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DEEP STRUCTURE OF THE RIO GRANDE RIFT FROM SEISMIC REFLECTION PROFILING

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Abstract. Seismic reflection surveys carried out by the Consortium for Continental Reflection Profiling (COCORP) in the Rio Grande Rift near Socorro, New Mexico, have successfully mapped large-scale structural variations down to the base of the crust. A total of 155 km of 24-fold common reflection point coverage was obtained parallel and transverse to major rift structure, including a continuous east-west profile which spans the central rift from the Sierra Lucero on the west (at the edge of the Colorado Plateau) to the Manzano Mountains on the east. Stacked seismic reflection sections derived from these surveys indicate coherent reflected energy from depths as great as 35 km, with substantial evidence of structural complexity extending throughout the crust. Among the major features apparent on these sections are (1) large-scale relief on the floor of the Albuquerque Basin, including a major intragraben horst standing over 3 km above the surrounding deep basin floors, and an extensive, shallow structural bench beneath the southeast portion of the basin, (2) well-developed, antithetic Tertiary normal faulting on the east side of the rift, (3) marked lateral and vertical variation in intrabasement reflection character, including relatively long correlatable events, distinct bands of short, discontinuous reflections, and seismically transparent zones characterized by few, if any, reflections, (4) a complex band of reflections from depths appropriate for the Moho, and (5) unusually strong, coherent reflections from a depth of approximately 20 km, corresponding closely with the top of an active magma body previously postulated to exist beneath the rift in this area on the basis of microearthquake studies. The magma body reflections are among the most prominent deep events yet recorded by COCORP; their geometry and relationship to surrounding structural elements provide important constraints on the depth of brittle faulting and new insight on magma migration and accumulation. Extensional faulting appears to be primarily near vertical near the surface with no clear evidence of large scale listric faulting at depth. The eastern rift boundary at Abo Pass is marked by a high-angle planar zone devoid of coherent reflections, possibly representing intense structural disruption extending through the crust. In contrast, the seismic section across the western rift boundary is defined by a gently dipping, relatively simple reflection of uncertain origin, with little evidence of a throughgoing fault in the lower crust. The seismic characteristics of many of the intrabasement features suggest particular igneous and/or metamorphic origins and support models which emphasize lateral and vertical heterogeneity at depth.

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Introduction

Recognizing the need for a higher-resolution view of the continental basement, the Consortium for Continental Reflection Profiling (COCORP) is systematically applying state-of-the-art multichannel seismic reflection techniques to the study of continental crust [Oliver, 1978]. The seismic reflection technique as practiced today has, in general, the highest resolution of any geophysical method for the study of most varieties of earth structure and is widely used in the petroleum industry to explore the sedimentary section. Until relatively recently, however, seismic reflection studies of the lower parts of continental crust have been rare [Oliver et al., 1976]. COCORP began its field program with an experiment in Hardeman County, Texas, in March 1975. Reflections and diffractions from lower crustal depths were identified and interpretated by Oliver et al. [1976]. Partly because of the success of this work, a second field experiment was carried out at the eastern boundary of the Rio Grande rift at Abo Pass, New Mexico, in the fall of 1975 (Figure 1). Since these early tests, the COCORP program has moved on to profile a wide variety of geological problems successfully [Schilt et al., 1979].

The Abo Pass surveys consisted of two profiles. Line 1 (29 km surface coverage, Figure 1) began in the Tertiary-Quaternary fill of the Albuquerque Basin, then ran east through Abo Pass onto the Paleozoic formations dipping east off the Manzano and Los Pinos mountains. A shorter profile, line 2 (9.4 km), was run transverse to line 1 and subparallel to the general strike of major rift structures to provide three-dimensional control. Preliminary analysis of the seismic records from these surveys confirmed that the effort was successful in obtaining deep reflections and emphasized the desirability of extending the surveys to include other components of the rift [Oliver and Kaufman, 1976]. Therefore in the fall and winter of 1976-1977, fieldwork was resumed in the area. Four new lines were obtained and designated the Socorro surveys (Figure 1). Line 1A (51 km) overlaps Abo Pass line 1 and extends coverage to the west and northwest across the Albuquerque Basin to the Sierra Lucero, at the edge of the Colorado Plateau. Line 2A (33 km) extends line 2 southwest toward Socorro. Lines 3 (20 km) and 4 (13 km), conducted west of Socorro in the La Jencia Basin, are not treated in this paper.

