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Summary of DATA

RIO GRANDE RIFT: PROBLEMS AND PERSPECTIVES

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INTRODUCTION

Our intent in this paper is to present new ideas concerning the Rio Grande rift which we hope are at least provocative, if not correct, and to emphasize important unanswered questions which will be fruitful areas for continued research. Our approach is to evaluate, relate, and interpret existing data derived from a number of disciplines, rather than primarily to present new data. We hope thereby to achieve a synthesis of sorts—a new and broad perspective on the Rio Grande rift. This perspective should help guide future observations by directing attention toward the more fundamental questions regarding rifting processes. In this paper we use the Basaltic Volcanism Study Project (1981, p. 838) definition of rifts as elongate depressions overlying places where the entire thickness of the lithosphere has ruptured in extension.

Among the topics and ideas that we address are: (1) the regional extent of the Rio Grande rift, (2) the structure of the crust and upper mantle, (3) whether the evidence for an "axial dike" (i.e., a composite mafic intrusion) in the lower crust is compelling, (4) the nature of faulting and extension in the crust, and (5) the structural and magmatic development of the rift. These issues provide important constraints on the nature of thermal and tectonic processes involved in formation of the Rio Grande rift.

The Rio Grande rift is a region where the lithosphere is being permanently altered through thinning and (probably less importantly) intrusion of mafic magmas. The rift is the culmination of a long and complex geologic history. Initiation of rifting probably resulted from plate-boundary interactions along the west coast of North America (e.g., Lucas and Ingersoll, 1981). However, the presence of a major thermal event associated with the immediately preceding subduction regime may have been a necessary condition for rifting.

DESCRIPTION AND REGIONAL EXTENT

The Rio Grande rift extends as a well-defined series of asymmetrical grabens from Leadville, Colorado, to Presidio, Texas, and Chihuahua, Mexico, a distance of more than 1,000 km (Fig. 1). North of Socorro the rift is a distinctive morpho-tectonic feature. The main rift grabens have undergone vertical structural offsets of as much as 6 km (e.g., near Bernalillo). South of Socorro the rift is not physiographically distinctive, yet can be distinguished from the adjacent Basin and Range province by a variety of geologic and geophysical signatures (Seager and Morgan, 1979).

Over much of its length the rift is part of a broad region of "rift-like" late Cenozoic extensional deformation, i.e., a region characterized by large crustal blocks separated by steeply dipping normal faults. In central New Mexico, west of Albuquerque and Socorro, this region is more than 200 km in width, extending southwestward across the physiographic Colorado Plateau to Springerville, Arizona, and perhaps farther (Baldridge and others, 1983). A broadly linear, northeast-trending array of late Cenozoic volcanic fields, commonly referred to as the Jemez lineament, separates this extended terrain (transition zone) along the southeastern margin of the plateau from the less deformed "core" to the northwest. The Jemez lineament must correspond to a major boundary or zone of weakness in the lithosphere. However, it is not an expression of a fault or fracture zone and does not correspond to any single, simple structure in the upper crust (Baldridge and others, 1983). Farther south, the extended region encompasses the entire Mogollon–Datil volcanic field. In southern New Mexico and northern Chihuahua, the rift is not physiographically distinguishable from the Basin and Range province extending across southern Arizona and southern New Mexico.

Although the style (i.e., extensional deformation) and timing of structural deformation of the entire extended region are similar to that of the main rift grabens, the magnitude of deformation is much less. Vertical offset on normal faults of the Colorado Plateau transition zone probably does not exceed a few hundred meters. In the Mogollon–Datil region of southwestern New Mexico, narrow, deep grabens exist near Datil, Reserve, and Silver City, but these are separated from the larger, main grabens of the rift by a region 100 km or so wide, in which extensional deformation formed only shallow grabens.

This recognition of a broad, "rift-like" region is fully in accord with definitions of the rift based on geophysical and topographic criteria (e.g., Cordell, 1978; Ander, 1981). Whether the term "Rio Grande rift" is applied to the entire region (e.g., Cordell, 1978), or is restricted to the deeper, physiographically distinguishable main grabens, as we prefer, is largely a semantic argument. The important point is that the continuity of structural style and timing across this broad region clearly indicates a continuity in underlying process. That is, formation of the Rio Grande rift and Colorado Plateau transition zone and breakup of the Mogollon–Datil volcanic field all resulted from, and were part of, the very widespread Basin and Range deformational event.

LITHOSPHERIC STRUCTURE

The fact that deformation associated with the Rio Grande rift affected such a large region suggests that it resulted from an event that perturbed the entire lithosphere. The structure of the lithosphere beneath the rift is anomalous in many respects (Fig. 2):

(1) A moderate amount of crustal thinning has taken place beneath the axis of the rift. At 35°N latitude, depth to Moho is 33 km beneath the Albuquerque–Belen Basin compared to about 45 km under the Arizona–New Mexico border and to 50 km under the Great Plains (Olsen and others, 1979). North–south interface dips are small (0–2°) so that uplifted Moho beneath the rift exists from the southern border of New Mexico northward to Colorado. Discontinuous reflection segments on COCORP lines within the rift basin have two-way travel times that correspond well to a depth of 33 km (Brown and others, 1979).

(2) The sub-Moho compressional velocity (P_n velocity) beneath the rift axis is 7.6–7.7 km/s, significantly lower than velocity of "normal" mantle (8.0–8.2 km/s) beneath the Great Plains. Gravity modeling suggests that a low-density layer, which we presume correlates with this low P_n material, persists westward from the rift beneath the Colorado Plateau at least as far as northeastern Arizona, where P_n has been directly measured at 7.8 km/s along the reversed Chinle–Hanksville line. Black and Braile (1982) argue that such low P_n values are primarily due to high temperatures in the mantle. Velocities of 7.6–7.9 km/s therefore strongly suggest that the asthenosphere is in direct contact with the base of the crust beneath the rift and southeastern Colorado Plateau, without any intervening normal mantle.

The very broad (long-wavelength) gravity low extending across the entire state (about 500 km) at the latitude of Albuquerque (Fig. 2) is related in part to the shape of the low-velocity zone in the upper mantle (the asthenosphere). Other east–west gravity profiles between latitudes 32° and 38° also show this asthenospheric diapir, which thus forms a ridge-like Moho upwarp approximately parallel to the surface trace of the rift (Cordell, 1978; Ramberg and others, 1978). Both heat-flow and electrical-conductivity measurements support this gravity picture of an asthenospheric upwarp. The axis of the asthenospheric low does not coincide with the axis of the rift, but instead is displaced about 200 km westward.

(3) The compressional velocity in the lower crust beneath the rift (6.4–6.5 km/s) is substantially less than lower-crust velocities (6.7–6.8 km/s) under the flanks both to the east and west. Lower-crust velocities on the order of 6.5 km/s are rather uncommon in continental North America, existing mainly in the Great Basin section of the Basin and Range province and in the Rio Grande rift (Prodehl, 1979; L. Braile and others, written comm. 1984). We interpret such subnormal lower-crust velocities to be mainly due to high crustal temperatures and to suggest that no major compositional difference exists between the lower

