + 3z$ °C, where z is depth in km. Using this approximation, the asthenosphere temperature ranges from about 1200°C at 50 km to 1650°C at 200 km. The crustal temperature increment due to crustal heat production and conductivity variations will be laterally variable as these parameters vary, but for average crustal parameters (e.g., see Morgan, 1984), will probably be about 150°C. Thus, ΔT is expected to be

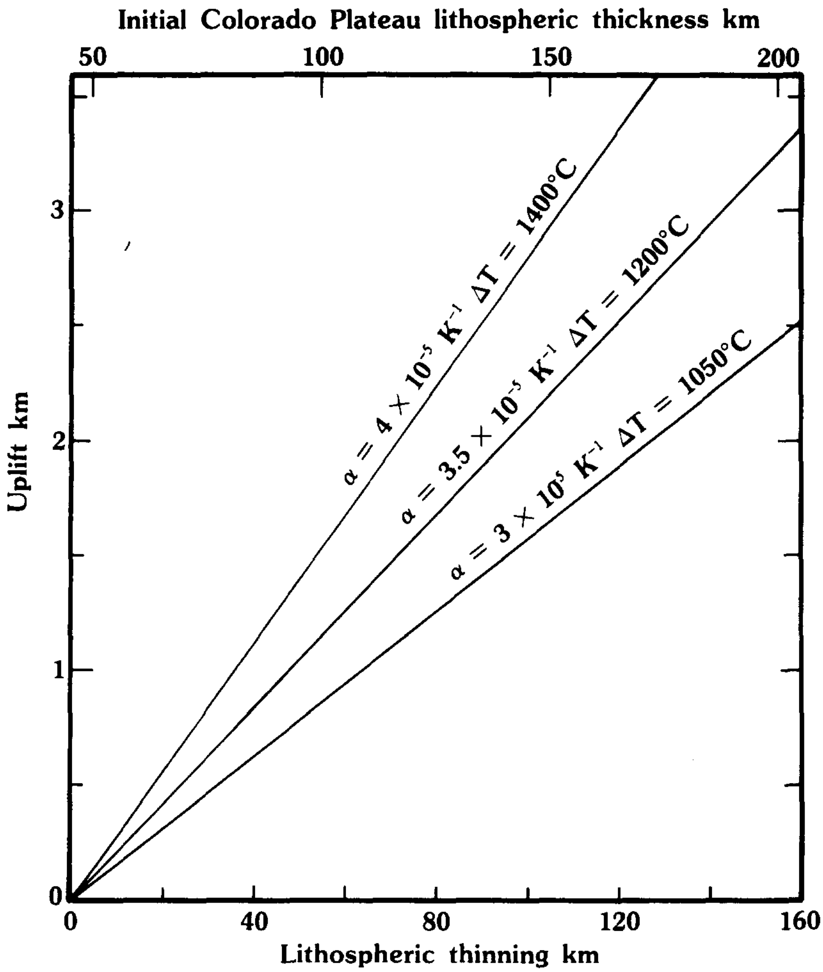

Fig. 5. Uplift due to thermal expansion in the lithosphere as a function of the amount of lithospheric thinning, and the potential uplift of Colorado Plateau as a function of its initial lithospheric thickness assuming that thinning continues to the Moho (45 km). Curves are shown for different values of the volume coefficient of expansion, α, and the effective lithospheric temperature drop, ΔT (see text for definition).

within the range 1050 to 1400°C. Uplift as a function of lithospheric thinning is shown in Figure 5 for these ranges of α and ΔT. If 2 km of uplift of the Colorado Plateau is due to simple thermal expansion, between 70 and 130 km of lithospheric thinning, or an equivalent heating of the crust and mantle down to the depth of isostatic compensation is suggested.

The second group of uplift mechanisms suggested for the Colorado Plateau invokes crustal thickening, either tectonic or magmatic. If temperature changes associated with the process of crustal thickening are

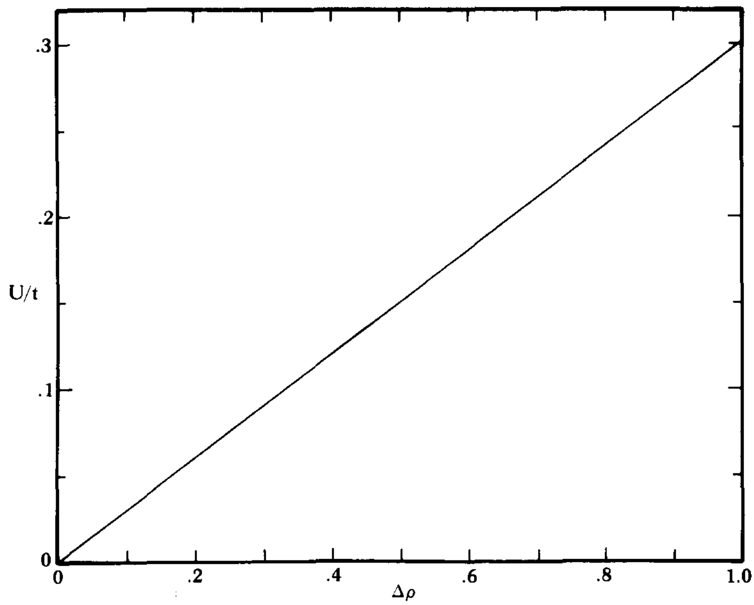

Fig. 6. Ratio of uplift, U, to crustal thickening or thickness of layer affected by a phase change, t, as a function of the density contrast at the Moho or the density change during the phase change, $\Delta\rho$, assuming a mantle or pre-phase change density of 3.3 Mg m^{-3}.

ignored, and assuming that the crust is in isostatic equilibrium both before and after thickening, it is easily shown that the ratio of the uplift, U, to crustal thickening, t, is given by

$$U/t = \Delta\rho/\rho_m \tag{3}$$

where $\Delta\rho$ is the density contrast between the crust and the uppermost mantle, and ρ_m is the density of the uppermost mantle. This ratio is plotted as a function of $\Delta\rho$ in Figure 6 for a value of ρ_m of 3.3 Mg m^{-3}. Seismic and gravity data indicate that $\Delta\rho$ is probably in the range 0.3 to 0.5 Mg m^{-3} (e.g., Drake and others, 1959; Woollard, 1966) which suggests that 13 to 22 km of crustal thickening is required if 2 km of uplift is caused solely by this mechanism.