In spite of the considerable attention given to continental rift systems, a number of fundamental questions regarding their deep

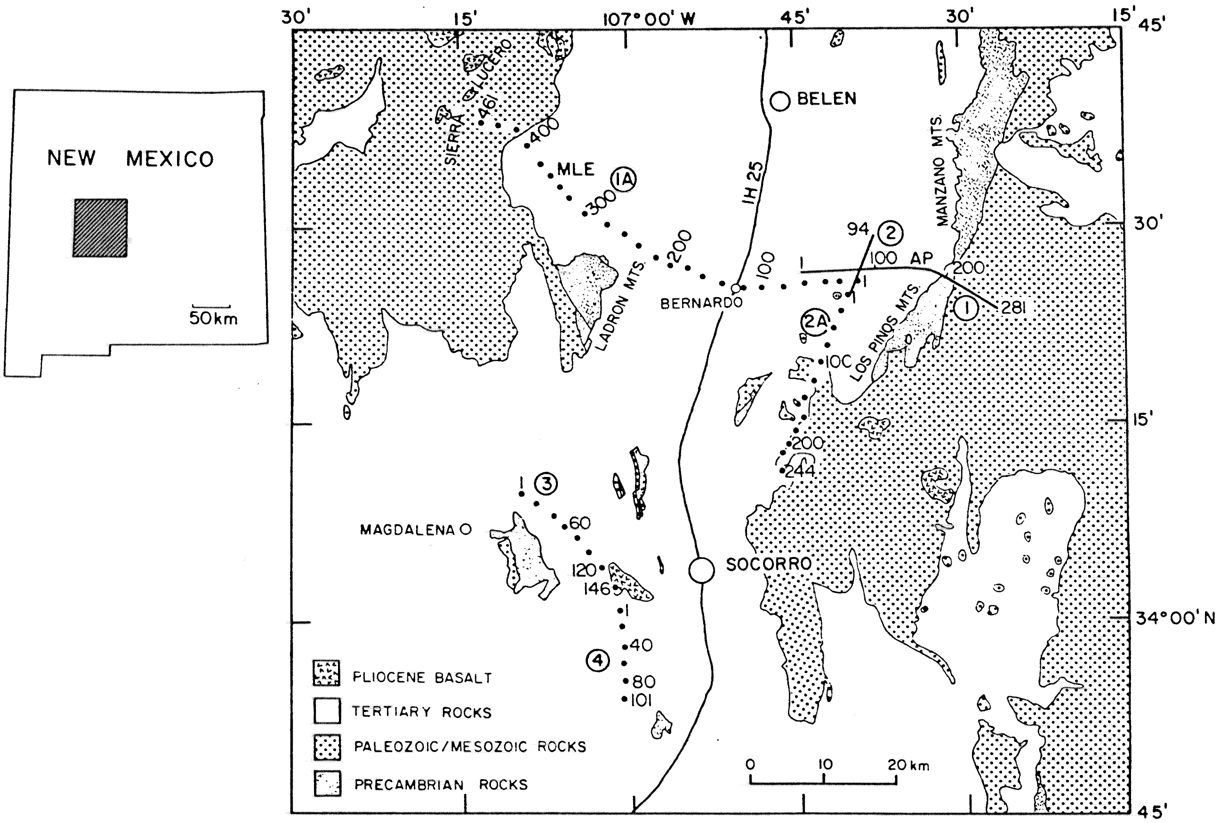

Fig. 1. Location of COCORP seismic reflection lines within the Albuquerque Basin of the Rio Grande rift. Line numbers are circled, and vibration points labeled. MLE is Monte Largo embayment, AP is Abo Pass. Geology after Dane and Bachman [1965]. Socorro lines 3 and 4, in the La Jencia Basin west of Socorro, are not treated in this paper.

structure have yet to be satisfactorily answered. For example, do the main rift boundary faults decrease in dip with depth? Are rift structures controlled by preexisting crustal weaknesses? How is magma transported and emplaced in continental crust? How does rifting affect the crust-mantle boundary? Since previously reported multichannel, near-vertical, deep seismic profiles across major intracontinental rifts are very few [e.g., Dohr and Meissner, 1975], the seismic reflection records collected by COCORP in the Rio Grande rift offer substantial new information bearing upon these and other problems. Brown et al. [1979] have reported briefly on some of this work. In this paper, the results of COCORP reflection surveys along Abo Pass lines 1 and 2 and their extensions (Socorro lines 1A and 2A) are presented and discussed more fully, with emphasis on the nature of (1) the rift boundaries, (2) intrarift faulting, (3) the Moho beneath the rift, and (4) an important midcrustal reflector believed to represent the top of an active magma body.