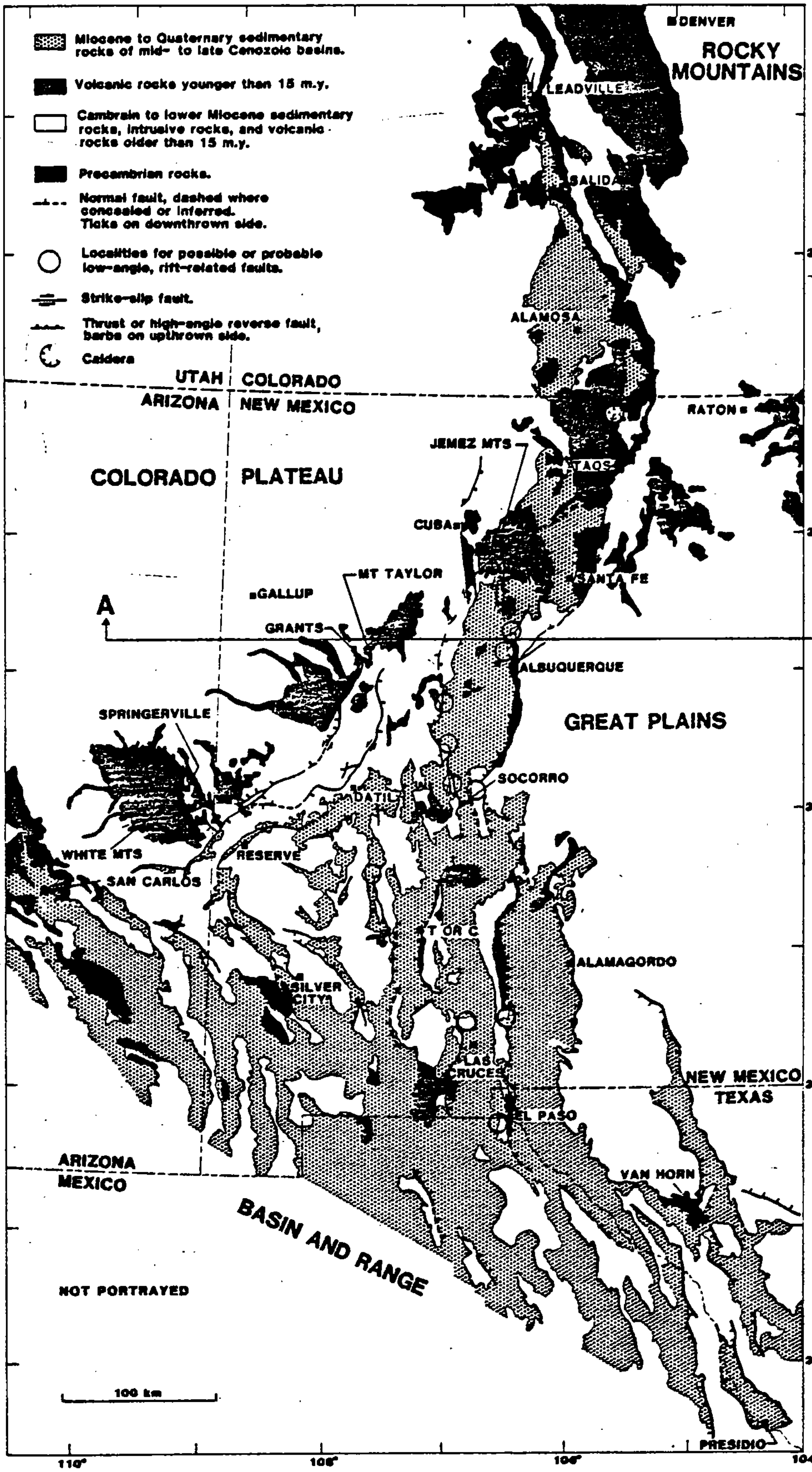


FIGURE 1. Generalized tectonic map of the Rio Grande rift. After Woodward and others (1978), Tweto (1978), NMGS Map (1982), and Baldridge and others (1983). A–A′ shows location of cross section in Figure 2.

Why evidence for so little volcanism in rift itself

East African Rift

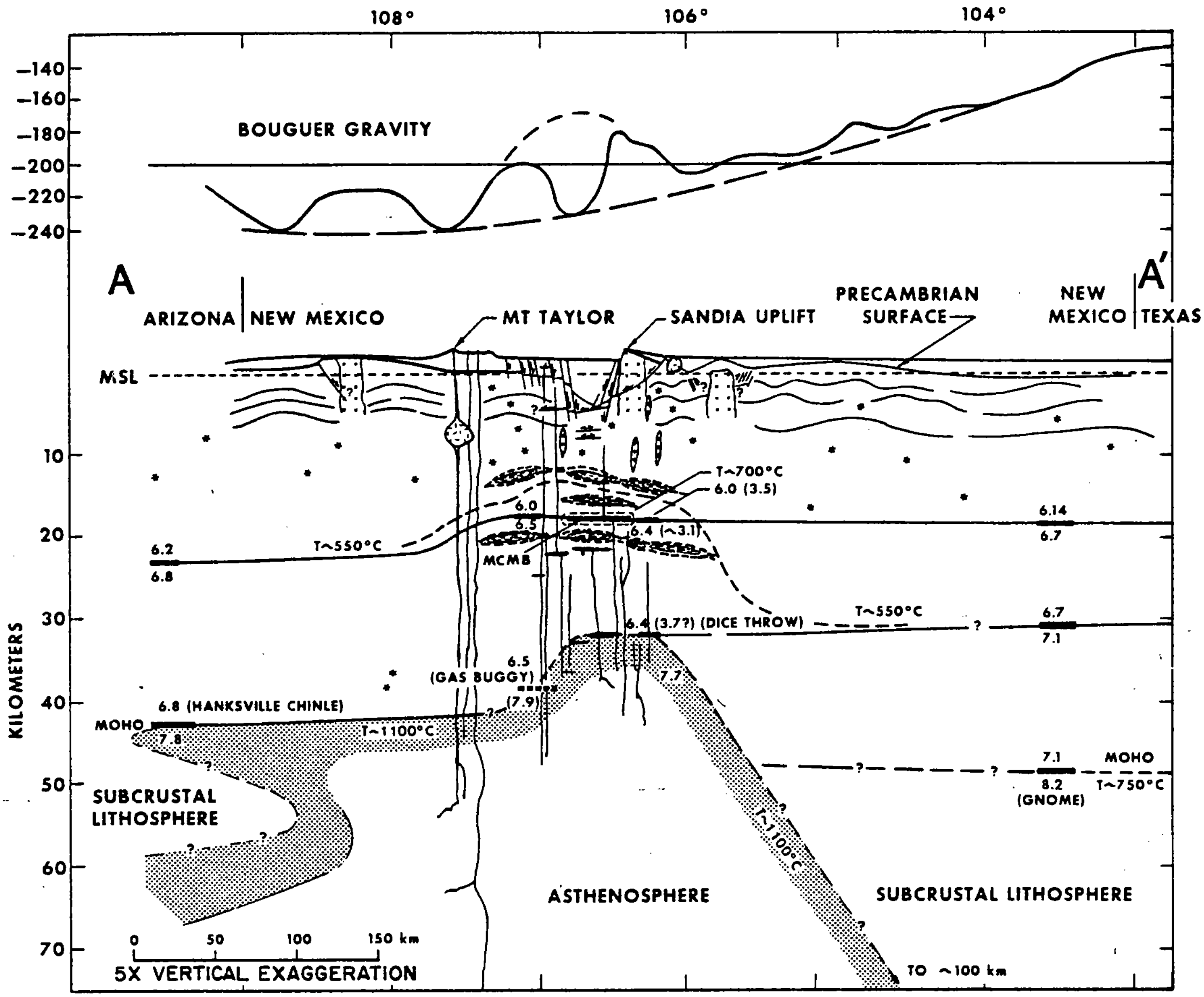
volcanism

volcanism

Use Wernicke simple shear extension

A

predict RGR will die
not getting pulled apart very quickly



IGURE 2. Cross section through northern Albuquerque–Belen Basin. See Figure 1 for location. Locations of intracrustal boundaries, interpreted from travel times, nd record-sections of correlated phases are plotted as short, heavy lines at the longitudes where seismic profiles intersect the cross section. Numbers are P-wave elocities in km/s (numbers in parentheses are S-wave velocities). Profiles are identified by locations of the shotpoints (Chinle–Hanksville: Roller, 1965) or by code ames of the main source explosions (DICE THROW: Olsen and others, 1979; GASBUGGY: Toppozada and Sanford, 1976; GNOME: Stewart and Pakiser, 1962). he Chinle, GASBUGGY, and GNOME profiles were obtained at a relatively early stage of U.S. crustal-profiling efforts, when station spacing was relatively coarse 10–50 km) and true amplitude/waveform data were inadequate to permit more than estimates of possible velocity gradients or fine structure in the principal crustal ayers. DICE THROW average station spacing was about 3 km, allowing better gradient estimates using modern synthetic seismogram-modeling techniques. ieneralized distribution of earthquake hypocenters in the upper crustal layer is shown by asterisks, as are the specific deep crustal events of 1976/1977 (Sanford nd others, 1979). A Bouguer gravity-anomaly profile at the latitude of the cross section is shown at the top (Cordell and others, 1982). The midcrustal magma ody (MCMB) at Socorro has been projected into the cross section. Locations of basaltic dikes, though schematic, are based partly on heat-flow data (e.g., Clarkson nd Reiter, this guidebook). Lensoidal "megaboudins" in the middle crust represent region of discontinuous ductile flow (see text).

crust beneath the rift and that adjacent to it. The intermediate-wavelength gravity high (Fig. 2) is often interpreted to indicate mafic igneous rocks intruded into the lower crust (an "axial dike" or, more properly, composite batholith). However, in the Rio Grande rift this intermediate-wavelength gravity signature can as well result from the equivalent replacement of lower-density crustal rock with higher-density mantle material because of the thinning of the lower crust. If there were a large intrusion of mafic material, such as a dike swarm, in the lower crust, we would expect a very considerable increase in density and hence a corresponding increase in seismic velocity (to 7.3–7.7 km/s) of the lower crust. In fact, the seismic results show a *decrease* in lower-crust seismic velocity compared to values in the flanks. Therefore, we do not consider the gravity high to be a persuasive argument for a large, lower-crust mafic-dike system; we believe crustal thinning can account for almost the entire effect.