Uplift caused by phase change in the lithosphere, either anhydrous or associated with hydration, can be calculated in the same manner as uplift due to crustal thickening, again assuming isostatic equilibrium before and after uplift and ignoring any thermal effects. In fact, if the phase change results in the conversation of mantle to crustal material, i.e., the Moho is a phase change, the phase change results in crustal thickening, and equation 3 applies directly as before, t being the thickness of the layer affected by the phase change and $\Delta\rho$ the density change during the phase change. If the phase change does not result in an interchange between the crust and the mantle, $\Delta\rho$ is the density change during the phase change, ρ_m is the density of the zone affected by the phase change prior to the phase change, and t is the thickness of the zone affected by the phase change. The ratio of uplift to thickness affected by the phase change as a function of density change during the phase change is again shown by the plot in Figure 6. If the Moho is a phase change, $\Delta\rho$ is probably in the range 0.3 to 0.5 Mg m^{-3} as before, and 2 km of uplift requires 13 to 22 km of mantle to be converted to crust by the phase change. Intra-crustal or intra-mantle phase changes are likely to be associated with smaller density changes than the Moho. Thus, if 2 km of uplift is the result of a phase change not associated with the Moho, the thickness of the layer affected by the phase change would probably be in excess of 25 km.

DISCUSSION OF UPLIFT MECHANISMS

As shown above, most of the uplift mechanisms suggested for the Colorado Plateau listed in Table I are physically capable of producing 2 km of uplift. However, the time scales required for some of the mechanisms and manifestations of the mechanisms other than uplift are not always compatible with the implications of available geological and geophysical data.

Low upper mantle seismic velocities and high electrical conductivity suggest that the upper mantle is anomalously hot beneath the Plateau, and thus some thermal expansion is to be expected if the pre-uplift Plateau lithosphere was relatively cold. For net uplift, thermal expansion must occur in the lithosphere underlying the Plateau in addition to any density effects in a postulated underlying remnant subducted slab (e.g., Damon, 1979). Thermal relaxation of a subducted slab cannot cause a net uplift on its own as it can only reverse subsidence caused by the original emplacement of the cool subducted slab.

Bird (1979) predicted about 1 km of uplift of the Plateau by thinning of the lithosphere from about 100 to 40 km by delamination. If the original lithospheric thickness was greater than 100 km, more uplift might be expected. However, dioritic magmatism in the extension of the Colorado Mineral Belt suggests temperatures of 700 to 900°C at depths of about 16 to 33 km at least locally prior to any uplift (D. Gust, pers. comm., 1984), indicating that the Plateau lithosphere was not very thick and cool prior to uplift, and thus the 1 km uplift predicted by Bird may be an upper rather than a lower limit to uplift by thermal expansion. Bird suggested two delamination events to explain mid-Tertiary and Pliocene to Recent uplift and volcanism, but the magnitude of the second predicted uplift is considerably less than that deduced from observations (Fig. 3), and there is poor correspondence between the delamination events and the timing of volcanism (Fig. 1 and 2). Bodell and Chapman (1982) predict nearly 2 km of uplift associated with unspecified thermal thinning of the lithosphere from 120 to 80 km, but most of this uplift comes not from thermal expansion but from the assumption of a constant density contrast between the lithosphere and asthenosphere. Unless the base of the lithosphere is associated with a phase or compositional change, this assumption is almost certainly incorrect. The lithosphere-asthenosphere density contrast assumed by Bodell and Chapman (1982) of 0.1 $Mg\,m^{-3}$ is about twice that predicted from reasonable thermal expansion considerations, and their model does not meet the constraints on uplift timing especially the young uplift event (Fig. 3).

The results of a refined version of the thermal thinning model of the Plateau lithosphere with resulting uplift and increased surface heat flow are shown in Figure 7. This model is based upon the analytical solution for a moving plane heat source in which the lithosphere is assumed to thin in response to a constant increase in heat flux at its base (Appendix; see also Morgan, 1983; Crough and Thompson, 1976; Wendlandt and Morgan, 1982; Spohn and Schubert, 1983; and others). This type of model can explain the uplift of the Plateau and apparently high mantle temperatures but moderate surface heat flow, but probably cannot explain more than 1 km of uplift of the Plateau, and again does not satisfy the constraints on uplift timing. Thus, while thermal expansion may make an important contribution to uplift of the Colorado Plateau we conclude that an additional uplift mechanism is also required, especially with respect to very young uplift of the Plateau.

The thickness of about 45 km for the crust of the Colorado Plateau is not anomalously thick for its present elevation, but it is considerably thicker than crust generally at sea level (e.g., see crustal thickness maps of the U.S. in Allenby and Schnetzler, 1983, and L. Braile and G. R. Keller, pers. comm., 1984). Thus it is probable that unless sea level was at least several hundred meters higher during the deposition of Cretaceous marine shales on the Plateau, the Plateau crust was thinner at this time. Minor crustal thickening (up to a few km) probably occurred during Late Cretaceous-early Tertiary crustal shortening and subsequent deposition on the Plateau. However, if modern crustal thicknesses on the eastern and southern margins of the U.S. can be used as a guide to the Plateau crustal thickness when its surface was at sea level, it is not unreasonable that its thickness has increased by about an additional 10 km during Cenozoic. This increase in thickness would explain about 1 km of uplift in addition to the uplift by thermal expansion.

