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# Cenozoic Thermal, Mechanical and Tectonic Evolution of the Rio Grande Rift

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Careful documentation of the Cenozoic geologic history of the Rio Grande rift in New Mexico reveals a complex sequence of events. At least two phases of extension have been identified. An early phase of extension began in the mid-Oligocene (about 30 Ma) and may have continued to the early Miocene (about 18 Ma). This phase of extension was characterized by local high-strain extension events (locally, 50–100%, regionally, 30–50%), low-angle faulting, and the development of broad, relatively shallow basins, all indicating an approximately NE-SW ±25° extension direction, consistent with the regional stress field at that time. Extension events were not synchronous during early phase extension and were often temporally and spatially associated with major magmatism. A late phase of extension occurred primarily in the late Miocene (10–5 Ma) with minor extension continuing to the present. It was characterized by apparently synchronous, high-angle faulting giving large vertical strains with relatively minor lateral strain (5–20%) which produced the modern Rio Grande rift morphology. Extension direction was approximately E-W, consistent with the contemporary regional stress field. Late phase graben or half-graben basins cut and often obscure early phase broad basins. Early phase extensional style and basin formation indicate a ductile lithosphere, and this extension occurred during the climax of Paleogene magmatic activity in this zone. Late phase extensional style indicates a more brittle lithosphere, and this extension followed a middle Miocene lull in volcanism. Regional uplift of about 1 km appears to have accompanied late phase extension, and relatively minor volcanism has continued to the present. We have estimated geotherms and calculated lithospheric strength curves for the two phases of extension, using geologic data to constrain earlier events and geophysical data to constrain the modern geotherm and crustal structure. A high geotherm was deduced for early phase extension, resulting in a shallow crustal brittle-ductile transition and negligible mantle strength. The lithosphere cooled after early phase extension, resulting in a deeper crustal brittle-ductile transition, and, perhaps more significantly, a considerable zone of mantle strength immediately beneath the Moho. These results indicate that early phase extensional style was controlled by a crustal decollement near the brittle-ductile transition, which was prevented during late phase extension by significant strength in the uppermost mantle. Late Cenozoic uplift of the rift zone cannot be explained by crustal thinning during extension and geotherm evolution predicted from simple cooling. However, this uplift does not appear to be restricted to the rift zone, and Pliocene to Recent volcanism and heat flow data suggest that uplift may be caused by magmatic thickening of the crust, perhaps unrelated to rifting. The complex interrelationship among regional and local prerifting, synrifting, and postrifting events in the Rio Grande rift suggests that rifting, at least in this region, should not be considered in isolation of other geologic events.

## INTRODUCTION

Much attention has recently been focused upon the mechanisms and processes of rifting. In this study we present a relatively detailed analysis of two areas of New Mexico which exhibit complex but similar Cenozoic histories of extensional tectonism. The locations of these areas are shown in Figure 1. The first study area is the Basin and Range province and southern Rio Grande rift in southern New Mexico; the second study area is the central Rio Grande rift in central and northern New Mexico, the southern San Luis basin, the Española basin, and the Albuquerque-Belen basin. At least two Cenozoic extensional events have occurred in each area with contrasting tectonic styles. We have carefully documented these events with relevant geological and geophysical data to attempt to understand the factors which controlled tectonic style during each extensional event through an analysis of the thermal and mechanical evolution of the lithosphere in these areas. As these data are crucial to the thermomechanical modeling presented in this contribution, we present detailed documentation of the geologic history of the southern Rio Grande rift study area below, followed by a more concise history of rift evolution in northern New Mexico [see also *Morgan and Golombek*, 1984]. It is beyond the scope of this paper to present similarly detailed documentation for the Cenozoic evolution of all areas of the Rio Grande rift: however, the general similarities in evolutionary history that we have deduced for our two study areas suggest that our results may be generally applicable to the Rio Grande rift.

Our analysis of extension in the Rio Grande rift differs from many previous rift modeling studies in that we ignore the causes of extension. Through this analysis, we show that, at least in our study areas, the extension events cannot be understood in

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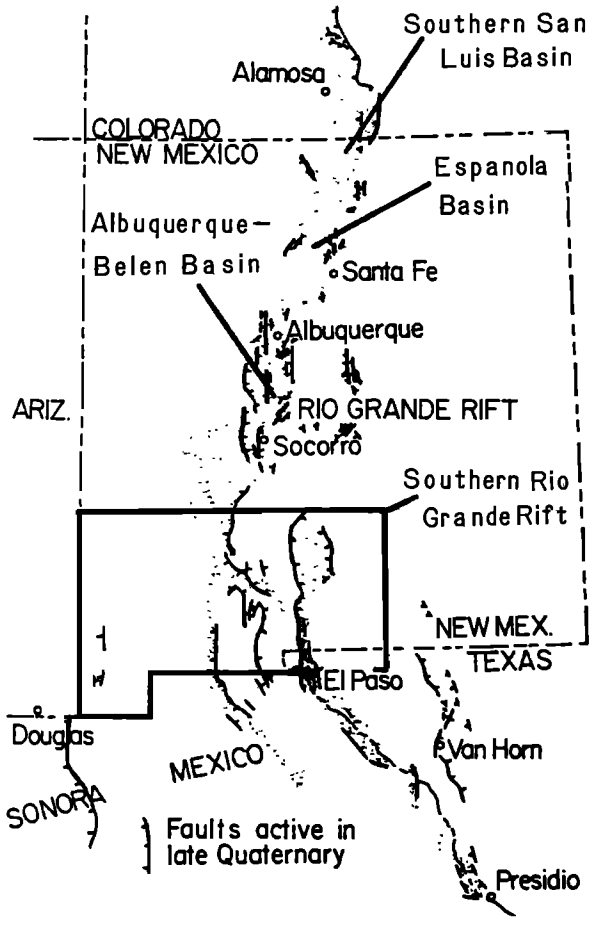


Fig. 1. Index map for study areas. The southern Rio Grande rift detailed study area is outlined by the box. The Albuquerque-Belen, Española and southern San Luis basins comprise the central Rio Grande rift study area (base map from *Seager and Morgan* [1979]).

isolation of other geologic events. We attempt to show that the interrelationship between regional and local events may be complex but that it is fundamental to the understanding of extensional tectonics. The interrelationship between local and regional events may be the key to understanding some of the complexities and diversities of continental rifting.

## CENOZOIC GEOLOGIC HISTORY OF SOUTHERN RIO GRANDE RIFT

### *Prerift Tectonics and Magmatism*

During the last 65 m.y., southwestern New Mexico has been the site of almost constant tectonic and/or magmatic activity (Figure 2). Most of southern New Mexico was a cratonic back arc region during the Laramide which failed under regional compression by forming a series of northwest trending, Wind River style basement block uplifts and complementary basins [*Seager*, 1983; *Seager and Mack*, 1986*a*]. Only in the southwestern corner of the state did a magmatic arc extend into the region, although isolated volcano-plutonic centers occur outside the arc at Tyrone, Pinos Altos, Santa Rita, and Hillsboro. All of the magmatism is Late Cretaceous to Paleocene in age.

Middle Tertiary magmatism affected virtually all of southwestern New Mexico. Although locally sheets of rhyolitic ash flow tuff are prominent, earliest eruptions were generally andesitic flows and lahars from widespread central volcanoes. By 35 Ma, calc-alkaline rhyolite and rhyodacite were erupted in huge volumes, largely as ash flow tuff sheets from cauldron complexes. Individual sheets spread across thousands of square kilometers of southwestern New Mexico, suggesting that neither active extension nor compression was producing topography that might control the distribution of volcanic rocks [e.g., *Chapin*, 1974; *Chapin and Seager*, 1975; *Seager et al.*, 1984).

*Elston* [1984*a*] and *Engebretson et al.* [1984] believe middle Tertiary arc magmatism was accompanied by strong interarc extension. On the other hand, *Price and Henry* [1984] and *Henry and Price*, [this issue] present data from west Texas that indicates ENE compression from Laramide time through the main pulse of middle Tertiary volcanism 39–32 Ma. Because of these contradictions, it is unclear what kind or orientation of stress field accompanied middle Tertiary volcanism in southwestern New Mexico. What does seem clear is that by about 30 ± 2 Ma changing patterns of magmatism, sedimentation, and deformation signify establishment of regional extension as a stable, dominating stress field.