The GNOME profile data (Stewart and Pakiser, 1962) also indicate a third (7.1 km/s), deeper layer beneath the usual 6.8 km/s lower-crust layer along the eastern New Mexico/west Texas border area. Deep, higher-velocity layers are commonly observed in craton and platform areas in other parts of the world and possibly reflect the growth of continental crust through such mechanisms as underplating. Another refraction line running eastward from GNOME (Romney and others, 1962) into central Texas was too sparsely instrumented to clearly show how far eastward this third crustal layer may extend.

(4) Unusually strong wide-angle reflections at precritical distances from refraction profiles in the southern Albuquerque Basin imply the presence of a thin, tabular, low-rigidity (i.e., abnormally low S velocities) layer at midcrustal depth (about 20 km) in the vicinity of the Socorro "midcrustal magma body" (Olsen and others, 1982). The refraction observations suggest that this anomalous sill-like layer probably surrounds, but is more extensive than, the partial melt or magma body between Socorro and Belen that has been extensively documented by Sanford and coworkers (Reinhart and others, 1979; Sanford and others, 1977). We tentatively correlate this feature with the boundary between the upper and lower crustal layers (Conrad discontinuity).

NATURE OF THE MOHO

Beneath the axis of the Rio Grande rift, existing seismic-refraction data (mainly from the DICE THROW profile) indicate that the Moho is a sharp transition at the resolution of 1.0–1.5 km provided by the dominant 1–10-Hz frequencies characteristic of regional earthquake and explosion sources. On the other hand, the more restricted-range COCORP-type deep-reflection experiments, where dominant frequencies are in the 10–80-Hz range, reveal a more complex, sandwich-like structure in the same region. Complex lamellae show considerable lateral variation, but the overall trend follows essentially the lower resolution "boundary" determined over much greater distances by refraction/wide-angle-reflection experiments. Thus, the few-hundred-meter-resolution reflection data indicate that the sub-rift Moho is a complicated structure which, in addition to representing a major petrological and compositional change with depth, has had a complex metamorphic and horizontal-stress history. The relatively small amplitudes of the P_n phase observed by refraction techniques along the Rio Grande rift axis indicate that the P-wave velocity in the asthenosphere actually decreases slightly with depth. Such decrease suggests that the sub-Moho temperature gradient is small and that nearly isothermal conditions apply in the top few kilometers of the asthenospheric upwarp.

Because refraction data for the adjacent Colorado Plateau and Great Plains are relatively sparse, with correspondingly more uncertain variations in amplitude vs. distance behavior of the relevant seismic phases used in analysis, we cannot describe details of the Moho transition and sub-Moho properties with the same confidence. However, by analogy with data from plateaus and shields in other parts of the world, we expect some significant differences from what has been inferred for the central part of the rift (Mueller, 1977).

NATURE OF THE CRUST

Our ideas concerning the nature of the upper crust and the mechanisms of crustal extension are greatly influenced by COCORP seismic-reflection profiling in the Socorro area and southern part of the Albuquerque–Belen Basin (Brown and others, 1979, 1980), and by observations of the Precambrian basement in New Mexico. Basically, we think that the upper and middle crust, and possibly the lower crust, are dominated by horizontal structures. COCORP data show a horizontally stratified upper crust, with local, transparent volumes which Brown and others (1979) suggest may be granitic plutons (e.g., line 2A). Below about 10 km are numerous discontinuous, subhorizontal reflection segments varying in length, dip, density, and amplitude. Brown and others (1979) suggest this type of signature would be expected from a highly deformed metamorphic terrain with a dominant subhorizontal fabric.

This interpretation fits well with our ideas concerning the structure of the upper crust, clues to which are present in the basement blocks surrounding the rift. The lithology and, to a lesser extent, the structure of the Precambrian basement of northern New Mexico are relatively well understood. Three major lithologic groups of Proterozoic age are exposed in the uplifted blocks of the central rift basin: volcanic and volcaniclastic rocks, quartzites and associated mature clastic sediments, and silicic plutons. With local exception, the volcanic sequence is oldest and is overlain structurally and perhaps stratigraphically by the quartzite sequence. Both sequences are intruded by younger batholithic to hypabyssal plutonic rocks. The sedimentary and volcanic rocks are folded into large (kilometers in wavelength) overturned folds, the axial-plane foliations of which yield the seemingly pervasive structural fabric of the basement. This fabric strikes northwest to northeast, with local easterly trends, and dips at a high angle. However, the folds commonly plunge less than 50°, suggesting that stratigraphic layering, although complexly folded, actually dips at a rather low angle. The presence of a low-angle lithologic boundary is also suggested by regional aeromagnetic patterns which show simple, broad, east-trending bands of contrasting magnetic intensity corresponding to outcrops of quartzite and metavolcanics (Zeitz, 1982).

Because of the high-angle fabric so obvious in outcrop, the basement appears (although incorrectly) to be characterized by a mosaic of angular blocks separated by high-angle zones of structural weakness. Late Miocene to Pliocene high-angle rift faults tend to follow basement fabric and reinforce this view. However, structural analysis of the basement suggests that the Proterozoic deformation produced large-scale horizontal transport of the lithologic units, giving rise to nappes and overthrust sheets that were subsequently folded into the high-angle structures. Thus, there was and is a pervasive low-angle fabric in the basement.

In extrapolating to the structure of the crust, the significance of Precambrian structure and lithology is the competence contrast produced by the juxtaposition of quartzite and volcanic terranes. At high levels in the crust, the quartzite is relatively brittle and contrasts strongly with the more incompetent metavolcanics and schists that surround it. Decollement and shear along stratigraphic boundaries are normal features of these terranes. Projecting downward into the crust, it is likely that strong layer contrasts will be found in the upper 5–10 km, and that horizontal fabrics will dominate where ductility contrasts along lithologic and structural boundaries are well developed. Perhaps the high-angle fabric exposed in the uplifted blocks is a manifestation of folding in an upper plate above a delamination surface of Proterozoic age. As previously noted, seismic-reflection data support the suggestion of low-angle lithologic contrasts in the deeper parts of the upper crust.

Strong evidence of a dominant horizontal structure in the middle crust within the rift, corresponding essentially to the Conrad discontinuity, comes collectively from: (1) unusual S-to-S-wave reflections from the horizontal interface from microearthquakes that overlie the reflector; (2) strong P-to-P-wave reflections from near-vertical-incidence COCORP profiling; and (3) unusually strong wide-angle reflections on refraction profiles that can best be explained in terms of an anomalously low S-wave velocity in the material just beneath the reflector interface (Olsen and others, 1982). Combined evidence from these various seismic techniques, plus supporting evidence from electrical, geodetic-leveling, and heat-flow measurements, have led to a widely accepted interpretation that this anomaly is due to a thin, sillke magma body or partially molten layer at midcrustal depths (18–20 :m) beneath the axis of the rift. This midcrustal magma body is most trikingly manifested beneath a 1,700 km² "central" area extending rom Socorro northward almost to Belen (Sanford and others, 1977; Rinehart and others, 1979). Within this region, Brocher (1981) has ummarized the evidence indicating the total thickness of this magma ody is of the order of only 1 km and consists of a highly complex, aminated series of sills of various degrees of fluidity and lateral "in-erfingering." Beyond the 1,700 km² central region, the wide-angle data Olsen and others, 1979, 1982) imply that the horizontal tabular ge-metry continues outward, but that the seismic-velocity anomaly is onfined almost entirely to subnormal S-wave velocities. Velocities for ' waves are almost indistinguishable from those in surrounding rocks t these depth ranges. This low-rigidity aureole suggests a metamorphic-eaction zone surrounding the central midcrustal-sill complex. A sur-rising and not well understood property of the magma body is its very orizontal upper surface, which is "flat" within the limits of resolution ±0.5 km) of the seismic techniques.