Crustal thickening can occur by crustal shortening or underthrusting, magmatic underplating or intrusion, and by a phase change at the Moho.

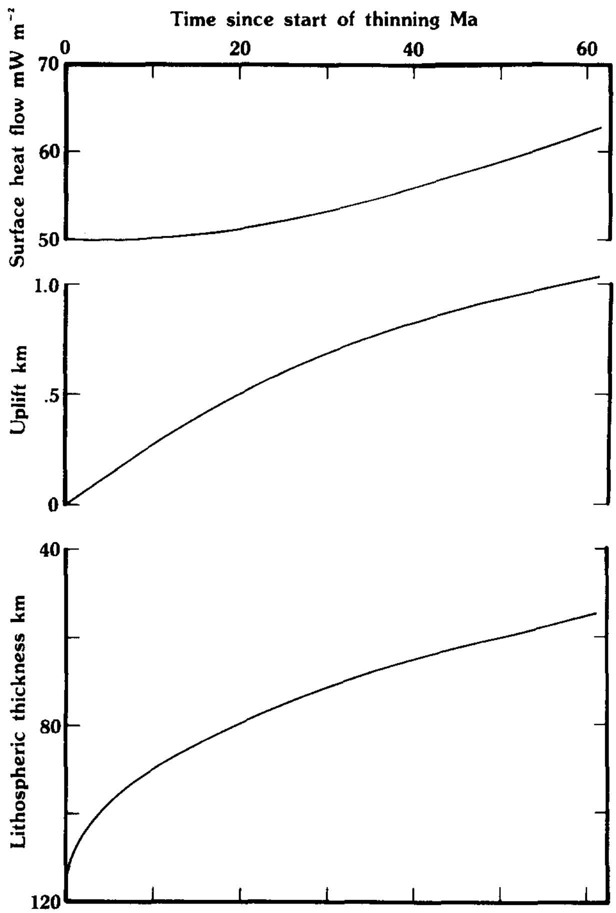

Fig. 7. Example of uplift and surface heat flow predicted by moving plane source model of lithospheric thinning (see text). Model parameters: initial surface heat flow - 50 mW m^{-2}; surface radiogenic heat production - 2 μW m^{-3}; heat production depth parameter 10 km; heat production depth distribution - average of exponential decrease with depth and constant heat production to 10 km; crustal thickness - 40 km; crustal thermal conductivity - 2.5 W m^{-1} K^{-1}; mantle thermal conductivity - 3.4 W m^{-1} K^{-1}; volume coefficient of expansion - 3×10^{-5} K^{-1}; lithosphere-asthenosphere boundary temperature - 1200°C; initial lithospheric thickness - 115 km; increase in heat flux at base of lithosphere - 60 mW m^{-2}; mantle diffusivity - 1 mm^2 s^{-1}.

As there is little evidence of major Cenozoic crustal shortening on the Plateau, and uplift does not correlate with compressional events, crustal shortening and underthrusting are not thought to be important for the late uplift of the Plateau. There is no evidence for major low rigidity zones beneath the Plateau, and thus expansion accompanying partial melting is thought to be insignificant. Hydration in the upper mantle and/or lower crust were thought to be significant and the possible cause of uplift of the

Plateau by Hess (1954), McGetchin and Silver (1972) and Silver and McGetchin (1978). However, recent studies of lower crustal xenoliths from the mid-Tertiary diatremes in the center of the Plateau in conjunction with studies of seismic velocity profiles of the Plateau by Padovani and others (1982) suggest that the Plateau lithosphere has not undergone hydration at depths greater than 15 km, and that uplift is not the result of a hydration phase change. Temperature dependent phase changes at the Moho or elsewhere in the lithosphere are possible, but the lack of correlation of the Moho or other seismic-density discontinuities with the temperature and pressure conditions of known phase changes (e.g., Bullard and Griggs, 1961; Mareschal and others, 1982), suggest that these phase changes are of secondary importance if they occur. Magmatic thickening of the crust of the Plateau is not unreasonable, however, especially with respect to the importance of Cenozoic magmatism in the western U.S. (e.g., Lipman and others, 1972; Christiansen and Lipman, 1972). Cenozoic volcanism has been relatively minor volumetrically on the Plateau, and we speculate that this could be at least in part due to deep crustal intrusion rather than eruption of the magmas.

Other data which suggest the possible significance of the uplift of the Plateau by magmatic crustal thickening come from the adjacent provinces, the Basin and Range, Southern Rocky Mountains-Rio Grande rift, and Great Plains. There is evidence that all of these areas have experienced recent (about 5 Ma or younger) uplift (e.g., Axelrod and Bailey, 1976; Lucchitta, 1979; Seager and others, 1984) and thus at least the most recent phase of Plateau uplift may not be restricted to the Plateau itself. Recent models of the thermal evolution of the Rio Grande rift in northern New Mexico by Morgan and Golombek (1984) conclude that recent uplift in this area must be due at least in part to crustal thickening, and there is evidence of very young magmatic intrusion in this area and the adjacent Plateau from seismic, heat flow and geomagnetic data (e.g., Sanford, 1983; Cook and others, 1978; Reiter and Clarkson, 1983b; Decker and others, 1984; Towle, 1980). Thus there is evidence that at least the latest phase of uplift of the Colorado Plateau should not be studied in isolation of the surrounding provinces.

Until further data are available, especially with respect to the timing of plateau uplift, the timing and significance of volcanism on the Plateau, and details of the structure of the deep crust and upper mantle, it will be very difficult to reliably constrain uplift models of the Plateau. From the analysis of the physical processes given above, and consideration of the available geological and geophysical data, it seems likely that at least two, possibly related, mechanisms are responsible for the uplift. Thermal expansion is a reasonable mechanism to account for about 1 km of uplift of the Plateau, although the details of this thermal expansion, lithospheric thinning by sub-lithospheric heating or lithospheric delamination, are poorly constrained. It seems likely that heating and thermal expansion is not yet complete as surface heat flow in the center of the Plateau is not yet high. Crustal thickening is probably the additional uplift mechanism, again accouting for about 1 km of uplift. Magmatic crustal thickening seems the most reasonable mechanism, and may not be restricted to the Plateau.