Geologic and Tectonic Setting

The Rio Grande rift is a NNE trending crustal break following the southern Rocky Mountain tectonic belt which separates the Great Plains on the east from the Colorado Plateau and Basin and Range on the west. Chapin [1971, 1979] summarizes many of the general aspects of the geology of the rift. Extending for over 1000 km from northern Chihuahua, Mexico, to the northern end of the upper Arkansas valley in Colorado, its constituent basins vary from 16 to 64 km in width [Kelley, 1956]. South of the upper Arkansas valley the rift exhibits a significant asymmetry, with greater structural relief along the east side. According to Condie and Budding [1978] (see also Foster and Stipp [1961]), the near surface Precambrian terrain of central and south-central New Mexico is dominated by granitic plutons (75%) intruded into metamorphic rocks, including schists (10%), quartzite and arkosite (8%), mafic metaigneous rocks (4%), siliceous metaigneous rocks (2%), and gneisses (1%). Radiometric ages for the Precambrian rocks of New Mexico range from 1.0 to 1.8 b.y. [Muehlberger et al., 1967; Condie and Budding, 1978]. Overlying the Precambrian basement in most parts of the rift is a cover of Paleozoic and Mesozoic sedimentary rocks of variable thickness, itself overlain in the basins by significant thicknesses of Tertiary and Quaternary fill, ranging from 120 m in the Estancia Basin to as much as 4500 m in the San Luis Basin [Davis et al., 1978].

The COCORP surveys were conducted in the southern part of the Albuquerque Basin, about 40 km north of Socorro. Here the basin is flanked on the west by the westward dipping Paleozoic rocks and Tertiary intrusives of the Sierra Lucero [Kelley and Wood, 1946] and bordered on the east by the Precambrian and Paleozoic rocks of the Los Pinos and Manzano mountains [Stark and Dapples, 1946; Reiche, 1949; Stark, 1956]. The regional geology of this area has recently been reviewed by Kelley [1977]. North of Bernardo (Figure 1) the rift consists of a series of linked structural depressions with raised margins, while to the south it broadens into a series of basins and intergraben horsts [Chapin and Seager, 1975]. Whether the southern Rio Grande rift merges with the Basin and Range province or retains a unique crustal identity is still debated [Seager and Morgan, 1979].

The Rio Grande rift was formed by crustal extension beginning about 32-27 m.y. ago [Chapin, 1979] and continuing to the present, with late Cenozoic rift structures apparently being superimposed upon late Paleozoic and Laramide uplifts [Kelley, 1952, 1977; Chapin and Seager, 1975; Woodward, 1977]. Volcanism concurrent with extension has occurred mainly along the medial axis and west side of the rift, with periods of greatest activity being from 32 to 20 m.y. ago and from 5 m.y. ago to the present [Chapin, 1971; Chapin and Seager, 1975]. Evidence that rifting is continuing at the present time includes a regional topographic bulge [Cordell, 1978], fault scarps cutting alluvial fans and Pleistocene surfaces [Sanford et al., 1972; Kelley, 1977], seismic activity [Sanford et al., 1972, 1979], high heat flow [Reiter et al., 1979] with numerous thermal springs occuring along the western margin [Summers, 1968], and contemporary uplift [Reilinger and Oliver, 1976].