Similar, but sparser and therefore less conclusive, geophysical ob-ervations suggest the midcrustal anomalous layer may extend inter-uittently along the rift and/or into the flank regions. For example, ermance and Pedersen (1980) argue that magnetotelluric electrical-onductivity data favor a contiguous set of tabular units with a common eneric origin along the rift. Although the midcrustal layer is undoubt-dly not a continuous lithologic or petrologic unit throughout and in-olves only a small volume of magma compared to volumes of magmatic ocks in other rifts (e.g., the Kenya rift), the layer undoubtedly plays very important but little understood role in the style of Rio Grande ft volcanism. Thus, we have projected some of the features of the ocorro magma body into our cross section (Fig. 2) in order to em-hasize our perception of its importance, even though the section is bout 100 km north of the best data.

In the middle crust, but still within the seismic upper crustal layer ig. 2), is a transition from brittle to ductile behavior. The best evidence r the depth of this transition in the Rio Grande rift region is provided / the distribution of earthquake hypocenters. Although the precision measurements is lower for hypocentral depths than it is for horizontal ordinates (epicenters), almost all regional earthquakes recorded since 960 are from depths less than 20 km and most are probably shallower an 10–15 km (Sanford and others, 1979, 1981; Olsen, 1979). At rain rates applicable to crustal extension processes for the rift and asin and Range provinces, the level of the brittle–ductile transition ne is probably at depths where temperatures are in the range 350–00°C. In some local regions of high heat flow, such as beneath the ung (1.1–1.4 m.y. old) calderas of the Jemez Mountains, the brittle-ctile transition comes within only a few kilometers of the surface, eating large aseismic areas. Focal mechanisms of the shallow-crust rthquakes are consistent with E–W tensional forces. The only ex-ptions to the generally shallow depths of regional earthquakes were o moderate (magnitude ~4) events which occurred in 1977 near the se of the crust northwest of Mount Taylor. It is speculated (Sanford d others, 1979) that these deep events may be related to magmatic ocesses of the Jemez lineament/Mount Taylor system.

Curie temperatures for crustal materials (~550°C) are very similar the higher temperature range for the brittle–ductile transformation; erefore, buried structures inferred by aeromagnetic data all exist within brittle domain.

The nature of the lower crust is considerably more uncertain, not ly because of limits of seismic resolution, but also because there are t enough exposed sections of the lower continental crust world-wide give a representative picture of the range of lower crustal conditions. wever, there now seems to be a consensus (Smithson and Brown, 77; Smithson and others, 1981; Kay and Kay, 1981; Bott, 1982) that lower continental crust consists predominantly of granulite-facies tamorphic rocks having an intermediate bulk composition similar to desite or diorite. Presence of water (and other volatiles) in the deep ust is of fundamental importance in that it can greatly affect seismic d electrical-conductivity properties. However, definitive laboratory d field studies are so few that while these effects are currently being actively debated (Shankland and Ander, 1983), they remain unresolved. There is little doubt that the dominantly metamorphic nature of the lower crust implies a highly complex structural grain that we at present can describe only in general terms (Smithson, 1978). While we recognize these complexities in both the lower and upper crust, we have refrained from trying to cartoon these in Figure 2 because we do not want to have these misunderstood as specific, proposed features in the Rio Grande rift.

As pointed out above, seismic-refraction data show a lateral variation in average P-wave velocity of the lower crust from about 6.7–6.8 km/s beneath the flanks to 6.4–6.5 km/s beneath the rift axis (Fig. 2). In addition to chemical composition, several other factors, such as pressure, temperature, fluids, and anisotropy due to preferential alignment of certain minerals (usually olivine) or major cracks, can affect observed elastic-wave velocities. P-wave velocities increase with increasing pressures, especially at shallow depths (upper crust) where cracks and voids are closed. For compositions and pressures characteristic of the lower crust, laboratory measurements (Christensen, 1979) indicate the pressure derivatives of wave velocities are small enough so that P and S velocities should not be *increased* by more than about 0.1 and 0.05 km/s, respectively, throughout the 20- to 40-km depth range. Temperature coefficients are negative, however, and Christensen's (1979) measurement of various granulites showed that $-(\delta V/\delta T)_p$ is in the range of $0.5–0.7 \times 10^{-3}$ km/s/°C at approximate lower-crust pressure conditions. Interpretations of the differences in the geotherms between the axial region of the rift and the flanks beneath the Great Plains and (more certainly) the Colorado Plateau indicate that the flank–rift temperature difference in the lower crust may be about 300°C at ~20-km depths and more than 400°C at the 30–35-km level. Thus, granulite temperature coefficients in combination with high sub-rift temperatures are sufficient to decrease lower-crust P velocities from the observed values of ~6.7 km/s in the flanks to 6.4–6.5 km/s beneath the rift *without requiring any substantial changes in chemical composition* between the two regions. This result leads us to conclude that there is no persuasive evidence for a *massive* mafic-intrusive complex or dike swarm ("axial dike") in the lower crust beneath the axis of the Rio Grande rift, as has sometimes been postulated (from gravity data only) for the Kenya rift (Olsen, 1983). Certainly, lithospheric thinning beneath the central rift will generate some (predominately vertical) magma feeder conduits and local dikes that must be the main source of high average temperatures in the lower crust (e.g., Clarkson and Reiter, this guidebook). However, the volumes and masses of mafic material intruded into the lower crust must be less than a few (possibly ten) percent. Otherwise, both the lower-crust density and averaged seismic velocities would be *increased* substantially beneath the rift contrary to observations. The composition of the lower crust beneath the axis of the Rio Grande rift is a problem for further modeling studies.

The process by which large amounts (50–200%) of crustal extension are produced has been concisely described by Hamilton (1982, 1983). Hamilton's model identifies three crustal layers, each with distinctive mechanical behavior: brittle fracturing and rotation in the upper crust, discontinuous ductile flow in the middle crust, and laminar ductile flow in the lower crust. The brittle–ductile transition is near the top of the middle crustal layer. A significant feature of this model is the lenticular, transposed nature of the middle crust (see Callender, 1983, fig. 2), with lenses of more competent rocks ("megaboudins") interspersed with less competent material along ductile-shear zones. We think this hypothesis adequately explains the crustal evolution of the Rio Grande rift, and have cartooned this crustal structure in Figure 2.

As noted by Eaton (1982), the development of this type of crustal structure is strongly temperature and strain-rate dependent. In the Rio Grande rift, significant crustal extension is confined to late Oligocene/early Miocene (see next section), implying that the appropriate combination of high (convective?) heat flow, rapid strain rate, and resulting upward mass transport must have existed only at this time. Subsequently, the rift must have experienced a decrease in strain rate and upward mass transport (both plastic flow and magmatic upwelling) in a manner similar to that described by Eaton (1982) for the Basin and Range province.

higher angle younger
low angle older

Transport paths of magma through a crust altered by significant extension is presumably complicated. In the upper crust, brittle fracturing to depths of 10 km would allow relatively steep, near-surface conduits to develop, and would result in the aligned volcanic centers which are commonly observed in the Pliocene and Quaternary volcanic fields of the rift. However, middle crustal structure would be lensoidal and horizontally stratified, and the contact between middle and upper crust would probably be characterized by a low-angle detachment surface. This crustal environment might yield magma "pillows" and stratiform, sill-like intrusions in the middle crust, such as the Socorro magma body. It would also produce strong seismic reflectors near the brittle–ductile transition.

STRUCTURAL DEVELOPMENT

We must keep in mind that the picture of the Rio Grande rift presented to this point is simply a "snapshot" of an evolutionary process. Features of all ages, including those inherited from the pre-rift history, are superimposed. To understand the rifting process, we must understand structural and magmatic developments of the rift as a function of time.