TECTONIC STABILITY OF THE COLORADO PLATEAU

Perhaps the most remarkable feature of the Cenozoic evolution of the Colorado Plateau has been its stability relative to the adjacent Basin and Range and Southern Rocky Mountain-Rio Grande rift tectonic provinces. It should be emphasized that the Plateau has not escaped Cenozoic tectonic deformation, e.g., early Cenozoic compressive warping and late Cenozoic faulting on its margins, but it has been deformed much less than adjacent crust. As discused above, Cenozoic uplift of the Plateau seems to require at least some heating of the Plateau lithosphere, and if the Plateau lithosphere was relatively cool at the end of the Mesozoic, strength in this lithosphere seems the most reasonable explanation for the relative resistance of the Plateau to Late Cretaceous-early Tertiary Laramide deformation. Studies of lithospheric strength as a function of the geotherm (e.g., Lynch, 1983; Smith and Bruhn, 1984; Morgan and Golombek, 1984) indicate that the lithosphere becomes considerably weaker as it is heated and thinned.

Until we have better definition of the uplift and magmatic history of the Plateau, it is difficult to reliably constrain geotherms and strength curves for different stages in the Cenozoic evolution of the Plateau relative to the adjacent provinces. However, as the modern Plateau has lower surface heat flow in its interior than the adjacent provinces, and Cenozoic volcanism has generally been much less abundant on the Plateau than in the adjacent provinces, it is not unreasonable to speculate that it has been cooler and thus stronger than the adjacent provinces. An example of the simplified hypothetical lithospheric strength curves for the mid-Tertiary Plateau, the modern Great Plains, and various stages in the evolution of southwestern New Mexico are shown in Figure 8. As the Plateau lithosphere has been heated in the Neogene and Quaternary, however, it has probably become weaker, especially in the Plateau margins where young volcanism is common. Thus, part of the reason for the relative stability of the Plateau may have been that it was cooler than the surrounding provinces, but heating of the Plateau may be causing it to lose strength.

In addition to lithospheric strength, local topography may also have contributed to stability of the Plateau. After early Tertiary compressive defor-

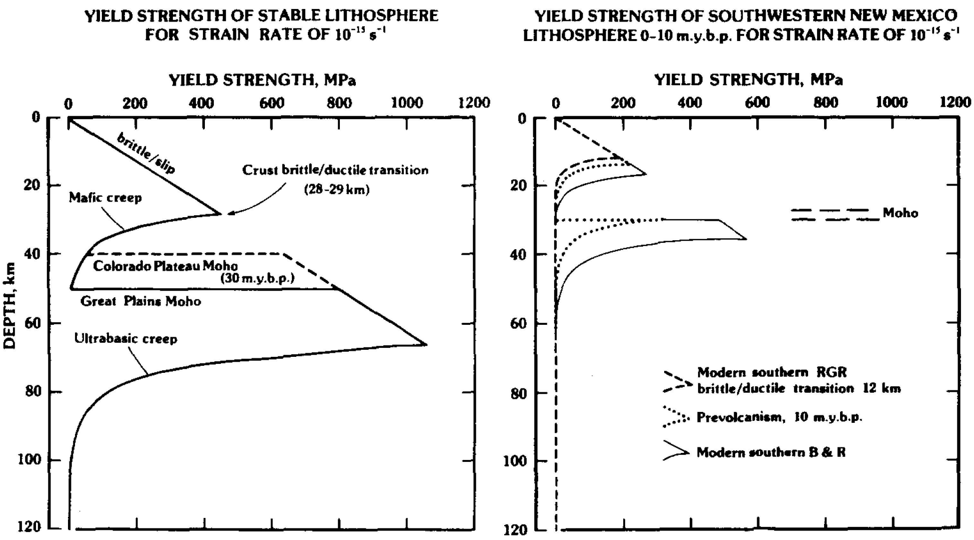

Fig. 8. Hypothetical lithospheric strength curves for the modern Great plains and mid-Tertiary Colorado Plateau (left) and various stages in the evolution of the southwestern New Mexico (right) (adapted from Morgan and Seager, 1983). Strength curves are for an extensional stress field and were generated using the deformation parameters of Lynch (1983). Great Plains and mid-Tertiary (30 m.y.b.p.) Colorado Plateau strength curves were generated assuming a surface heat flow of 50 mW m^{-2}. Southwestern New Mexico strength curves were generated assuming heat flow of 125 mW m^{-2} for the modern southern Rio Grande rift (RGR), 85 mW m^{-2} for the modern Basin and Range (B&R), and 90 mW m^{-2} for the lithosphere at 10 Ma (m.y.b.p.), prior to the most recent episode of volcanism in this area. It is thought that the geotherm in this area was probably similar to the modern rift geotherm during the Late Oligocene and early Miocene, and thus there was a strong strength contrast at this time between the Colorado Plateau and the adjacent provinces. Additional examples of strength curves calculated for this region and further discussion of the geotherm evolution are given by Morgan and Golombek (1984).

mation, the western U.S. has experienced primarily extensional tectonism. Prior to about 10 Ma the Plateau was topographically low with respect to the adjacent Basin and Range province, and this topography would have generated a compressional stress field at least in the uppermost crust of the Plateau (e.g., see Bott, 1981; Crough, 1983). At about 10 Ma when the Plateau first rose above the adjacent Basin and Range, this stress field would have been reversed, especially in the Plateau margins, as illustrated in Figure 9. This reversal of the locally generated stress field, combined with the effects of the regional stress field and heating of the Plateau margins is probably responsible for the development of young faults on the periphery of the Plateau, and if these trends continue, may mark the initiation of breakup of the Plateau.