### *Early Phase Extension*

*Magmatism.* Several kinds of evidence document the existence of an extensional regime in southern New Mexico in late Oligocene time. One of these is the striking and locally thick sequence of "basaltic andesite" that overlies middle Oligocene ash flow tuffs or associated rocks almost everywhere. Ranging in age from about 29 to 18 Ma, these mafic flows locally comprise central shield volcanoes or are associated with

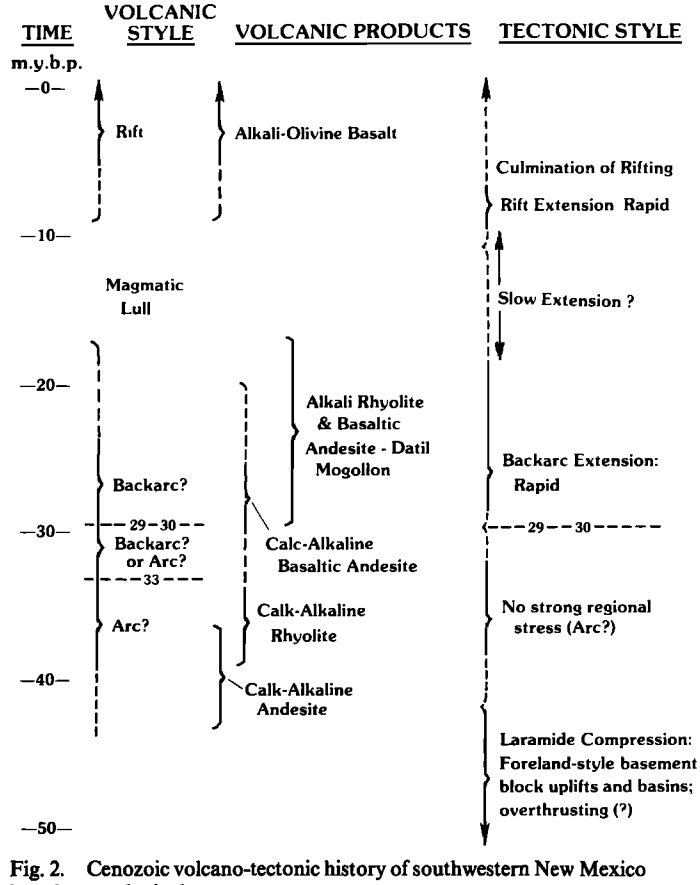


Fig. 2. Cenozoic volcano-tectonic history of southwestern New Mexico based on geologic data.

the remains of cinder cone fields or diatremes. At least one 27 Ma cinder cone was constructed on a throughgoing normal fault [*Seager and Clemons,* 1975], implying that extensional stresses were active at this time.

Although the basaltic andesites are generally dark colored, they range in composition from mafic andesite to latite and, like their predecessor tuffs and andesites, are calc-alkaline [*Chapin and Seager,* 1975; *Bornhorst,* 1980]. Strontium isotopic ratios are also similar to the older volcanics (0.706–0.709, [*Stinnett and Steuber,* 1976]). High silica (rarely peralkaline) rhyolitic lavas and tuffs are interbedded locally with the lavas, especially in the Mogollon-Datil Plateau where important ash flow tuff cauldrons of early Miocene age are well known [e.g., *Ratte et al.,* 1984; *Elston,* 1984*b*]. As *Elston* [1984*a,b*] has emphasized, the passive emplacement of batholithic roots of these cauldrons implies an active extensional stress field.

The structural, stratigraphic, and chemical-isotopic data from the "basaltic andesites" and related rocks discussed above indicate that Oligocene calc-alkaline magmatism continued into early Miocene time but modified to the extent that a regional, stable extensional stress field, firmly established about 30 ± 2 Ma, promoted weakly bimodal but fundamentally basaltic andesite volcanism until about 18 Ma. Another important manifestation of this extension was development of "early rift" basins.

*"Early rift" basins.* Across broad tracts of southern New Mexico, thick basinal sedimentary rocks intertongue with upper Oligocene-lower Miocene "basaltic andesite" flows. Locally as much as 1.9 km thick, these deposits accumulated in broad "early rift" basins as alluvial fan, alluvial flat, and playa facies. Clasts can be traced to late Oligocene to middle Miocene fault blocks, and they also record the erosional "unroofing" of these blocks, some of which are ancestral to modern ranges. Fanglomerate wedges and interformational unconformities document progressive deformation along basin margins during the Miocene. Upper parts of basin fill, however, lack evidence of associated volcanism and, locally at least, represent broad, stable basin floors contemporaneous with the middle to late Miocene magmatic lull in southern New Mexico [*Chapin and Seager,* 1975; *Seager et al.,* 1984]. This period of possibly less vigorous tectonic activity and clear absence of volcanism was followed in latest Miocene time by renewed faulting that modified, or in places completely disrupted, "early rift" basins.

*Low-angle normal faults.* Whereas broad basins seem to have been a widespread product of the early phase of extension, a more local but nevertheless significant additional feature is low-angle faulting. Faults dipping 40° to near horizontal are exposed in several ranges of southern New Mexico (Figure 3). The faults are generally closely spaced in contrast to the widely spaced, high-angle faults which cut them and border modern fault block mountains. It is seldom clear whether the low dips represent flat parts of listric faults or domino-style rotation of originally high-angle faults. In general, we believe that the faults have been substantially rotated during the course of early and/or late phase extension, even if they were originally listric [*Seager,* 1981; *Seager and Mack,* 1986*b*].

Unfortunately, the low-angle faulting cannot be closely dated. About all that can be said is that the faults are latest Oligocene or younger and that they are older than the modern high-angle range boundary faults. They thus fall into the general age of better dated (31–10 Ma) low-angle faults reported by *Chamberlin* [1983] and *Chamberlin and Osburn* [1984] from the Rio Grande rift near Socorro.

There is little evidence to tie low-angle faulting genetically to development of broad early rift basins. Few low-angle faults are known to cut early rift sediments. Most occurrences of low-angle faulting are in the vicinity of major late Oligocene volcano-plutonic complexes (e.g., Organ Mountains, East Potrillo Mountains), whereas regions lacking such igneous centers also lack well-developed low-angle faults (e.g., Caballo Mountains, northern and central San Andres Mountains, Sacramento Mountains). On the other hand, not all volcano-plutonic complexes are associated with low-angle fault systems (e.g., Emory cauldron, Tres Hermanas Mountains). Such inconsistencies make it difficult to correlate directly low-angle faulting with shallow plutonism and volcanism. Nevertheless, the overlap of later phases of middle Tertiary volcanism with the development of closely spaced low-angle faults seems significant. We believe that the faulting was probably a response to extension of an upper crust that had been thermally weakened by 10–20 m.y. of shallow pluton emplacement and volcanism.

*Amount of early phase extension.* Estimates of the amount of early phase extension are difficult to make. Using assumptions and arguments of *McKenzie* [1978] and *Morgan and Golombek* [1984], broad early rift basins up to 1.9 km deep, such as we have described, may indicate as much as 27% extension assuming a 100-km-thick lithosphere prior to extension and that thermal relaxation after rifting did not deepen the basins. Because of uncertain fault geometry, estimates of extension resulting in low-angle faults also is imprecise. Using calculations described by *Thompson* [1960], *Wernicke and Burchfiel* [1982], and *Angelier and Colletta* [1983] based upon palinspastic reconstruction of mapped fault blocks, we believe extension of 50–100% was achieved locally in areas of closely spaced, low-angle faults such as Bishop Cap hills and East Potrillo Mountains. Much less extension is apparent in areas not affected by middle Tertiary magmatism or low-angle faulting. Thus 30% extension may be a reasonable estimate for total early phase extension across the southern rift, and locally extension may have approached 100%.

*Orientation of early phase extension.* We have used the trends of late Oligocene to middle Miocene faults, basins, and dikes to judge the orientation of the regional extensional stress field during the early phase of rifting. Although there are confusing anomalous trends, there also seems to be a preferred orientation of structures between N-S and N60°W, averaging N30°–40°W. This indicates NE-SW extension during the early phase of rifting [*Seager,* 1981; *Newcomer et al.,* 1983; *Seager and Mack,* 1986*b*].

North-northeast directed extension in southern New Mexico during late Oligocene to middle Miocene time is consistent with the direction of extension determined in the northern rift by *Lipman* [1983] and *Golombek et al.* [1983], in west Texas by *Price and Henry* [1984] and *Henry and Price* [this issue], and in Arizona by *Rehrig and Heidrick* [1976]. It is also supportive of Miocene stress fields determined by *Zoback et al.* [1981] for the Basin and Range province in general.

### *Late Phase Extension*

The waning of early phase extension and onset of late phase extension was transitional across the span of the middle Miocene magma gap. Evidence for the late phase or episode of rifting include: new, widespread, coarse-grained fanglomerate wedges at the top of early rift basin fills; the appearance of alkali-olivine basalt flows (Selden Basalt, 9.6 Ma); major offset of upper Miocene rocks; and striking unconformities between