The geometry of rift basins and Miocene to Pliocene high-angle faults in the Rio Grande rift has been characterized by numerous authors (e.g., Chapin, 1971; Chapin and Seager, 1975; Cordell, 1978; Seager and Morgan, 1979). Until recently, it was generally accepted that extension occurred dominantly on high-angle normal faults (e.g., Chapin, 1971; Woodward and others, 1975; Brown and others, 1980). These faults were thought by some to listrically shallow near the brittle–ductile transition (e.g., Woodward, 1977). Commercial seismic-reflection profiling in the rift (L. Russell, pers. comm. 1981), reinterpretation of the COCORP deep seismic-reflection profile (Cape and others, 1982), and recent structural analyses of rift-bounding faults (e.g., Seager, 1981; Chamberlin, 1983a; Lipman, 1983; Rhoades and Callender, 1983) now suggest this model is incomplete.

Instead, there appears to be a significant component of low-angle faulting in the Rio Grande rift. It is unclear whether low-angle faulting in the rift involves domino-style movement (Morton and Black, 1975; Chamberlin, 1983a), listric faulting (Cape and others, 1982), regional crustal shear (Wernicke and Burchfiel, 1982; Bird, 1979), detachment faulting (Frost and Martin, 1982; Callender and Zilinski, 1976; Rhoades and Callender, 1983), or some combination of the above models.

Figure 3 presents published examples of late Oligocene to Miocene low-angle faults in the Rio Grande rift. The north end of the Sandia uplift (Fig. 3A) plunges approximately 20°N. In down-plunge projection, the map of Kelley (1977) shows the San Francisco–Placitas fault zone to be a listric fault dipping west (L. Russell, pers. comm. 1981). Seismic-reflection profiling in the Albuquerque Basin west of the Sandia uplift supports this interpretation (L. Russell, pers. comm. 1982). In addition, recent work in the Placitas area confirms low-angle, northwest-dipping faults and shears within the Paleozoic section and at or near the Phanerozoic–Precambrian contact (J. Callender, unpubl. data). Rhoades and Callender (1983) described similar low-angle structures along the west side of the Sandia uplift, 10–20 km south of Placitas. In all cases, late Miocene to Pliocene high-angle normal faults truncate the low-angle structures.

Figure 3B shows the low-angle Carrizo fault of late Oligocene/early Miocene age (Callender and Zilinski, 1976) along the western edge of the Albuquerque Basin. The Carrizo fault is intruded by 27-m.y.-old intermediate and mafic hypabyssal rocks, suggesting the fault zone acted as a conduit for local magmatism. Similar structures have been described in the Joyita Hills on the eastern edge of the Socorro Basin, 60 km southeast (Rosen, 1983; Osburn and Eggleston, 1983; Smith, 1983). These low-angle faults are confined to the Phanerozoic section and are commonly associated with large ductility contrasts between stratigraphic units. The low-angle faults in the Lucero uplift and Joyita Hills are truncated by late Miocene to Pliocene high-angle normal faults.

One of the most carefully studied areas in the Rio Grande rift that contains numerous low-angle normal faults is the Lemitar Range near Socorro (Fig. 3C). Chamberlin's (1983a, fig. 5) restoration of this cross section (Fig. 3C) indicates nearly 200% of domino-style extension from late Oligocene to late Miocene time, which was contemporaneous with periods of silicic volcanism from 31 to 20 m.y. ago and from 12 to 7 m.y. ago. Late Miocene to Pliocene high-angle faulting and block-tilting are superimposed on this earlier deformation.

The Jeter fault in the Ladron Mountains (Fig. 3D), one of the first low-angle faults described in the Rio Grande rift (Black, 1964), has been characterized as a thrust fault (Black, 1964), low-angle normal fault (Kelley, 1977), domino-style fault (Chamberlin, 1983b), or detachment zone (T. Shackleford, pers. comm. 1982). It shows many of the characteristics of a regional upper-plate detachment surface (K. Gillespie-Nimick, unpubl. data), including intense brittle deformation in both sedimentary and Precambrian crystalline rocks. It is truncated by late Miocene to Pliocene high-angle normal faults and is commonly the locus for metallic mineralization.

Similar low-angle structures exist in the southern part of the Rio Grande rift (Fig. 3E–H). With the exception of the Franklin Mountains (Fig. 3H), Tertiary low-angle faults are clearly cut by late Miocene to Pliocene high-angle normal faults. In the Franklin Mountains, the age of low-angle faulting is uncertain, but comparison with the carefully studied Organ Mountains to the north (Seager, 1981) suggests that these faults are similar in style and age to late Oligocene to early Miocene low-angle faults in the Organ Mountains.

Additional low-angle structures have been described in the northern Rio Grande rift (e.g., Spiegel and Baldwin, 1963; Lipman, 1983; Muehlberger, 1979). As studies of intrabasin and basin-margin uplifts continue, undoubtedly many more such features will be found.

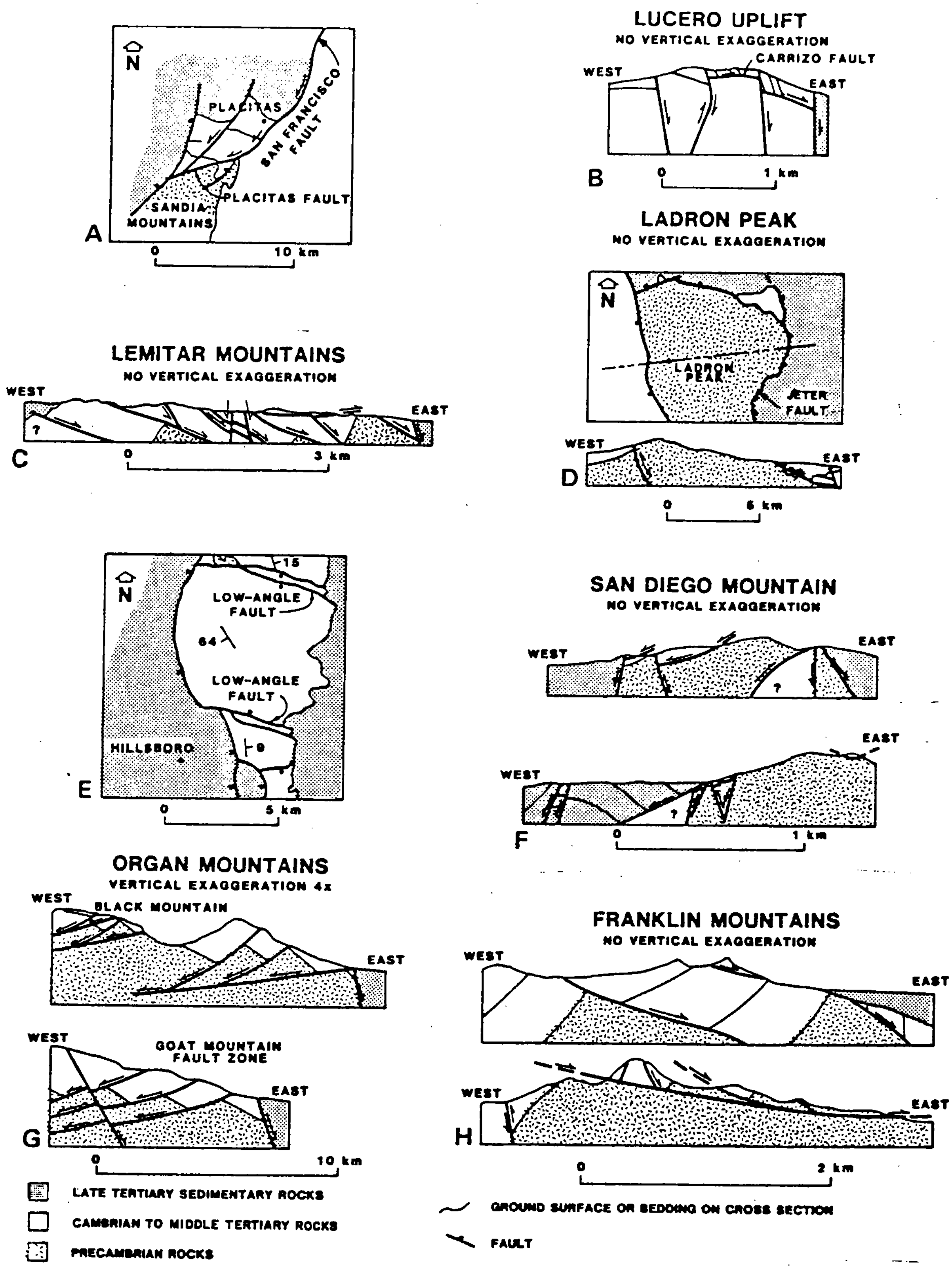
Thus, late Oligocene to early Miocene low-angle faults may be common features of the Rio Grande rift. They predate the high-angle normal faults of late Miocene to Pliocene age and suggest a major shift in structural style and mode of crustal extension with time. The low-angle structures overlap in time with the later phase of intense intermediate to silicic volcanic activity in southwestern New Mexico (Datil–Mogollon volcanic field), which culminated in early Miocene time. Abnormal crustal heat flow may have allowed significant high-level crustal stretching along low-angle structures. This may have been accomplished by the rise of the brittle–ductile transition in the upper crust. Convective-heat transfer, by massive intrusion of silicic magmas into the continental crust in the middle Tertiary (Elston, 1984), may have provided the thermal input to allow this stretching.