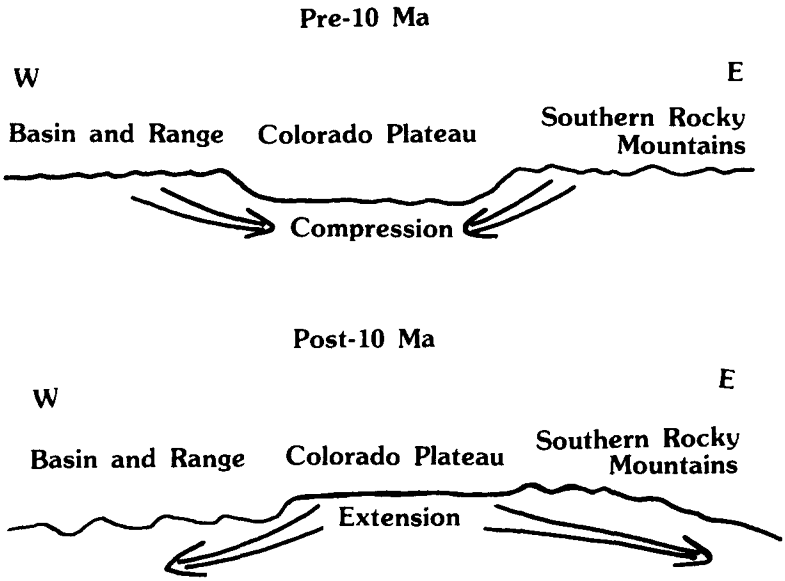

Fig. 9. Cartoon to illustrate the change in relative altitudes and related stress fields between the Colorado Plateau and the adjacent Basin and Range and Southern Rocky Mountain provinces at about 10 Ma.

CONCLUDING REMARKS

No simple single-event model seems able to explain the Cenozoic uplift of the Colorado Plateau and satisfy the available geological and geophysical constraints on the geological evolution of the Plateau. It seems probable that the uplift is related at least in part to a thermal process, and we estimate that about half the uplift has been caused by thermal expansion related to lithospheric heating and thinning, and that about half the uplift could have been caused by crustal thickening, probably primarily magmatic thickening. There is little definitive data to constrain the timing of these events, but the young uplift of areas adjacent to the Plateau concurrent with plateau uplift suggests that the most recent uplift has at least a component of crustal thickening, and that this process is not restricted to the Plateau.

The source of the relatively stability of the Plateau is probably related primarily to its thermal evolution. While it has been cooler than the surrounding regions it has been stronger, and this effect may have been amplified by further heating of the adjacent provinces during their deformation. Heating and uplift of the Plateau during the Cenozoic have both probably contributed to a reduction in the tectonic stability of the Plateau. However, before this problem, and the problems of Plateau uplift can be addressed in greater detail, additional geological and geophysical data are required especially with respect to the timing of uplift and magmatic activity in the Plateau, and the structure of the lower crust and upper mantle of the Plateau.

ACKNOWLEDGMENTS

D. Gust is thanked for several useful discussions about the significance of Cenozoic volcanism on the Colorado Plateau. We thank an anonymous reviewer for constructive criticism of a draft of this manuscript. P.M. made his contribution to this study while at the Lunar and Planetary Institute, which is operated by the Universities Space Research Association under contract no. NASW-3389 from the National Aeronautics and Space Administration. This is LPI contribution no. 561.

REFERENCES

Allenby, R. J. and Schnetzler, C. C., 1983. United States crustal thickness, Tectonophysics, 93: 13–31.

Armstrong, R. L., 1969. K-Ar dating of laccolithic centers of the Colorado Plateau and vicinity, Bull. Geol. Soc. Am., 80: 2081–2086.

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

Bird, P., 1979. Continental delamination and the Colorado Plateau, J. Geophys. Res., 84: 7561–7571.

Bodell, J. M. and Chapman, D. S., 1982. Heat flow in the north-central Colorado Plateau, J. Geophys. Res., 87: 2869–2884.

Bott, M. H. P., 1981. Crustal doming and the mechanism of continental rifting, Tectonophysics, 73: 1–8.

Bullard, E. C. and Griggs, D. T., 1961. The nature of the Mohorovicic discontinuity, Geophys. J. Roy. Astro. Soc., 6: 118–123.

Christiansen, R. L. and Lipman, P. W., 1972. Cenozoic volcanism and 0late-tectonic evolution of the Western United States. II. Late Cenozoic, Phil. Trans. R. Soc. Lond. A., 271: 249–284.

Cook, F. A., Decker, E. R. and Smithson, S. B., (1978) Preliminary transient heat flow model of the Rio Grande rift in southern New Mexico, Earth Planet. Sci. Lett., 40: 316–326.

Crough, S. T., 1979. Hotspot epeirogeny, Tectonophysics, 61: 321–333.

Crough, S. T., 1983. Rifts and swells: geophysical constraints on causality, Tectonophysics, 94: 23–37.

Crough, S. T. and Thompson, G. A., 1976. Numerical and approximate solutions for lithospheric thickening and thinning, Earth Planet. Sci. Lett., 31: 397–402.

Damon, P. E., 1979. Continental uplift at convergent boundaries, Tectonophysics, 61: 307–319.

Damon, P. E., 1983. Continental uplift, compensation and shunting during trench-spreading center collision, Tectonophysics, 99: T1–T8.

Decker, E. R., Bucher, G. J., Buelow, K. L. and Heasler, H. P., 1984. Heat flow and radioactivity in the Rio Grande rift zone, central and northern Colorado, New Mexico Geological Society Guidebook, 35: 45–50

Drake, C. L., Ewing, M. and Sutton, G. H., 1959. Continental margins and geosynclines, Phys. Chem. Earth, 3: 110–198.