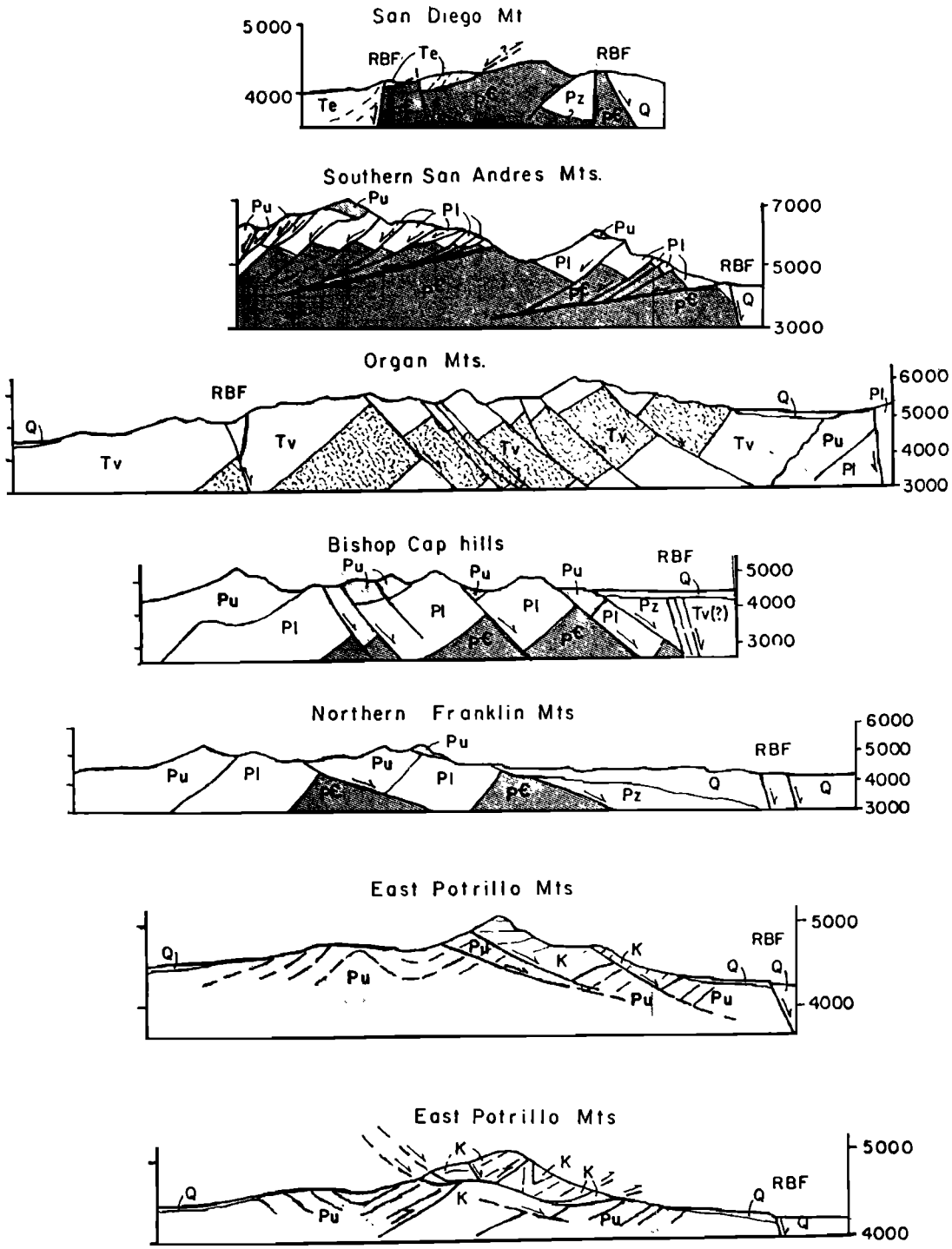


Fig. 3. Low-angle faults in south central New Mexico. No vertical exaggeration; elevations in feet. Note different scales. San Diego Mountain [after *Seager et al.*, 1971]; southern San Andres Mountains, Organ Mountains, and Bishop Cap hills [after *Seager*, 1981]; Franklin Mountains [after *Harbour* 1972]; East Potrillo Mountains [after *Seager and Mack*, 1985*a*]. RBF, high-angle range boundary fault; P€, Precambrian rocks; Pz, Paleozoic rocks undifferentiated; Pl, Lower Paleozoic rocks; Pu, Upper Paleozoic rocks; K, Cretaceous rocks; Tv, Oligocene volcanic rocks of Organ cauldron; Te, "early rift" basin sedimentary rocks; Q, Quaternary rocks.

deformed upper Miocene rocks and undeformed upper Pliocene strata or lavas.

*Timing of late phase.* Several workers along the length of the rift have bracketed the latest phase of rifting, the major episode of faulting that outlined the modern basins, between 10 and 3 Ma [e.g., *Chapin and Seager,* 1975; *Chapin,* 1979; *Baltz,* 1978; *Manley,* 1979; *Manley and Mehnert,* 1981; *Golombek et al.,* 1983; *Seager et al.,* 1984]. In southern New Mexico, high-angle block faulting was vigorously active between 9.6 and 4.5 Ma, continuing to the present but probably with less intensity [*Chapin and Seager,* 1975]. Locally, *Seager et al.* [1984] bracket the main pulse of block faulting between 9.6 and 7.1 Ma in the Las Cruces area.

*Structural style, amount and orientation of extension.* The late phase of block faulting produced relatively widely spaced, high-angle normal faults that border the modern fault block mountains and valleys. The uplifted blocks are simple horsts or tilted blocks, and the downdropped segments are either simple or composite grabens or half-grabens, generally narrow (5–20 km wide) and 1 km or more deep. Although range boundary faults are steep (65°–75°) and seemingly planar at present levels of exposure, some may flatten downward into a midcrustal zone of extension. This is suggested by broadly curved fault traces; by moderate (20°) rotation of fault blocks, by the development of zones of reverse drag and/or intense antithetic faulting in basin fill adjacent to some boundary faults [*Hamblin,* 1965; *Seager,* 1980, 1981], and by interpretation of seismic profiles [*Cape et al.,* 1983]. Five to twenty percent extension appears to be characteristic of late phase faulting based upon reconstructions of the preextension geometry of mapped fault blocks and, considering the small amount of observed tilt of fault blocks, is consistent with either planar or listric fault models [*Wernicke and Burchfiel,* 1982].

Fault blocks and grabens of the late phase trend northerly, although many fault segments trend northwest or even easterly. Overall, the structural grain is northerly, however, resulting in structural truncation of older tectonic trends. It seems clear that the grabens, horsts, and tilted fault blocks are a product of E-W extension in latest Miocene and Pliocene time, in contrast to the NE-SW extension that characterize early phase deformation.

*Late phase magmatism.* Besides differences in stress orientation and structural styles between late and early phase deformation, associated igneous activity is also very different. The onset of late phase extension is marked by the appearance of alkali-olivine to rarer tholeiitic "true" basalts, chemically and isotopically unlike the "basaltic andesites" of the early extensional phase [*Chapin and Seager,* 1975; *Stinnet and Steuber,* 1976]. Although seemingly a product of the late phase of extension, the basalts curiously are more abundant after 5.0 Ma, after the intensity of block faulting waned. Apparently, volcanism was somehow delayed until, or accelerated after, 5.0 Ma. At any rate, basaltic volcanism has continued until as recently as 0.18 Ma in the southern rift. As *Baldridge et al.* [1984] point out, however, the volume of volcanism generated in the late phase of rifting was small in the southern rift.

*Regional uplift.* Evaluation of regional uplift associated with rifting is based largely on a few scattered paleobotanical studies. Palms and other tropical plants, suggestive of near sea level conditions, are present in Eocene and middle Oligocene volcanic rocks in southern New Mexico and have also been reported from middle Miocene strata in northern New Mexico [*Axelrod and Bailey,* 1976]. Subalpine floras of latest Oligocene-early Miocene age from the Black Range (Hillsboro fauna [*Meyer,* 1983]) may be associated with topographically high volcanic structures, probably related in part to the Emory cauldron, and therefore not indicative of latest uplift. Based on the middle Miocene fossil palm, *Axelrod and Bailey* [1976] estimate 1100 m of epeirogenic uplift of the northern Rio Grande rift in the last 13 m.y. or so. In the Socorro area, *Meyer* [1983] indicates about 700 ± 300 m of uplift since middle Miocene, based on his study of fossil Junipers. Much of this regional uplift documented by *Axelrod and Bailey* [1976] and *Meyer* [1984] probably accompanied movements on faults during the late phase of Rio Grande rift extension, although this has not yet been demonstrated conclusively [*Chapin and Seager,* 1975; *Chapin,* 1979].

## CENOZOIC GEOLOGIC HISTORY OF CENTRAL RIO GRANDE RIFT

Laramide compressional deformation in most of north central New Mexico and Colorado produced a province of foreland basement uplifts and basins known as the southern Rocky Mountains [*Baltz,* 1978; *Chapin and Cather,* 1981]. These compressional and strike-slip related basins are not coincident with later extensional basins [*Stearns,* 1953].

### *Early Phase Extension*

The early phase of extension in the central Rio Grande rift occurred during the latter part of a major phase of mostly Oligocene volcanism. Evidence for this volcanic event can be found in the San Luis basin (rhyolitic through andesitic rocks) beneath the Taos Plateau volcanic field [*Lipman and Mehnert,* 1979] and into Colorado [*Tweto,* 1979]; in the Latir volcanic field, adjacent to the San Luis basin (andesite through rhyolite) [*Lipman,* 1981]; in and around the Española basin (andesite, latite, limburgite, and basalt of the Ortiz Mountains, Cerrillos Hills and other localities in the southeastern part of the basin) [*Stearns,* 1953; *Bachman and Mehnert,* 1978; *Baldridge et al.,* 1980; *Kautz et al.,* 1981], and in and adjacent to the southern Albuquerque-Belen basin [*Osburn and Chapin,* 1983].