Hence, the structural development of the rift apparently occurred during two discrete time intervals: (1) an early phase characterized by low-angle faults, which began about 30 m.y. ago and lasted an estimated 10–12 m.y.; and (2) a later phase, dominated by high-angle normal faults, which extended from 9 or 10 m.y. ago to about 3 m.y. ago (e.g., Chapin and Seager, 1975; Baldridge and others, 1980; Seager and others, 1984). These phases of deformation are similar to extensional events in the Basin and Range province, where low-angle faults are typical of the early phases of crustal extension, in contrast to later high-angle deformation in the same area (Eaton, 1982; Frost and Martin, 1982). The similarities in style of deformation between the rift and the Basin and Range illustrate that both regions are manifestations of the same event.

Furthermore, possible widespread low-angle faulting early in the formation of the rift has major implications for the total amount of extension. All models of low-angle faulting demand a significant amount of crustal stretching. Extension of more than 200% is possible with this geometry (Chamberlin, 1983a), whereas high-angle normal faulting yields extension of only 5–30% (Woodward, 1977).

VOLCANISM

Most of the volcanism associated with the Rio Grande rift, with some important exceptions such as the Jemez volcanic field (Fig. 1), is basaltic and occurred less than 5 m.y. ago. [Volcanism in the Jemez field began more than 13 m.y. ago and includes enormous volumes of intermediate to silicic magmas (Smith and others, 1970; Gardner and Goff, this guidebook)]. These basalts consist of a variety of nepheline- and hypersthene-normative compositions which are generally similar to other late Cenozoic basaltic rocks of the western U.S. However, the most voluminous of the late Miocene to Recent basaltic volcanic fields, the Taos Plateau field, consists dominantly of a distinctive, low-alkali tho-



IGURE 3. Low-angle Neogene faults in the New Mexico part of the Rio Grande rift. See text for explanation. **A,** Map of northern Sandia uplift, after Kelley 1977); **B,** cross section through northern Lucero uplift, after Callender and Zilinski (1976); C cross section through Lemitar Mountains, after Chamberlin (1983a); ', map and cross section of Ladron Peak area, after Black (1964) and Chamberlin (1983b); **E,** map of the Hillsboro area, after Kuellmer (1956); **F,** cross sections ırough San Diego Mountain (Tonuco uplift), after Seager and others (1971); **G,** cross sections through east side of Organ Mountains (4 × vertical exaggeration), ter Seager (1981); **H,** cross sections through northern Franklin Mountains, after Harbour (1972).

leiitic magma (Servilleta Basalt), with lesser volumes of more alkaline (but still hypersthene-normative) magmas. Minor quantities of intermediate to silicic magmas represent melts formed by crystal fractionation processes and by melting of the lower crust. The Taos field apparently reflected a major thermal source which generated large quantities of magmas at or near the top of the mantle, and began to involve the crust in a major way (Lipman and Mehnert, 1979; Basaltic Volcanism Study Project, 1981, pp. 108–131; Moorbath and others, 1983; Dungan and others, this guidebook).

In the central and southern rift, Servilleta-type magmas were not erupted (except for minor quantities near Mt. Taylor) and lavas are not dominantly of any single composition. Neither do their compositions correlate in any unique way with their tectonic setting. (Perhaps because so few volcanic rocks occur, their compositional complexity is the more obvious.) In the Cerros del Rio volcanic field near Santa Fe, the next most voluminous late Cenozoic basaltic field, a variety of compositions were erupted: high-alkali tholeiite, basaltic andesite, alkali-olivine basalt, and basanite (Baldridge, 1979). Farther south, both tholeiitic and alkaline magmas were erupted (Aoki and Kudo, 1976; Kelley and Kudo, 1978; Baldridge and others, 1982). East and west of the main rift grabens (primarily along the Jemez lineament) volcanic fields include more alkaline compositions (nephelinites, basanites, mugearites) than occur within the rift, but also range in composition to include high-alkali tholeiitic rocks that characterize much of the central rift. Minor volumes of Servilleta-type magmas occur near Mt. Taylor (Fig. 1) (e.g., Baldridge and Perry, 1983).

Hence, no single composition is dominant along the rift; neither is there any evolution of composition with time nor unique correlation of composition with tectonic setting. In the Taos area, Lipman and Mehnert (1975) contrasted the Servilleta basalts to basalts outside the axial graben and proposed that the basalt compositions delineated the asthenospheric upwarp. That is, tholeiitic basalts in the axial graben were derived from very shallow depths (near the base of the crust) at the top of the upwarp, whereas more alkaline basalts outside the grabens were derived from the flanks of the uplift at much greater depths. We do not disagree with this model for the Taos area; certainly, our lithospheric profile (Fig. 2) is compatible with it. We emphasize, however, that the Servilleta basalts are compositionally distinctive compared to most other basalts in the rift (and perhaps unique compared to other basalts of the Southwest?) and thus indicate that the Taos Plateau volcanic field represents a unique and localized heat source. Hence we do not generalize the petrogenetic model of Lipman and Mehnert (1975) to the rift as a whole. We think instead that the situation is much more complex. The compositions of volcanic rocks in the rift as a whole seem to indicate that basalts were generated at various levels in the asthenospheric upwelling, implying an unintegrated heat source with only local melting. The fact that volcanic fields outside of the axial grabens include some compositions more alkaline than those inside suggests that at least some magmas from deeper in the asthenosphere are erupted to the surface.

Most workers agree that, in addition to the post-5-m.y. pulse of volcanism, an early pulse also occurred (about 30–18 m.y. ago), coinciding generally with early structural development of the rift. These two magmatic events were separated by a mid-Miocene lull in volcanism (Chapin and Seager, 1975; Baldridge and others, 1980; Seager and others, 1984). We also agree with this picture of two separate magmatic pulses, but emphasize that this earlier event—as a rift-wide phenomenon involving eruption of true basalts—was very minor (Fig. 4). Most of the mafic volcanism during this interval occurred in southwestern and southern New Mexico and involved basaltic andesites (e.g., Chapin and Seager, 1975; Seager and others, 1984). In this region, the enormous amount of basaltic-andesite volcanism was related not to rifting, but to (and was part of) the Mogollon–Datil volcanic field (erupted about 43–20 m.y. ago) of dominantly intermediate to silicic magmas (Chapin and Seager, 1975; Elston and others, 1976). This does not mean that rifting was not an immediate cause of at least some of this volcanism. The extensional tectonic environment that was part of the rifting event may have allowed magmas to escape to the surface, whereas in the pre-rift environment magmas were not able to ascend. [Alternatively, these magmas may have been able to reach the surface only after the upper-crust batholiths had solidified so that they could be brittly fractured (Eichelberger and Gooley, 1977.)] It does mean, however, that the heat source which generated these magmas was part of the previous (subduction-related?) event (e.g., Seager and others, 1984) and not of rifting. This volcanic (and thermal) event obviously must have affected the lithosphere and crust, introducing compositional changes (Elston, 1984) and perhaps, through convective-heat transfer (e.g., Spohn and Schubert, 1983), initiating the rise of the brittle–ductile transition in the crust. The distinction between a prior major thermal event and one specifically associated with rifting is important in recognizing thermal constraints on the rifting process.