Dutton, C. E., 1880. Report of geology of the high plateau of Utah, Geophys. and Geol. Surv. of the Rocky Mt. Reg., Washington, D.C., 307 pp.

Dutton, C. E., 1882a. The physical geology of the Grand Canyon district, U.S. Geol. and Geog. Survey Terr. 2nd Ann. Rept.

Dutton, C. E., 1882b. The Tertiary history of the Grand Canyon district, U.S. Geol. Survey Mon. 2.

Dutton, C. E., 1885. Mt. Taylor and the Zuni Plateau, U.S. Geol. and Geog. Survey Terr. 6th Ann. Rept., p. 165–182.

Dutton, C. E., 1892. On some of the greater problems of physical geology, Philos. Soc. Washington, 11: 51–64.

Eardley, A. J., 1962. Structural Geology of North America, 2nd edn., Harper and Row, New York, 743 pp.

Gilluly, J., 1963. The tectonic evolution of the western United States, Quart. J. Geol. Soc. Lond., 119: 113–174.

Green, D. H. and Ringwood, A. E., 1967. An experimental investigation of the gabbro to eclogite transformation and its petrological applications, Geochim. Cosmochim. Acta., 31: 767–833.

Hack, J. T., 1942. Sedimentations and volcanism in the Hopi Buttes, Arizona, Geol. Soc. Am. Bull., 53: 335–372.

Hamblin, W. K., Damon, P. E. and Bull, W. B., 1981. Estimates of vertical crustal strain rates along the western margins of the Colorado Plateau, Geology, 9: 293–298.

Helmsteadt, H. and Schulze, D. J., 1978. Petrologic constraints for upper mantle models of the Colorado Plateau (abs). In: Papers Presented to the Conference of Plateau Uplift: Mode and Mechanism, Lunar and Planetary Institute, Houston, 16–18.

Hess, H. H., 1954. Serpentines, orogeny and epeirogeny. In: Crust of the Earth–a Symposium, Poldervaart, A., ed., Geol. Soc. Am. Spec. Pap. 62: 391–408.

Hess, H. H., 1962. History of ocean basins. In: Petrologic Studies: a Volume in Honor of A. F. Buddington, Geological Society of America, Boulder, Colorado, 599–620.

Hunt, C. B., 1956. Cenozoic geology of the Colorado Plateau, U.S. Geol. Surv. Prof. Pap. 279, 99 pp.

Hunt, C. B., 1976. Grand Canyon and the Colorado River, their geologic history. In: Geology of the Grand Canyon, Breed, W. J. and Roat, E., eds., Museum of Natural History, Flagstaff, Arizona, 129–141.

Keller, G. R., Braile, L. W. and Morgan, P., 1979. Crustal structure, geophysical models and contemporary tectonism of the Colorado Plateau, Tectonophysics, 61: 131–147.

Kelley, V. C., 1955. Regional tectonics of the Colorado Plateau and relationship to origin and distribution of uranium, Univ. New Mexico Publ. Geol. 5, 120 pp.

Kelley, V. C., 1979. Tectonics of the Colorado Plateau and new interpretation of its eastern boundary, Tectonophysics, 61: 97–102.

Lachenbruch, A. H. and Sass, J. H., 1978. Models of an extending lithosphere and heat flow in the Basin and Range province. In: Cenozoic Tectonics and Regional Geophysics of the western Cordillera, Smith, R. B. and Eaton, G. P., eds., Geol. Soc. Am. Memoir 152: 209–250.

Lipman, P. W., Prostka, H. J. and Christiansen, R. L., 1972. Cenozoic volcanism and Plate-tectonic evolution of the Western United States. I. Early and Middle Cenozoic, Phil. Trans. R. Soc. Lond. A, 271, 217–248.

Lovering, J. F., 1958. The nature of the Mohorovicic discontinuity, Trans. Am. Geophys. Un., 39: 947–955.

Luedke, R. G. and Smith, R. L., 1978a. Map showing distribution, composition, and age of late Cenozoic volcanic centers in Arizona and New Mexico, U.S. Geol. Surv. Misc. Invest. Ser. Map I-1091-A, 1:1,000,000.

Luedke, R. G. and Smith, R. L., 1978b. Map showing distribution, composition, and age of late Cenozoic volcanic centers in Colorado, Utah, and southwestern Wyoming, U.S. Geol. Surv. Misc. Invest. Ser. Map I-1091-B, 1:1,000,000.

Lucchitta, I., 1979. Late Cenozoic uplift of the southwestern Colorado Plateau and adjacent lower Colorado River region, Tectonophysics, 61: 63–95.

Lynch, H. D., 1983. Numerical models of the formation of continental rifts by processes of lithospheric necking, Ph.D. Dissertation, New Mexico State University, Las Cruces, New Mexico, 290 pp.

Mareschal, J.-C., 1981. Uplift by thermal expansion of the lithosphere, Geophys. J. R. Astro. Soc., 66: 535–552.

Mareschal, J.-C., Gangi, A. F. and Lamping, N. L., 1982. The Moho as a phase change: a test of the hypothesis, J. Geophys. Res., 87: 4723–4730.

Mayer, L., 1979. Evolution of the Mogollon rim in central Arizona, Tectonophysics, 61: 49–62.

McGetchin, T. R. and Silver, L. T., 1970a. Compositional relations in minerals from kimberlite and related rock in the Moses Rock dike, San Juan County, Utah, Amer. Mineral., 55: 1738–1771.

McGetchin, T. R. and Silver, L. T., 1970b. A crustal-upper mantle model for the Colorado Plateau based on observations of crystalline rock fragments in a kimberlite dike, Phys. Earth Planet. Int., 3: 471.