Faulting in local areas during the early phase of extension was low angle and either planar (domino style) or curved (listric). Where this phase is well documented, in the Questa caldera in the Latir volcanic field [*Lipman,* 1981] and in the Lemitar Mountains [*Chamberlin,* 1983], faulting was both locally pervasive and involved very large strains (greater than 100%). This combination of large strains on closely spaced faults produced highly rotated strata. Constraints on the timing of these faulting events, Questa 23 Ma, and Lemitar 31–28 Ma, implies that they are not synchronous but are temporally and spatially associated with major magmatic events. Other examples attributable to this phase of activity have been suggested flanking the Albuquerque basin [*Baldridge et al.,* 1984]. Between these regions there is little evidence of very large strains, and early phase extension probably involved broad basin development accompanied by small to moderate (~ 30%?) extension.

Broad shallow basins formed along the trend of the central rift by late Oligocene time [e.g., *Chapin,* 1979]. The best documented of these is the Española basin, which began forming about 26 Ma [e.g., *Baltz,* 1978; *Baldridge et al.,* 1980; *Manley and Mehnert,* 1981]. The early Española basin was a broad, shallow, north trending downwarp that was roughly coincident

with the present basin (but wider). Evidence for this geometry comes from early rift sediments extending beyond the present basin margins and thin deposits of rift sediments along the basin edges [e.g., *Manley*, 1979; *Kelley*, 1978].

The least principal stress direction during the early phase of extension is poorly documented. *Lipman* [1981] has shown that prior to 5–10 Ma, NW trending normal faults were predominant in and adjacent to the San Luis basin and thus imply a NE-SW extension direction.

### *Late Phase Extension*

The late phase of extension in the central Rio Grande rift followed a mid-Miocene lull (or at least a significant reduction) in magmatism that lasted from 20 to 13 Ma [e.g., *Chapin*, 1979; *Baldridge et al.*, 1980]. In the central Rio Grande rift this phase of extension produced narrow elongate fault bounded basins along the locus of the earlier shallow broad basins. The present-day basins in the northern rift are north trending basins arranged en echelon to the right resulting in an overall NNE strike to the rift [*Kelley*, 1979, 1982]. In the Española basin, faulting produced the present narrow basin about 10 Ma [*Manley*, 1979], as indicated by a 9.8 Ma dike [*Bachman and Mehnert*, 1978] intruded along a NE trending western border fault. Local and regional approaches indicate that this phase of extension was quite accelerated at about 10 Ma [e.g., *Golombek et al.*, 1983; *Golombek*, 1984; *Dethier and Martin*, 1984; *Gardner and Goff*, 1984]. This period of accelerated activity coincides with a regional change in the least principal stress direction from WSW-ENE to WNW-ESE [e.g., *Zoback et al.*, 1981; *Lipman*, 1981; *Golombek et al.*, 1983].

In contrast to the earlier phase of extension, faulting during the late phase resulted in much less tilted and rotated strata (implying high angle faults). Although there is some debate concerning the exact form of the faults at depth (i.e., listric or planar [e.g., *Woodward*, 1977; *Brown et al.*, 1980; *Brocher*, 1981; *Cape et al.*, 1983; *De Voogd et al.*, this issue], there is little question that vertical movements were more important than block rotation during the late phase. The narrow basins produced are typically asymmetric with major border faults on one side only (half- graben) [e.g., *Kelley*, 1979, 1982]. Attempts to estimate the extension across the central rift suggest roughly 10% of extension [*Woodward*, 1977; *Brocher*, 1981; *Cordell*, 1982; *Golombek*, 1981; *Golombek et al.*, 1983]. Given the structural relief across the modern basins (several to 10 km) [e.g., *Birch*, 1982; *Cordell*, 1979; *Kelley*, 1977; *Woodward*, 1977], the late phase of extension can be characterized as having small amounts of extension and large vertical displacements. Unlike the early extension event, volcanism is, in general, poorly associated with the late phase. If anything, most volcanics in the central rift are low-lying basalt fields that formed after the accelerated phase of activity ended (in the last 5 m.y.) [e.g., *Kudo*, 1982; *Lipman and Mehnert*, 1979; *Aubele*, 1979].

Regional uplift has accompanied the late phase of extension in the central Rio Grande rift. Fossil flora [*Axelrod and Bailey*, 1976], uplifted erosion surfaces [*Scott*, 1975; *Taylor*, 1975], and fission track dates [*Kelley and Duncan*, 1984; this issue] all indicate relatively rapid uplift of more than 1 km in the past 7–10 m.y.

## SUMMARY OF GEOLOGIC HISTORIES

In summary, two extensional regimes of different origin (but transitional with each other through the Miocene) can be interpreted from rocks and structures formed within the last 30 ± 2 m.y.. The older regime is characterized by emplacement of basaltic andesite and related flows with relatively high strontium isotope ratios, and lesser, more siliceous rocks; by formation of broad, shallow northwest trending basins; and by the local development of closely spaced, low-angle faults, generally in the areas of volcano-plutonic complexes. This regime evolved under an extensional stress field oriented NE-SW and extension of the order of 30% was characteristic and may locally have approached 100% or more in regions of high heat flow and closely spaced faulting. This early extension phase closely matches in both style, timing, and associated volcanism a similar phase of extension in the Basin and Range (e.g., pre-Basin and Range event of *Zoback et al.* [1981] that lasted from 30 to 13 Ma). Although it cannot be demonstrated that low-angle faults are related to the formation of the early broad basins, the similarity in timing between early basin formation and low-angle faulting, coupled with the same regional association in parts of the Basin and Range [e.g., *Zoback et al*, 1981; *Eaton*, 1982] implies a genetic relationship.

The younger episode or phase of extension seemingly represents a renewal or acceleration of block faulting and later volcanism beginning about 9 to 10 Ma, after a long transition period during the middle Miocene when volcanism was absent (or minor) and tectonism perhaps less vigorous. This latest phase resulted in segmentation of earlier rift basins into narrower horsts and grabens, formation of northerly trending modern rift uplifts and basins by movement on widely spaced, high-angle faults, and the later renewal of volcanism, this time dominated by relatively primative alkali-olivine basalt. Late phase deformation followed a change to E-W extension and was accompanied by epeirogenic uplift of 700–1100 m. Also at about 10 Ma, this change in stress direction resulted in opening of the northern Basin and Range province [*Zoback et al.*, 1981] that can be related to a change in the predominantly transform motion between the Pacific and North American plates. Reduced geothermal gradients and thicker, brittle upper crust are indicated by the surficial style of faulting and by lesser volcanic activity compared to the early phase of deformation. Only 5–20% extension resulted from late phase extension, but vertical displacements were in general at least as great or greater than in early phase extension.

## GEOPHYSICAL AND OTHER EVIDENCE FOR MODERN LITHOSPHERIC STRUCTURE

Geophysical data relating to crustal structure in the Rio Grande rift were summarized by *Cordell* [1978] and in more detail for the southern Rio Grande rift by *Seager and Morgan* [1979]. These data show anomalous crustal and upper mantle structure associated with the southern Rio Grande rift. The seismic data indicate that the crust of the southern Basin and Range and southern Rio Grande rift (23–28 km) is significantly thinner than the adjacent Colorado Plateau (40 km) and Great Plains provinces (50 km) [*Toppozada and Sanford*, 1976]. New seismic refraction data indicate that the crust thickens from approximately 23 km in the Basin and Range of southwestern Arizona [*Sinno et al.*, 1981] to approximately 31 km at the Arizona–New Mexico border [*Gish et al.*, 1981] but thins again to between 27 and 28 km beneath the southern Rio Grande rift [*Cook et al.*, 1979; *Daggett*, 1982; *Sinno et al.*, this issue]. Long-wavelength Bouguer gravity data are also consistent with crustal thinning beneath the southern Rio Grande rift [*Ramberg et al.*, 1978; *Daggett*, 1982; *Daggett et al.*, this issue]. Thus there

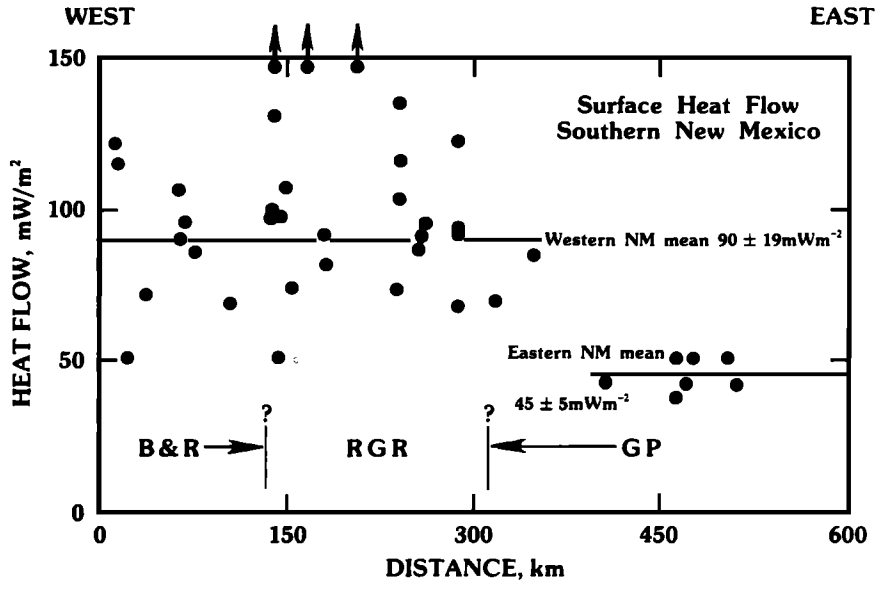


Fig. 4. Surface heat flow data from southern New Mexico projected onto a west to east profile showing little apparent difference between the southern Rio Grande rift and adjacent Basin and Range (redrawn from *Daggett*, [1982]).

is good evidence for anomalous crustal and upper mantle structure associated with the late phase extension event which formed the modern southern Rio Grande rift, and similar, but less definitive data suggest similar anomalous lithospheric structure to the north [e.g., *Baldridge et al.*, 1984].