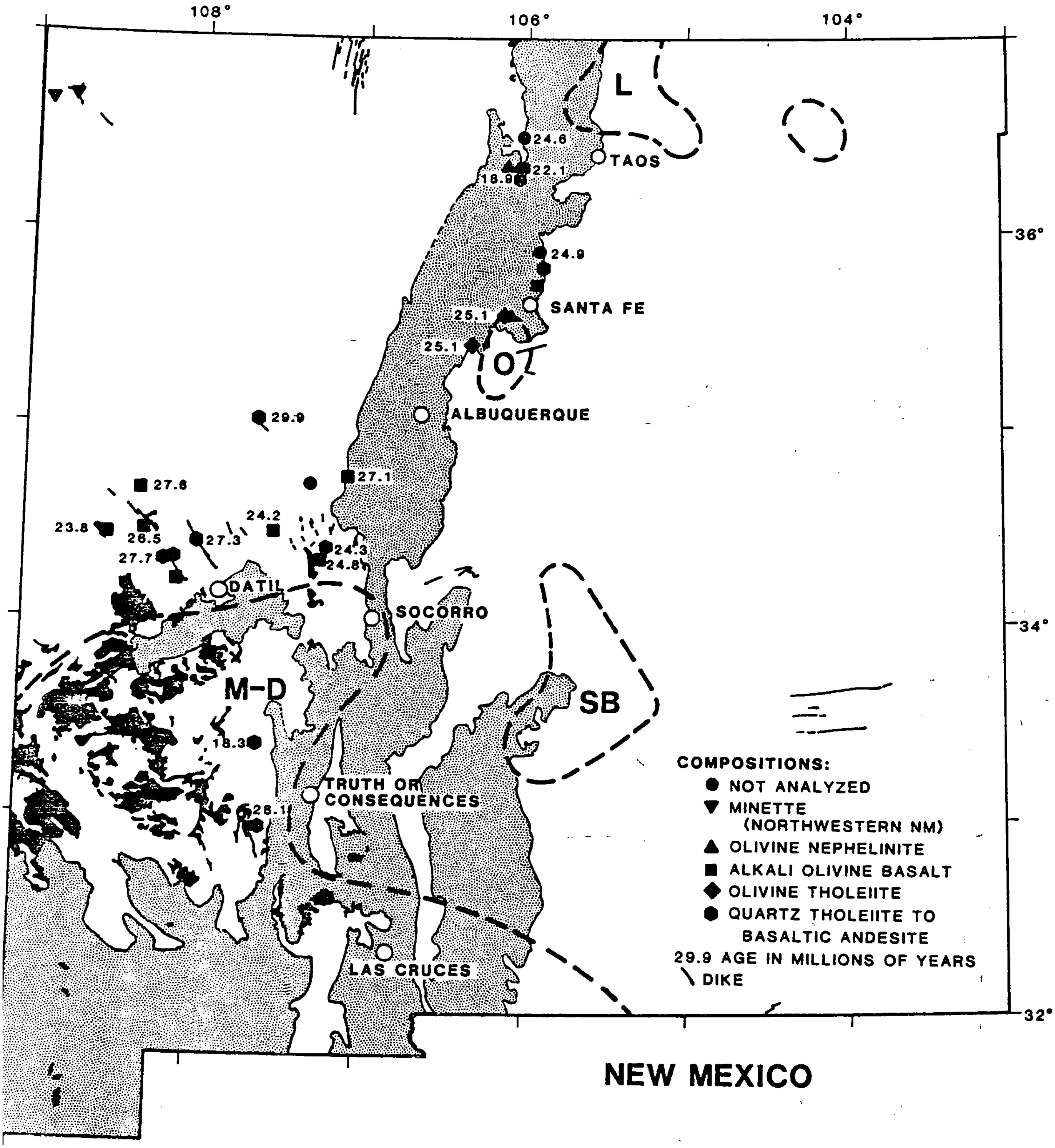
Much of the large volume of mafic lavas erupted 30–18 m.y. ago in southern and southwestern New Mexico is basaltic andesite related to the more silicic, calc-alkaline magmas. However, the compositions of magmatic rocks emplaced during this interval north of Socorro range greatly, with no obvious temporal or spatial pattern (Fig. 4). For example, olivine nephelinites erupted southwest and north of Santa Fe 19–25 m.y. ago (Baldridge and others, 1980) were derived from depths greater than 50 km, possibly as great as 90 km (Green, 1970a, b). Yet in the same general area, quartz-normative tholeiitic basalts, derived from much shallower levels in the mantle, were erupted approximately contemporaneously. This lack of a temporal pattern in magma composition confounds attempts to construct simple models of diapir growth beneath the Rio Grande rift, as was done for instance by Wendlandt and Morgan (1982) for the East African rift. Instead, the data seem to require an unintegrated heat source beneath the rift with only highly localized melting.

A striking feature of magmatism associated with the Rio Grande rift is the very small volume of volcanic rocks of any age in the rift. The largest fields within the rift grabens (Jemez and Taos Plateau volcanic fields) are part of the northeast-trending Jemez lineament (Fig. 1). Volcanic rocks are sparse in the rest of the rift. An essential question is, then, whether the low volume of volcanic rocks in the rift is due to the fact that magmas were never generated, or that they were never able to penetrate the crust and erupt on the surface. Based on evidence (largely seismic) presented above, we see no reason to think that large quantities of rifting-related magma were ever generated. If this conclusion is correct, it has enormous consequences for the magnitude of the thermal event accompanying rifting. For example, it would imply that no major heat source is associated with the rifting event and that little or no deep upwelling occurred beneath the rift. This conclusion is compatible with the diverse compositions of volcanic rocks in the rift.

LITHOSPHERIC PROCESSES

The goal of our studies is an understanding of the processes of rifting: what drives the lithospheric system, and what are the effects of rifting on the crust and mantle. Much discussion of rifting is cast into the simplified concepts of "active" and "passive" (e.g., Basaltic Volcanism Study Project, pp. 842–844). Certainly, our interpretation of the volcanism associated with the Rio Grande rift does not lead us to think that it is in any sense "active" (Fig. 5a). There are neither large volumes of volcanic rocks (of any age) in the rift, nor are there convincing arguments for large quantities of magmatic rocks at depth. Thus, though significant crustal thinning has occurred during rifting, this thinning has apparently been accomplished with little addition of heat into the system during the rifting process. A significant component of the modern, relatively high heat flow in the Rio Grande rift may be related to conductive-heat transfer from the middle Tertiary thermal episode, with the delay time dependent on crustal thickness and thermal time constant.

Pre-rift, subduction-related(?) volcanism, however, may have substantially changed the mechanical and lithologic character of the lithosphere and crust, providing an important component of convective-heat transfer into the crust and uplifting isotherms and the brittle–ductile transition. Rifting, particularly when low-angle faulting and large amounts of local crustal extension are involved in the upper crust, probably requires significant heat flow into the crust. Thus formation of the Rio Grande rift may have resulted from the (fortuitous?) combination of a thermally weakened lithosphere with a superimposed tensional-stress



ɔURE 4. Basaltic rocks inferred or known to have been emplaced between about 30 and 18 m.y. ago. Diagram differs considerably in degree of detail represented: ›rth of about 34° virtually every data point and outcrop are plotted; south of this latitude, outcrops are generalized and most data points are not plotted. Much ta exists, however, indicating that these rocks are generally basaltic andesites. In the Socorro and Las Cruces areas, large volumes are buried beneath rift sediments. oad, dashed lines show areas of Oligocene to Miocene (>43–20 m.y.), intermediate to silicic volcanism: M-D – Mogollon-Datil volcanic field, SB – Sierra anca, O – Ortiz, L – Latir (including "early rift volcanic field" of Lipman and Mehnert, 1979). Data from Sun and Baldwin (1958), Callender and Zilinski)76), Massingill (1977), Laughlin and others (1979, 1983), Lipman and Mehnert (1979), Baldridge and others (1980), Kautz and others (1981), NMGS Map)82), Lipman (1983), Seager and others (1984), and Vaniman (unpubl. data).