McGetchin, T. R. and Silver, L. T., 1972. A crustal-upper mantle model for the Colorado Plateau based on observations of crystalline rock fragments in the Moses Rock dike, J. Geophys. Res., 77: 7022–7037.

McGetchin, T. R., Nikhanj, Y. S. and Chodos, A. A., 1973. Carbonatite-kimberlite relations in the Cane Valley Diatreme, San Juan County, Utah, J. Geophys. Res., 78, 1854–1869.

McGetchin, T. R., Burke, K. C., Thompson, G. A. and Young, R. A., 1980. Mode and mechanism of plateau uplifts. In: Dynamics of Plate Interiors, Geodyn. Ser., v. 1, A. W. Bally, P. L. Bender, T. R. McGetchin and R. I. Walcott, eds., American Geophysical Union, Washington, D.C., and Geological Society of America, Boulder, Colorado, 99–110.

Melosh, H. J. and Ebel, J., 1979. A simple model for thermal instability in the asthenosphere, Geophys. J. R. Astro. Soc., 59: 419–436.

McKenzie, D., 1984. A possible mechanism for epeirogenic uplift, Nature, 307: 616–618.

Morgan, P., 1983. Constraints on rift thermal processes from heat flow and uplift, Tectonophysics, 94: 277–298.

Morgan, P., 1984. The thermal structure and thermal evolution of the continental lithosphere, Phys. Chem. Earth, 15: 107–193.

Morgan, P. and Golombek, M. P., 1984. Factors controlling the phases and styles of extension in the northern Rio Grande rift, New Mexico Geological Society Guidebook, 35: 13–19.

Morgan, P. and Phillips, R. J., 1983. Hot spot heat transfer: its application to Venus and implications to Venus and Earth, J. Geophys. Res., 88: 8305–8317.

Morgan, P. and Seager, W. R., 1983. Thermal, mechanical and tectonic evolution of the southern Rio Grande rift (abs), Geol. Soc. Am. Abs. w. Prog., 15, 320.

Naeser, C. W., 1971. Geochronology of the Navajo-Hopi diatremes, Four Corners area, J. Geophys. Res., 76: 4978–4985.

O'Connell, R. J. and Wasserburg, 1967. Dynamics of the motion of a phase change boundary to changes in pressure, Rev. Geophys., 5: 329–410.

Oetking, P., Feray, D. E. and Renfro, H. B., compilers, 1967. Geological highway map of the Southern Rocky Mountain region, U.S. Geol. Highway Map Ser., Map No. 2, American Association of Petroleum Geologists, Tulsa, Oklahoma, approx. 1:1,900,800.

Otton, J. K. and Brooks, W. E., 1978. Tectonic history of the Colorado Plateau margin, Date Creek Basin and adjacent areas, west-central Arizona (abs). In: Papers Presented to the Conference on Plateau Uplift: Mode and Mechanism, Lunar and Planetary Institute, Houston, 31–33.

Padovani, E. R., Hall, J. and Simmons, G., 1982. Constraints on crustal hydration below the Colorado Plateau from V_p measurements on crustal xenoliths, Tectonophysics, 84: 313–328.

Pakiser, L. C., 1963. Structure of the crust and upper mantle in the western United States, J. Geophys. Res., 68: 5747–5756.

Pedersen, J. and Hermance, J. F., 1981. Deep electrical structure of the Colorado Plateau as determined from magnetotelluric measurements, J. Geophys. Res., 86: 1849–1857.

Peirce, H. W., Damon, P. E. and Shafiquallah, M., 1979. An Oligocene (?) Colorado Plateau edge in Arizona, Tectonophysics, 61: 1–24.

Reiter, M. and Clarkson, G., 1983a. A note on terrestrial heat flow in the Colorado Plateau, Geophy. Res. Lett., 10: 929–932.

Reiter, M. and Clarkson, G., 1983b. Geothermal studies in the San Juan basin and Four Corners area of the Colorado Plateau II. Steady-state models of the thermal source of the San Juan volcanic field, Tectonophysics, 91: 253–269.

Rowley, P. D., Anderson, J. J., Williams, P. L. and Fleck, R. J., 1978a. Age of structural differentiation between the Colorado Plateau and Basin and Range Provinces in southwest Utah, Geology, 6: 51–55.

Rowley, P. D., Anderson, J. J., Williams, P. L. and Fleck, R. J., 1978b. Reply, Geology, 6: 572–575.

Roy, R. F., Beck, A. E. and Touloukian, Y. S., 1981. Thermophysical properties of rocks. In: Physical Properties of Rocks and Minerals, Touloukain, Y. S., Judd, W. R. and Roy, R. F., eds., McGraw-Hill, New York, 503–548.

Sanford, A., 1983. Magma bodies in the Rio Grande rift in central New Mexico, New Mexico Geological Society Guidebook 34, 123–125.

Seager, W. R., Shafiqullah, M., Hawley, J. W. and Marvín, R. F., 1984. New K-Ar dates from basalts and the evolution of the southern Rio Grande rift, Geol. Soc. Am. Bull., 95: 87–99.

Shaw, H. R. and Jackson, E. D., 1973. Linear island chains in the Pacific: result of thermal plumes or gravitational anchors, J. Geophys. Res., 78: 8634–8652.

Shoemaker, E. M., 1978. The Colorado Plateau problem, Conference on Plateau Uplift: Mode and Mechanism, Flagstaff, Arizona, 14–16 August, 1978, oral presentation.

Silver, L. T. and McGetchin, T. R., 1978. The nature of the basement beneath the Colorado Plateau and some implications for plateau uplifts (abs). In: Papers Presented to the Conference on Plateau Uplift: Mode and Mechanism, Lunar and Planetary Institute, Houston, 45–46.

Smith, R. B. and Bruhn, R. L., 1984. Intraplate extensional tectonics of the eastern Basin-Range: Inferences on structural style from seismic data, regional tectonics and thermal-mechanical models of brittle/ductile deformation, J. Geophys. Res., 89, 5733–5762.