Heat flow data and deep electrical sounding data indicate high lower crustal and upper mantle temperatures associated with the Rio Grande rift and adjacent Basin and Range province. In the south, deep electrical sounding data indicate that the thermal anomaly is more intense beneath the rift than the adjacent Basin and Range [*Seager and Morgan*, 1979], but the surface heat flow data do not make a clear distinction between the Basin and Range and Rio Grande rift provinces: contour maps of these data are somewhat subjective (compare *Reiter et al.* [1975], *Seager and Morgan* [1979], *Swanberg* [1979], and *Sass et al.* [1981]). A large scatter in the surface heat flow data from southwestern New Mexico makes an objective analysis of these data very difficult, as shown in Figure 4. The Great Plains and southern Rio Grande rift provinces are clearly thermally distinct, but there are no major thermal boundary between the southern Rio Grande rift and adjacent Basin and Range province. *Decker and Smithson* [1975] suggested that the southern Rio Grande rift is characterized by a heat flow of approximately 100 mW $m^{-2}$, and a crustal geotherm derived from this value is consistent with lower crustal xenolith pressure and temperature data [*Padovani and Carter*, 1977; *Seager and Morgan*, 1979]. This geotherm exceeds temperatures of 1000°C just below the Moho, however, which together with upper mantle xenolith pressure and temperature data indicates a convective heat transfer regime in the upper mantle beneath the southern Rio Grande rift [*Seager and Morgan*, 1979]. Scatter in the heat flow in the southern Rio Grande rift around the mean value possibly results in part from magmatic activity associated with this convective regime. To the north a clearly identified surface heat flow high is associated with the rift, and calculations based upon these data suggest geotherms that exceed the crustal solidus above the Moho [e.g., *Clarkson and Reiter*, 1984; *Decker et al.*, 1984; *Morgan and Golombek*, 1984].

Reduced heat flow data suggest that the southern Rio Grande rift and adjacent Basin and Range province are thermally distinct. As shown in Figure 5, there is no systematic correlation for heat flow–heat generation data from the southern rift, but data from the adjacent Basin and Range show the same correlation as data from other areas of the Basin and Range province where young volcanism has not occurred [e.g., *Roy et al.*, 1972; *Blackwell*, 1978]. These data indicate generally high, but laterally variable, reduced heat flow in the southern Rio Grande rift, with elevated, but fairly uniform, heat flow in the adjacent Basin and Range province.

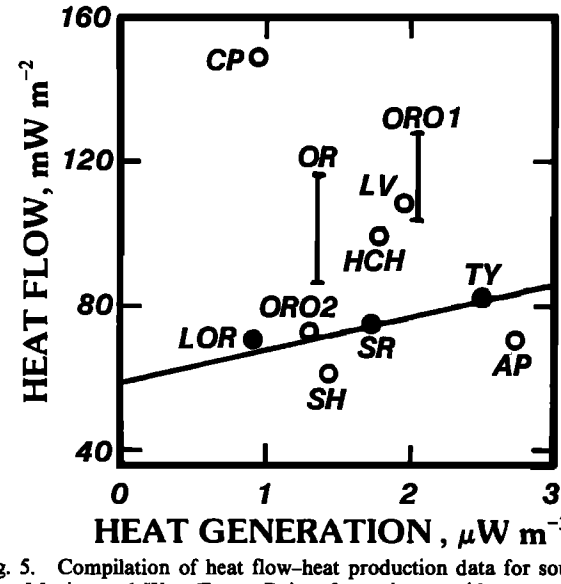


Fig. 5. Compilation of heat flow–heat production data for southern New Mexico and West Texas. Points from sites outside areas of Rio Grande rift volcanism (less than 15 Ma; solid circles), i.e., LOR, SR and TY, fall close to the "Basin and Range" heat flow–heat production line, indicating a predictable lithospheric thermal structure at these sites. Points from Rio Grande rift sites (open circles and bars) show large scatter indicating disturbed geotherms. Key to data: CP, Cooks Peak; LOR, Lordsburg; OR, Organ Mountains; ORO1 and ORO2, Orogrande; SH, Shaefer, Texas; SR, Santa Rita; HCH, Hachita; LV, Lake Valley; TY, Tyrone; AP, Animas Peak (data from *Cook et al.* [1978]).

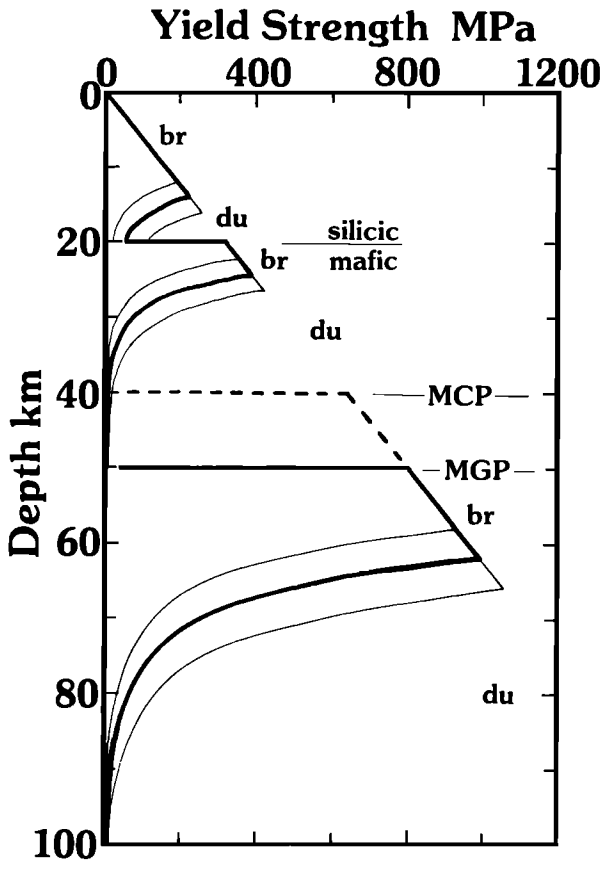


Fig. 6. Calculated lithospheric strength curves for the Great Plains (solid curve) and early Tertiary Colorado Plateau (dashed curve) assuming a shield geotherm [from *Lachenbruch and Sass,* 1978]. Heavy curves calculated for strain rate of $10^{-15}$ $s^{-1}$. Light curves above and below heavy curve calculated for strain rates of $10^{-16}$ and $10^{-14}$ $s^{-1}$, respectively. Depths to Great Plains and Colorado Plateau Mohos are indicated by MGP and MCP, respectively (depths from *Cordell* [1978]).

In summary, geophysical and other subsurface data indicate that the present crustal and lithospheric structure is most strongly controlled by the late phase, Rio Grande rift extension event. Perhaps rather surprisingly, the much larger magnitude early phase extension event has left little permanent imprint on gross crustal structure. We now discuss the mechanical implications of the two extension events and use the geologic history of the region and constraints from modern crustal and lithospheric structure to attempt to reconstruct the thermal and mechanical history of the lithosphere in this region.

## BASIN GEOMETRY AND EXTENSIONAL STYLE

Two distinct phases and styles of Cenozoic extension have been identified in the Rio Grande rift. During the first phase of extension, broad relatively shallow basins were formed associated with local high extensional strains. In contrast, relatively small extensional strains formed deep narrow basins during late phase extension. As discussed by *Morgan and Golombek* [1984], these contrasting basin styles, both formed by extension, can be described in terms of end-member mechanisms for extensional basin formation. Broad, relatively shallow basins can most easily be explained by a mechanism in which extensional strain is distributed over a wide area and in which the lithosphere is thin and weak, so that it is always locally in isostatic equilibrium. Downwarp is caused by thinning of the buoyant crust during extension [*McKenzie,* 1978]. Narrow, deep basins can most easily be explained by fault-controlled subsidence in local isostatic disequilibrium. By this mechanism, local downwarp is purely a function of fault geometry, with high-angle normal faults resulting in large vertical displacements for relatively small lateral movements [*Vening Meinesz,* 1950; *Bott and Mithen,* 1983]. This mechanism may be expected to prevail in stronger, thicker lithosphere as flexural strength in the lithosphere supports local isostatic disequilibrium, and isostatic equilibrium will only be obtained on a regional scale, with some of the basin downwarp compensated by upwarp of the flanks of the basin. On a regional scale the net subsidence for the fault downwarp model will be the same as for the stretched crust model, but the fault subsidence model produces greater local subsidence than the stretching model for the same extensional strain [*Morgan and Golombek,* 1984]. More complete analyses of the stretching-subsidence mechanism is given by *McKenzie* [1978] and of the fault subsidence mechanism by *Bott* [1976] and *Bott and Mithen* [1983].