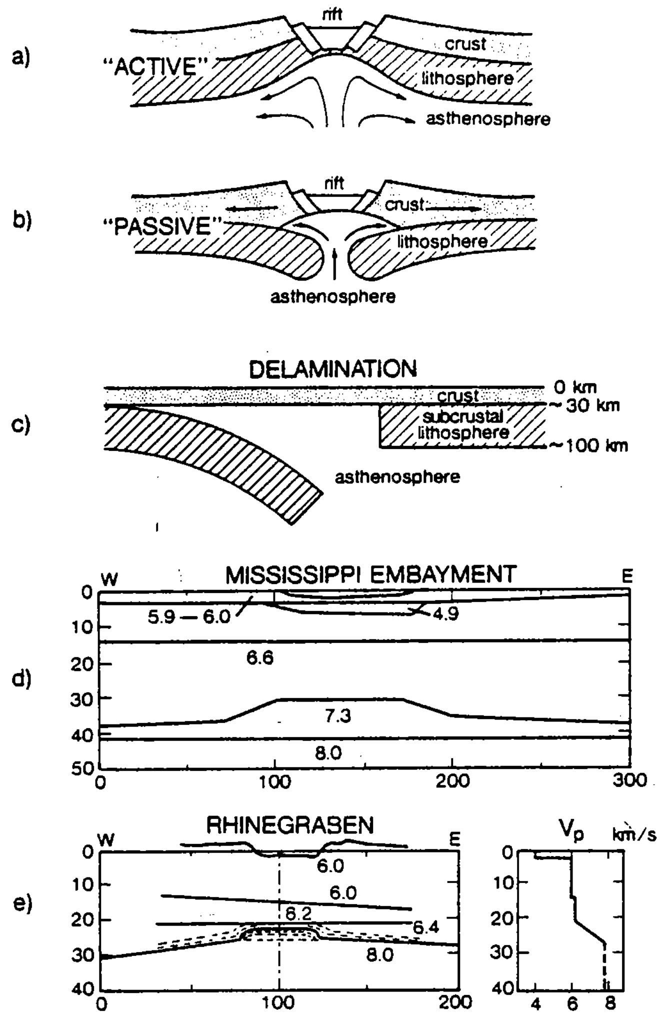


FIGURE 5. Schematic cross sections of the Mississippi embayment and Rhinegraben, and generalized models showing aspects of the rifting process.

regime. Middle Tertiary convective heat flow may have been critical in softening the crust and allowing it to yield.

Our lithospheric profile (Fig. 2) contains information that may pertain to mechanisms of rifting. One particular model for purely passive rifting (Turcotte and Emerman, 1983) postulates an asthenospheric "mushroom" geometry (Fig. 5b), in which asthenosphere is floored by subcrustal lithosphere with a probable compressional seismic velocity of about 8.0 km/s. As noted below, the axial seismic-refraction lines along the Rio Grande rift do not seem to support the mushroom or floored "pillow" picture, perhaps because existing refraction lines coincidentally overlie the "broken window" in the subcrustal lithosphere. Additional profiles in the near flanks might possibly reveal such structures. We feel, however, the existing deep seismic and gravity interpretations of the Rio Grande rift are much more in accordance with the broadly domed picture of Figure 5a than with the mushroom concept. Preliminary interpretation of teleseismic delays by Davis and others (this guidebook) suggests the asthenospheric low-velocity "upwarp" anomaly broadens with depth and may extend to depths of 200 km or deeper beneath the rift in northern New Mexico and southern Colorado. There is, however, an important difference from the model of Figure 5a in that the Rio Grande system is very asymmetric with upper-mantle/asthenospheric anomalies extending westward under the Colorado Plateau (Ander, 1981).

Thus, the concepts of active and passive rifting are obviously highly oversimplified, and probably neither is directly relevant to the Rio Grande rift. A most important feature of the rift is that it is superimposed on a region which underwent a long and complex history of deformation and heating prior to rifting. In order to understand the rifting process, it is imperative to sort out phenomena uniquely associated with rifting (i.e., with crustal extension and lithospheric thinning) from those inherited from previous events. We do not see convincing evidence for major, active, deep lithospheric involvement during rifting. Instead, the discontinuous nature of rifting and the fact that the rift is part of a very broad region of deformation suggest that plate processes exert a dominant control on the formation and evolution of the Rio Grande rift. As we have stated above, however, this does not imply that the major thermal and magmatic event preceding rifting did not influence, or perhaps even control, the initiation of rifting.

The significance of the Rio Grande rift to the evolution of the continental lithosphere is illustrated by contrasting it to other rifts. Based on gravity modeling and several axial and crossing seismic profiles, Mooney and others (1983) derived a crustal model for the northern Mississippi embayment (Fig. 5c). This area is the site of a buried late Precambrian rift that was reactivated in the Mesozoic. Mooney and others (1983) note the presence of the 7.2-km/s "domed" layer, which implies the lower crust in the rift vicinity has been altered by injection of mantle material. They suggest such a "high-to-intermediate"-velocity intrusion or lower-crust alteration may be a ubiquitous feature of continental rifts. A somewhat similar "pillow" feature has been inferred beneath the central region of the Rhinegraben (Fig. 5d), except there the rift pillow takes the form of a 5-km-thick velocity-gradient zone between 6.4 and 8.0 km/s rather than a more uniform 7.2–7.4-km/s layer. In the flanks outside the Rhinegraben, the Moho returns to a first-order discontinuity (Prodehl and others, 1976).

Although there is evidence for a 7.1-km/s deep crustal layer under the Great Plains (Fig. 2), axial-profile data within the Rio Grande rift and even to the west (DICE THROW, GASBUGGY, and Chinle–Hanksville) do not support the interpretation of such a floored, high–intermediate-velocity layer above an approximately 8-km/s mantle. We have made synthetic seismograms for floored models and compared them to DICE THROW data where P_n (at a velocity of 7.7 km/s) can be clearly observed *as a first arrival* between distances of 150 and 350 km. If there were an 8-km/s "floor" at depths of 5–15 km beneath the observed (33-km depth) 6.4–7.7-km/s transition, 8-km/s apparent-velocity headwaves and wide-angle reflections from the floor layer would be very strong and would severely mask the observed 7.7-km/s P_n phase and associated wide-angle reflections.

At present, the origin of the deep, approximately 7.1-km/s layer that appears to occur in some rifts is not well understood. This layer may represent lower crust that has been densified by intrusion of basaltic material, such as in magmatically active rifts (e.g., the East African rift). Its presence in the fossil Mississippi embayment might also suggest that such velocities are attained after significant cooling of the lower crust. The Rhinegraben may represent an intermediate evolutionary stage between a fossil system and a currently dynamic system such as the Rio Grande rift.

Bird's (1979) suggestion of continental delamination (Fig. 5c) may possibly help account for the asymmetry of the Rio Grande rift/Colorado Plateau system. The lower continental (subcrustal) lithosphere is in unstable mechanical equilibrium because it is denser than the underlying asthenosphere. If some process, such as fault breakage throughout the subcrustal lithosphere or plume erosion in the vicinity of the present rift, provided a line of weakness and an elongated conduit connecting asthenosphere with the base of the crust, the denser subcrustal-lithosphere boundary layer could peel away and sink into the chemically equivalent but hotter and less viscous asthenosphere. Observed moderately low P_n velocities of 7.8 km/s extending from the rift westward under the Colorado Plateau (Fig. 2) strongly suggest the existence of hotter asthenospheric material in contact with base of the crust, perhaps as far westward as the Arizona–New Mexico border. Bird (1979) suggests several possible geological and petrological consequences of this model that need to be explored in greater detail in connection with rift evolution.

CONCLUSIONS

Rifting is a long-term process involving changes in the entire lithosphere. Manifestations in the crust of the rifting process are a relatively urficial part of a deeper process. Geophysical studies yield information on the deeper crust and mantle, whereas geological and geochemical tudies can provide an historical record of rifting. It is important in all of these studies to focus on the whole process of rifting.

The evolution of the Rio Grande rift does not fit either simple active or passive models. For example, no compelling evidence exists for a najor thermal event associated with rifting. Yet heat, inherited from he immediately preceding deformation regime, was a critical factor in, nd may have been a necessary condition for, rifting.

Even though volcanism and extension associated with the Rio Grande ift are relatively minor, the lithosphere beneath the rift is being pernanently altered. For example, compositional changes in the lithophere probably lead to a density increase. The effect of rifting will robably not be continental splitting. Rather, we suspect that the lithopheric anomaly associated with the Rio Grande rift will ultimately ead to major basin formation.

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