Snyder, W. S., Dickinson, W. R. and Silberman, M. L., 1976. Tectonic implications of space-time patterns of magmatism in the western United States, Earth Planet. Sci. Lett., 32: 91–106.

Spohn, T. and Schubert, G., 1983. Convective thinning of the lithosphere: a mechanism for rifting and mid-plate volcanism on Earth, Venus, and Mars, Tectonophysics, 94: 67–90.

Towle, J. N., 1980. New evidence for magmatic intrusion beneath the Rio Grande rift, New Mexico, Geol. Soc. Am. Bull., 91: 626–630.

Thompson, G. A. and Zoback, M. L., 1979. Regional geophysics of the Colorado Plateau, Tectonophysics, 61: 149–181.

Watson, K. D., 1967. Kimberlite pipes of northeastern Arizona. In: Ultramafic and Related Rocks, Wyllie, P. J., ed., John Wiley, New York, 261–269.

Wendlandt, R. F. and Morgan, P., 1982. Lithospheric thinning associated with rifting in East Africa, Nature, 298: 734–736.

Williams, H., 1936. Pliocene volcanoes of the Navajo-Hopi country, Geol. Soc. Am. Bull., 47: 111–172.

Woollard, G. P., 1966. Regional isostatic relations in the United States. In: Earth Beneath the Continents, Steinhart, J. S. and Smith, T. J., eds., Geophys. Mono. 10, American Geophysical Union, Washington, D.C., 557–594.

Young, R. A., 1979. Laramide deformation, erosion and plutonism along the southwestern margin of the Colorado Plateau, Tectonophysics, 61: 25–47.

APPENDIX

Analytical thermal model for lithospheric thinning

The thermal effects of lithospheric thinning in response to a constant increase in heat flux at the base of the lithosphere can be modeled using the analytical solution for a plane heat source in a moving medium (Morgan, 1983). In modeling lithospheric thinning, the medium (the lithosphere) is kept stationary, the plane heat source is moved and assumed to lie on the base of the lithosphere, its strength being equal to the constant increase in heat flux at the base of the lithosphere to be modeled. A moving image source is used to maintain the boundary condition of a stable temperature (arbitrary zero) at the Earth's surface. The geometry of the problem is illustrated in Figure 10.

If the lithosphere has an arbitrary temperature distribution (geotherm) prior to thinning given by $T_0(z)$, where z is depth, the temperature at any time during thinning is given by the superposition of the temperature perturbations due to the moving heat sources upon the original geotherm. Thus, using the notation in Figure 10, and using the solution given by Morgan (1983), the temperature T_p at any depth z within the lithosphere,

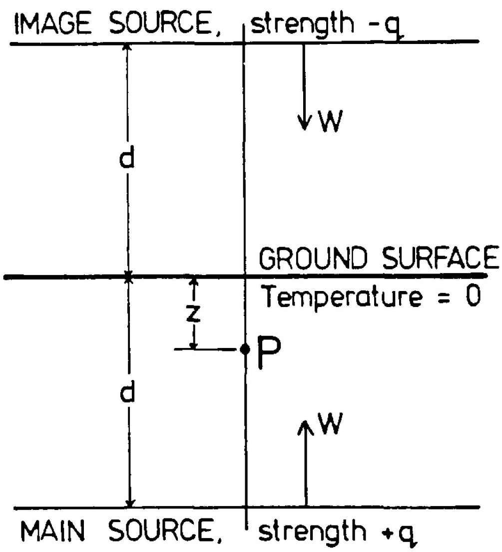

Fig. 10. (Appendix) Geometry used to compute the thermal perturbation due to a moving planar heat source analogous to thermal thinning of the lithosphere in response to a constant increase in heat flow at its base.

when the lithosphere has thickness d and is being thinned at a rate W, is given by:

$$T_p = T_0(z) + (q/\rho cW)\{\exp[-W(d-z)/k] - \exp[-W(d+z)/k]\} \qquad (A1)$$

where q is the plane heat source strength, per unit time, per unit area, and ρ, c and k are the density, specific heat and thermal diffusivity of the medium into which the heat is propagating. At the base of the lithosphere (on the heat source), the temperature T_b is given by setting $z=d$ in equation A1:

$$T_b = T_0(z) + (q/\rho cW)[1 - \exp(-2Wd/k)] \qquad (A2)$$

The condition for lithospheric thinning, however, requires that the temperature at the base of the lithosphere is equal to the asthenosphere temperature. Thus, if an asthenosphere temperature is assumed, the rate of lithospheric thinning for a particular lithospheric thickness can be determined for any basal heat flux increase, q, by solving equation A2 numerically for W. Once the thinning rate W has been determined, the instantaneous temperature for any point in the lithosphere can be determined using equation A1.

The mean temperature increase above the base of the lithosphere is of interest for calculations of thermal expansion within the lithosphere and resulting uplift. This mean temperature increase T_{inc} is given by the integral

$$\begin{aligned} T_{inc} &= 1/d \int_0^d [T_p - T_0(z)]\,dz \\ &= qk[1 + \exp(-2Wd/k) - 2\exp(-Wd/k)]/\rho cW^2 d) \qquad (A3) \end{aligned}$$

If no mass is lost from the lithospheric/asthenospheric column during thinning, surface uplift can be calculated from the thermal expansion in the lithosphere due to its mean temperature increase (see also Morgan, 1983).

Surface heat flow during thinning can be calculated from the surface geothermal gradient calculated by differentiating equation A1 and setting $z=0$. This gradient G_s is given by:

$$G_s = dT_0(z)/dz + (2q/k\rho c)\exp(-Wd/k) \qquad (A4)$$

By integrating the rates of lithospheric thinning derived from equation A2, lithospheric thickness as a function of time can be determined, with corresponding uplift and surface heat flow values calculated using equations A3 and A4.