Quantitative modeling of subsidence associated with extension is limited by a number of unknown or poorly constrained parameters. For example, uncertainties in fault geometries and displacements may make an accurate estimate of the amount of extension difficult. Initial model parameters may also be difficult to define, as rifting perhaps rarely takes place in "normal" lithosphere, as we emphasize in our examples from

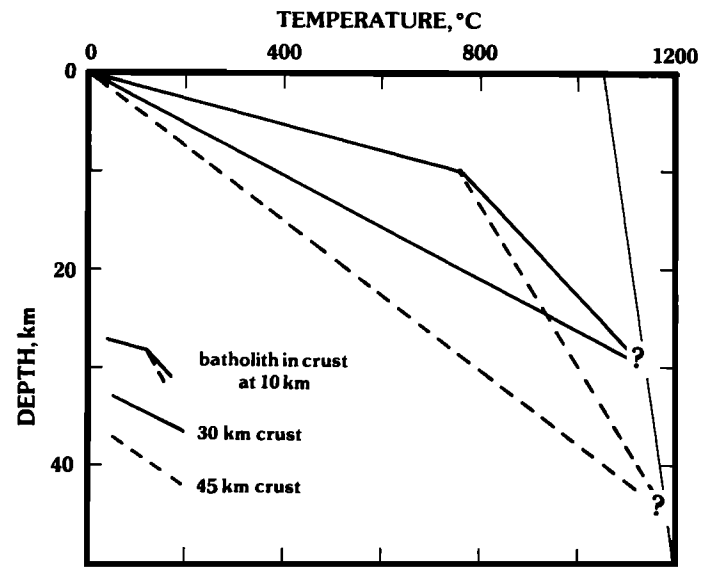


Fig. 7. Simplified estimated geotherms for southwestern New Mexico at 30–25 Ma for different crustal thicknesses and with a batholith in the crust at 10 km. The 30-km crust represents the crust at the time of early phase extension, assuming no crustal thinning during extension (thickness maintained by addition of magmas). The 45-km crust represents probable maximum crustal thickness assuming maximum crustal thinning during extension for 50% extension. Fine line on right represents the basalt dry solidus [from *Lachenbruch and Sass,* 1978], and geotherms are constrained to intersect this solidus at the Moho, based on the geochemical evidence for crustal melting and magma contamination. The batholith was assumed to be intruded at the solidus temperature for tonalite of 760°C at 10 km [from *Wyllie,* 1977]. All curves are schematic, especially at higher temperatures, and do not accurately represent the transient effects of intrusions. However, the strength curves generated from these geotherms are not sensitive to the geotherm at high temperatures and are considered to be valid for purposes of qualitative comparisons.

New Mexico (compare *McKenzie* [1978]). In addition, the lithosphere may not be a closed system during extension, and material may be added by magmatism prior, during or postextension [e.g., *Lachenbruch and Sass,* 1978; *Royden and Keen,* 1980]. Thus, at this point we use only the observation that two distinct extensional styles are overprinted in the southern Rio Grande rift, an early ductile extension event indicating a thin weak lithosphere and a later high-angle fault-controlled event indicating a stronger thicker lithosphere. We attempt to reconcile the geologic history of the region with this observation.

## GEOTHERM EVOLUTION AND IMPLICATIONS

The overprinting of the two extension events suggest that the control on extensional style is not fundamental to the crust but is a temporally varying parameter. The two main temporally variable parameters which control the mechanical strength of the lithosphere are strain rate and temperature. Strength increases as strain rate increases, but decreases as temperature increases (see the appendix). In New Mexico, the high strain (early extension) event was characteristic of a thin weak lithosphere. The low strain (late phase extension) event was characteristic of a thicker, stronger lithosphere. Thus, if total strain can be taken as a measure of strain rate, the behavior of the lithosphere during these two extension events is opposite to that expected from the strain. As a first approximation, therefore, we ignore strain rate in our analysis and concentrate on temperature changes, or the evolution of the geotherm. The techniques used in this study for calculation of lithospheric strength curves from lithospheric structure and the geotherm are given in the appendix.

Our results are useful in a qualitative rather than a quantitative sense, and we use a reference lithosphere with which to compare changing geotherms and changing lithospheric strength curves.

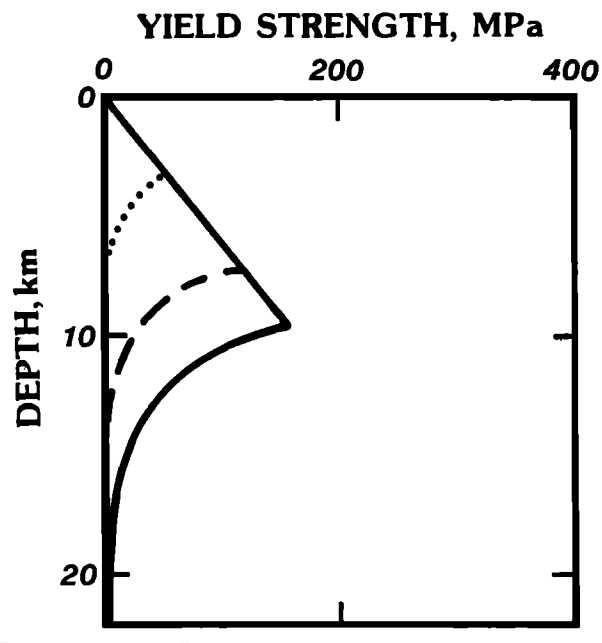


Fig. 8. Calculated lithospheric strength curves for southwestern New Mexico at 30–25 Ma based upon the geotherms shown in Figure 7 and assuming an extensional strain rate of $10^{-15}$ $s^{-1}$. Solid curve represents 45-km crust; dashed curve represents 30-km crust; dotted curve represents crust with batholith at 10 km for both crustal thicknesses.

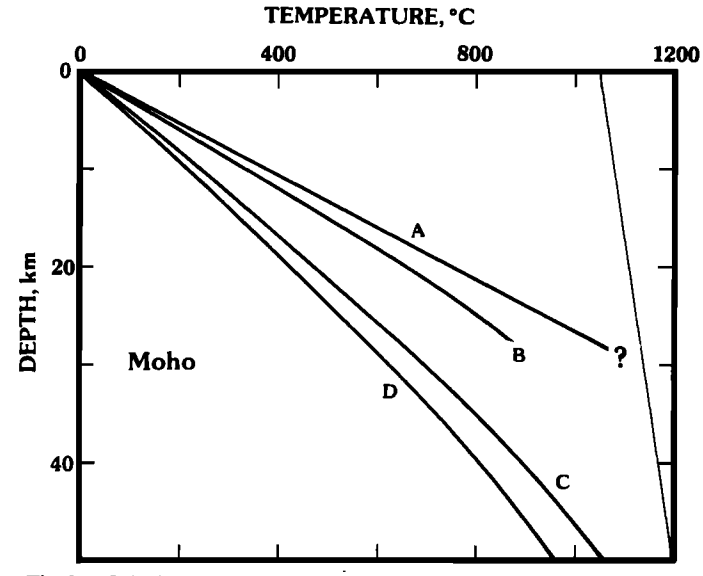


Fig. 9. Calculated geotherms for southwestern New Mexico, assuming conductive cooling after early phase extension and the modern southern Rio Grande rift geotherm. Curve A, assumed geotherm at 25 Ma (see Figure 7); curve B, modern Rio Grande rift geotherm [from *Seager and Morgan,* 1979]; curve C, calculated geotherm for 10 Ma; curve D, calculated geotherm for 0 Ma. Curves C and D were calculated from curve A assuming conductive cooling of the lithosphere starting at 25 Ma, using the cooling solution given by *McKenzie* [1978] and assuming a crustal thermal conductivity of 2.5 W $m^{-1}$ $K^{-1}$, a diffusivity of 25 $km^2$ $my^{-1}$, and an equilibrium lithospheric thickness of 125 km. Curve D closely matches the predicted Basin and Range geotherm for a reduced heat flow of 59 mW $m^{-2}$ [*Roy et al.,* 1972]. See note in caption to Figure 7 regarding validity of geotherms.

For this lithosphere, we use a shield geotherm appropriate to the Great Plains province [*Lachenbruch and Sass,* 1978], from which the lithospheric strength profile shown in Figure 6 was calculated. For this strength curve and all other curves in our analysis, we have assumed a strain rate of $10^{-5}$ $s^{-1}$ (approximately 3%/m.y.). This strain rate is probably reasonable for late phase extension and is assumed constant to allow examination of the effect of the changing geotherm.

To determine the geotherm at the start of the early phase extension (approximately 30 Ma), we must rely upon indirect evidence for lithospheric temperatures. Approximately 15 m.y. of intense magmatic activity preceded this extension event, and we interpret strontium isotope ratios of the magmas just prior to extension to indicate that crustal melting was caused by this magmatic activity just prior to extension. As a lower bound to temperature, therefore, we use the solidest temperature at the Moho and assume that thermal equilibrium was developed in the lithosphere during this long period of magmatic activity. Several major upper crustal batholiths were implaced at this time, indicating higher temperatures, and a higher geotherm associated with this magmatic activity is also considered. One final unknown at this time is crustal thickness. If no magma was added to the crust during extension, from the regional extensional strain of approximately 30–50% and the modern crustal thickness of 30 km (for the southern rift), we estimate a preextension crustal thickness of 40–45 km. This may be regarded as an upper bound to crustal thickness, and a lower bound is the modern crustal thickness, or 30 km, assuming all extension to be accommodated by the addition of magma

to the crust. Schematic geotherms based upon these geological inferences are shown in Figure 7, and strength curves based on these geotherms are shown in Figure 8. Similar geotherms and strength curves with a thicker crust applicable to the rift in northern New Mexico were given by *Morgan and Golombek* [1984]. These strength curves indicate that prior to early phase extension, the lithosphere in the area of the rift zone of New Mexico was very weak relative to the Great Plains, with no significant mantle strength and a very shallow crustal brittle-ductile transition (10–12 km). This brittle-ductile transition may have been locally much shallower, associated with higher upper crustal temperatures around recently intruded batholiths.

To calculate the geotherm and strength curves at the start of the late phase extension, we have assumed that the lithosphere conductively cooled following early phase extension. Starting with the 30-km-thick crust geotherm from Figure 7, we have used the cooling equations of *McKenzie* [1978] to calculate geotherm at 10 Ma and the present, assuming that cooling started at 25 Ma, the approximate age of the start of waning of magmatic activity. These geotherms and the modern southern Rio Grande rift geotherm are shown in Figure 9, and the strength curves calculated from these geotherms are shown in Figure 10. A comparison of the pre-early phase and pre-late phase extension strength curves in Figures 8 and 10 shows that the crustal brittle-ductile transition depth increased from around 10–12 km to about 14 km between the extension events. Perhaps the most important difference between the strength curves, however, is the appearance of significant uppermost mantle strength in the pre-late extension curves and a zone of lower crust strength. The cooling model predicts that the geotherm should continue to cool and that the lithosphere should continue to strengthen to the present. The modern predicted geotherm shown in Figure 9 is consistent with the reduced heat flow data from regions of the Basin and Range province of southwestern New Mexico where young (<12 Ma) volcanism has not occurred. Petrologic and surface heat flow data from the southern Rio Grande rift which has recent volcanism indicate a higher modern geotherm for the rift as shown. Similar geotherms and strength curves are applicable to Rio Grande rift evolution in northern New Mexico, as presented by *Morgan and Golombek* [1984]. However, thicker crust in the north results in less strength in the uppermost mantle for all stages of geotherm evolution. As discussed below, we consider the late Cenozoic reheating of the lithosphere in the Rio Grande rift to be primarily a postrifting phenomenon based upon recent volcanism following the middle Miocene volcanic lull in this zone and the different reduced heat flow characteristics of the Rio Grande rift and adjacent Basin and Range (Figure 5).

The main results of our geotherm and strength curve evolution deductions can be summarized as follows: (1) Prior to early phase extension, the geotherm was very high and the lithosphere extremely thin and weak, its weakest points associated with the intrusion of upper crustal batholiths, (2) if the lithosphere is assumed to have conductively cooled after early phase extension, lithospheric strength is predicted to have increased prior to late phase extension, with significant strength in the uppermost mantle as well as the upper crust and some strength in the lower crust, (3) the model of conductive cooling from the early extension event to the present is consistent with reduced heat flow data from the Basin and Range province of southwestern New Mexico where there are no recent volcanics, and (4) there appears to have been a reheating event of the lithosphere associated with the southern Rio Grande rift, possibly postrifting, related to recent magmatism.

## CONTROLS ON STRUCTURAL STYLE OF EXTENSIONAL TECTONICS

### *Early Phase Extension*

The lithospheric strength curves shown in Figure 8 for the time of early phase extension in New Mexico suggest a simple mechanism for the generation of broad relatively shallow basins during this extension event. As illustrated in Figure 11*a*, only the uppermost crust will be expected to behave brittlely with shallow detachment and ductile extension in the remainder of the crust and upper mantle. To accommodate the transformation from brittle to ductile extension, a decollement would be expected to form at the brittle-ductile transition, with normal faults in the uppermost crust becoming listric and flattening at this depth. Faulting in the uppermost crust causes primarily block rotation with little net vertical relief. Subsidence is caused primarily by the isostatic effect of crustal necking. The close association of high strain with intrusions during this phase of extension suggests that some subsidence may even have been reduced by the addition of magma to the crust at these locations.

### *Late Phase Extension*

The tectonic style of crustal extension at the time of late phase extension is less easy to deduce from the strength curves predicted for this period. The growth of lower crustal and significant uppermost mantle strength during cooling of the lithosphere from early to late phase extension and the contrasting styles of the two extension events suggest that uppermost mantle strength played an important role in controlling tectonic style during late phase extension. We suggest that the growth of this lower crustal and uppermost mantle strength prevented crustal decollement, as predicted for the early phase extension, and this in turn prevented flattening of the normal faults associated with extension. As actual fault failure is expected to be a very high strain rate event, brittle failure may have penetrated the crust into the upper mantle (Figure 11*b*) [*Vening Meinesz,* 1950]. Alternatively, high-angle normal faults in the upper crust may be terminated by ductile extension and perhaps flattening in the lower crust (Figure 11*c*) [*Bott and Mithen,* 1983] as suggested by minor rotation of fault blocks (see above). The exact mechanism of crustal failure is unimportant for the present arguments, as ductile relaxation of the lower crust and uppermost mantle will be expected to produce similar final crustal geometries from either throughgoing crustal faulting or uppermost crustal faulting only. The important concept suggested by the predicted strength curves and the observed structural styles is the control of normal fault geometry by the growth of strength in the lower crust and uppermost mantle as the lithosphere cools.

## DISCUSSION

From our analysis of extensional tectonics in the framework of the geological evolution of New Mexico, we suggest that extensional strain may be localized by earlier geological events. Extension in the Basin and Range and Rio Grande rift provinces was preceded by major magmatic activity which raised the geotherm and reduced lithospheric strength [*Morgan and Golombek,* 1984; *Baldridge et al.,* 1984]. This magmatic activity

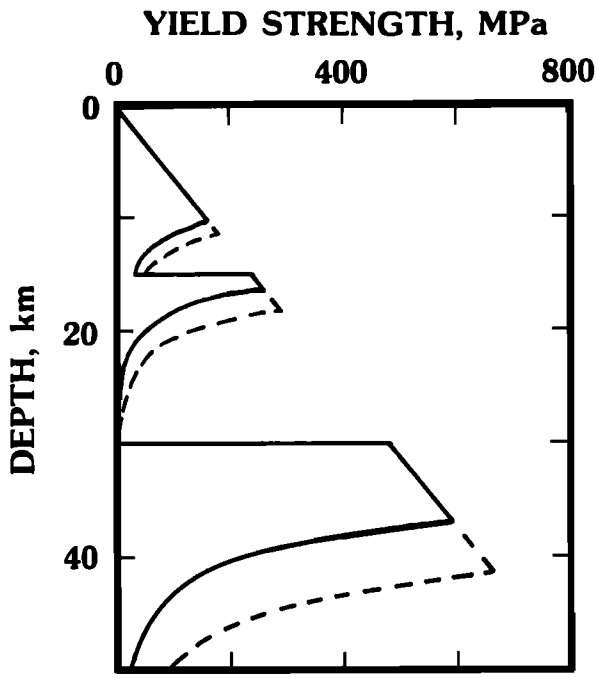


Fig. 10. Calculated lithospheric strength curves for geotherms calculated for 10 Ma (solid curve) and 0 Ma (dashed curve) shown in Figure 9, assuming an extensional strain rate of $10^{-15}$ $s^{-1}$.

was not widespread in the adjacent Great Plains province, and we suggest that the Great Plains lithosphere was resistant to extensional stresses by virtue of high strength resulting from its cool geotherm. The close spatial association of high-strain early phase extension and magmatic intrusions suggests a strong localization of high strain by elevated upper crustal temperatures during this event. The direction of extensional strain in this early phase event is consistent with the regional stress field at that time [*Zoback and Zoback,* 1980], but rapid lateral variations in extensional strain suggest isolation of the areas of strain from the relatively rigid blocks of the Great Plains and perhaps the Colorado Plateau during this event.

Wherever a boundary between a rigid and deformed block has a component of strike parallel to the axis of extension (or compression), a strike-slip decoupling between the two blocks must exist to accommodate the difference in strain across the boundary. Thus, to accommodate early phase extension in southwestern New Mexico, there must have been strike-slip decoupling between the Colorado Plateau and the area to the southeast, in the region now covered by the Datil-Mogollon volcanics, although strike-slip faulting of the same age as early phase extension has not yet been recognized in the Datil-Mogollon field.

During late phase extension, extensional strain was still confined to areas of early and mid-Cenozoic magmatic activity, but extension was not strongly controlled locally by magmatic activity. We tentatively suggest that extensional strain has continued later in the Rio Grande rift (and the Basin and Range of the Great Basin) than in the adjacent southern Basin and Range during late phase extension due to the stabilizing effect of the Colorado Plateau: East-west extension requires strike-slip decoupling of the Colorado Plateau from the southern Basin and Range (the region south of the plateau), whereas areas east and west of the plateau are not influenced by the plateau during east-west extension. The direction of extension during the late phase is consistent with the contemporary regional stress field [*Zoback and Zoback,* 1980]. Thus strain was localized during early phase extension by earlier and concurrent heating of the lithosphere. Some localization of strain in late phase extension may have resulted from resistance of the Colorado Plateau to east-west extension.

The Rio Grande rift shares its two-phase extension history with much of the Basin and Range province [e.g., *Eaton,* 1982; *Zoback et al.,* 1981; *Golombek et al.,* 1983]. Possible association of a changing geotherm with changing extensional style has also been suggested for the Basin and Range [e.g., *Lucchitta and Suneson,* 1982], and we suggest that the concepts developed for the Rio Grande rift in the present study may be generally applicable to the Basin and Range province.

The lithospheric geotherm may be a strong factor in controlling the style of extensional tectonics. Ductile extension producing broad relatively shallow basin is favored by a high geotherm, and fault block tectonics producing narrow deep basins is favored by a cooler geotherm. This observation suggests that it may be difficult to predict the initial thermal condition for ductile extension and basin formation, and it is interesting to note that in many applications of *McKenzie's* [1978] stretching model, it has been necessary to invoke mechanisms that in effect introduce more heat into the system, such as dike intrusion or two-layer stretching [e.g., *Royden and Keen,* 1980; *Hellinger*

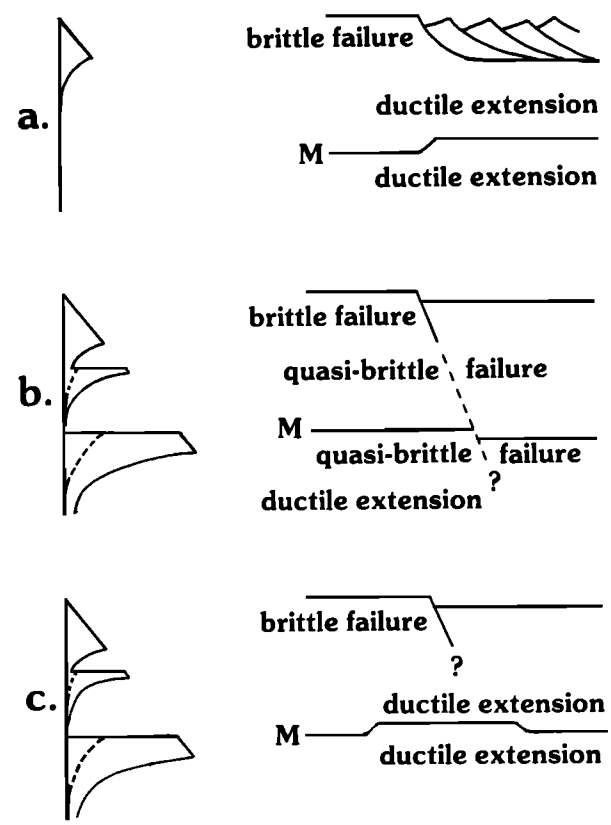


Fig. 11. Relationships between strength curves (left) and styles of extensional tectonics (right). (*a*) Early phase extension. (*b*) and (*c*) Late phase extension. Not to scale. See text for explanation. Upwarp of the Moho (M) is shown schematically in Figures 11(*a*) and 11(*c*) Only the direct effects of faulting on the Moho are shown in Figure 11(*b*) Solid strength curves represent southern New Mexico (Figures 8 and 10); dashed curves represent northern New Mexico [from *Morgan and Golombek,* 1984].

*and Sclater,* 1983]. Difficulties inherent in predicting the preextension geotherm may thus limit the usefulness of these extension-subsidence models.

Further problems with simple extensional-subsidence models are evident when the regional elevation history of southern New Mexico is examined. Up to 1.25 km of regional subsidence is predicted by 30% regional crustal extension during early phase extension and more if greater extension is assumed. No such regional subsidence has yet been documented, suggesting that crustal extension may have been accommodated by the addition of new material to the crust. In fact, *Elston* [1984*a*] has suggested uplift of at least 1–2 km by magmatic crustal thickening during this early phase of extension in southwestern New Mexico. Additional subsidence is predicted during late phase extension, and the 2–3 km of crustal thinning indicated by seismic and gravity data associated with the southern Rio Grande rift are consistent with approximately 10% extension during this event. However, the elevation changes predicted by the late phase extension event cannot explain the approximately 1 km of uplift determined for the last 10–13 m.y. in the Rio Grande rift. Post-middle Miocene uplift may be associated with postrifting, magmatic thickening of the crust, and the modern high heat flow and anomalous uppermost mantle beneath the southern Rio Grande rift may be associated with this magmatic activity rather than directly with late phase extension (P. Morgan et al., manuscript in preparation, 1986) [see also *Cook et al.,* 1978; *Morgan and Golombek,* 1984; *Morgan and Swanberg,* 1985]. The recent rift uplift is perhaps not restricted to the rift zone but includes the Colorado Plateau [e.g., *Morgan and Swanberg,* 1985], the Southern Rocky Mountains, and much of the western United States: If the uplift in all this region is related to magmatism, the young magmatism in the rift may not be directly related to rifting. The complex interrelationship between the regional stress field and preextension and postextension magmatic events suggests that the Rio Grande rift can only be explained in terms of the interrelationship between local asthenosphere-lithosphere conditions and the regional stress field.

## CONCLUDING REMARKS

Detailed geological information and reasonable geophysical control of lithospheric structure have allowed us to examine the sequence and style of extensional tectonics in New Mexico during the Cenozoic. We are able to produce a consistent model of geotherm evolution and tectonic style, but this model suggests a complex interrelationship between regional and local geologic events. Some of the paradoxes apparent from analyses of different rifts may be resolved by these models, but in turn they suggest that models of extensional tectonics may only be applicable on a local rather than a global basis. Rifting events cannot be analyzed in the isolation of the geologic events that precede and follow crustal extension.

## APPENDIX: LITHOSPHERIC STRENGTH CURVES

Lithospheric yield strength curves used in this study are based upon a compilation of laboratory and field experimental data by *Lynch* [1983] and *Lynch and Morgan* [1986] for brittle failure and ductile creep. For essentially qualitative comparison uses of lithospheric strength curves, we have assumed a constant strain rate and that no stress relaxation occurs in the lithosphere before faulting, which allows the following strength relationships to be used [from *Lynch,* 1983 and *Lynch and Morgan,* 1986].

Brittle strength under extension $S_b$ is given by

$$S_b = 16\ z\ \text{MPa}$$

where z is depth in kilometers.

Ductile strength $S_d$ is given by

$$S_d = \left(\frac{\dot{\epsilon}}{\dot{\epsilon}_0}\right)^{1/3} \exp\left(\frac{Q^* + 293\ z}{3RT}\right) \text{MPa}$$

where $\dot{\epsilon}$ is strain rate ($s^{-1}$), R is the gas constant ($J\ mol^{-1}\ K^{-1}$), T is absolute temperature (K), and $\dot{\epsilon}_0$ and $Q^*$ are constants dependent on rock type given by:

| Rock Type | $\dot{\epsilon}_0$, $MPa^{-3}\ s^{-1}$ | $Q^*$, $J\ mol^{-1}$ |
|---|---|---|
| Silicic | $2.5 \times 10^{-8}$ | $1.4 \times 10^5$ |
| Mafic | $3.2 \times 10^{-3}$ | $2.5 \times 10^5$ |
| Ultramafic | $1.0 \times 10^3$ | $5.5 \times 10^5$ |

Lithospheric yield strength S is then given by:

$$S = S_b \qquad S_b > S_d$$
$$S = S_d \qquad S_b < S_d$$

Strength curves used in this paper have been based upon the arbitrary assumption that the upper half of the crust is silicic in composition and the lower half of the crust is mafic in composition. Different (reasonable) assumptions about crustal composition change the details of the curves but do not change the qualitative results that we extract from the curves. The effect of allowing stress relaxation in the lithosphere is to make the brittle-ductile transition(s) more shallow and change the shape of the section of the strength curve relating to ductile strength [e.g., *Lynch,* 1983; *Kusnir and Park,* 1984], but again, the qualitative conclusions that we derive from the strength curves are not changed by these modifications to the models.

*Acknowledgements.* Many of the concepts developed in this study have been suggested by previous students of Basin and Range and Rio Grande rift geology, and in particular, we have benefited from discussions with Ivo Lucchitta. Our analyses of lithospheric strength have relied heavily upon studies of this subject by H. D. Lynch. We are grateful to F. Cook and an anonymous reviewer for their careful reviews of this manuscript. P.M. started this study while at the Lunar and Planetary Institute, which is operated by the Universities Space Research Association under contract NASW-3389 from the National Aeronautics and Space Administration. Work by M.P.G. on this study was carried out at the Jet Propulsion Laboratory, California Institute of Technology, under contract from the National Aeronautics and Space Administration. This is LPI contribution 582.

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