Previous Geophysical Studies

The Rio Grande rift has been the subject of a variety of detailed as well as reconnaissance geophysical investigations, many of which have been summarized by Cordell [1978]. Numerous geothermal studies [Decker, 1969; Decker et al., 1975; Decker and Smithson, 1975; Reiter et al., 1975, 1978, 1979] have demonstrated that heat flow in the rift is considerably greater than it is in the Colorado Plateau, Great Plains, or even the southern Basin and Range province. Decker and Smithson [1975] and Reiter et al. [1978] interpret the heat flow measurements to indicate the existence of hot, possibly molten, material within either the crust or upper mantle. Very high local heat flow values near Socorro have been interpreted by Reiter and Smith [1977] and Reiter et al. [1979] as evidence of molten rock at shallow depths. Geomagnetic and magnetotelluric studies [e.g., Schmucker, 1964, 1970; Warren et al., 1969; Porath and Gough, 1971; Hermance and Pedersen, 1978; Jiracek et al., 1979] support the inference of hot, partially melted material at shallower depths beneath various parts of the rift than beneath flanking regions. The existence of an upwarped 'cushion' of relatively low density mantle material beneath the rift is also supported by gravity [Decker and Smithson, 1975; Ramberg et al., 1978] and seismic refraction [Cook et al., 1979a] studies. Gravity measurements have also been the basis of other studies of rift structure [e.g., Sanford, 1968; Ramberg and Smithson, 1975; Ramberg et al., 1978; Cordell et al., 1978; Cordell, 1979]. A detailed analysis of gravity measurements near Abo Pass line 1 has been reported by Wongwiwat [1970].

Gross crustal structure along the rift, as determined by Toppozada and Sanford [1976] from refraction analysis of local explosions, earthquakes, and the Gasbuggy nuclear explosion, consists of an 18-km-thick upper crust (excluding the sedimentary cover) with a velocity of 5.8 km/s overlying a lower crust with a velocity of 6.5 km/s, with the Moho dipping about 2° northward (Figure 2). Depth to the Moho at the latitude of COCORP Line 1 is estimated to be 38 km on the basis of these data. Phinney's [1964] analysis of long period, body wave spectra for the Albuquerque region suggests a similar crustal thickness. More recently, Olsen et al. [1979] have derived a refraction model for the crust near Abo Pass from the DICE THROW explosions (Figure 2). This model consists of a 2 km thick sedimentary layer (3.3 km/s) over an 18.71 km thick upper crustal layer (6.0 km/s), with a 12.5 km thick lower crust (6.4 km/s). The total depth to the Moho near COCORP line 1 based on these results is therefore about 33 km, consistent with surface wave dispersion studies [Keller et al., 1979]. Such values are somewhat greater than the 30 km depths to Moho estimated for the southern Basin and Range [Prodehl, 1970; Warren and Healy, 1973] but considerably less than the 50 to 55 km depth estimates for the Moho beneath the Great Plains of eastern New Mexico [Stewart and Pakiser, 1962; Mitchell and Landisman, 1971].

The seismicity of the rift is generally limited to depths of less than 13 km and is indicative of normal faulting [Sanford et al., 1972, 1979]. Of particular interest are the results of microearthquake studies by Sanford and co-workers which indicate the existence of an active magma body beneath the rift in the vicinity of the COCORP surveys. Sanford and Long [1965] found unusually strong shear waves arriving late from microearthquakes, interpreted by Sanford et al. [1973] as reflections from a zone of partially molten material. Subsequent studies of reflected shear phases, swarm-like seismicity, shear wave screening, P wave delays, heat flow, and contemporary relative uplift support the conclusion that an extensive body of molten or partially molten rock exists beneath the rift in this area, possibly in association with local bodies of molten material at shallower depths (see Sanford et al. [1977] for a recent review of this evidence). The top of this magma body is at a depth of 20 km near COCORP lines 1 and 1A [Rinehart et al., 1979].

Data Acquisition and Processing

The field techniques employed by COCORP to obtain the seismic reflection data examined here were only slightly modified from standard VIBROSEIS (Trademark Continental Oil Company) survey practices used by oil exploration crews. Fieldwork and subsequent processing were executed by Petty-Ray Geophysical (a division of Geosource, Inc.) under contract to COCORP. The compressional wave source in these surveys consisted of five vibrators operating synchronously to transmit a sweep signal with frequency varying linearly from 10 to 32 Hz. The duration of each sweep was 20 s for the Abo Pass surveys (lines 1 and 2), with a total recording time of 40 s, resulting in 20 s of correlated reflection data. A sampling interval of 8 ms was

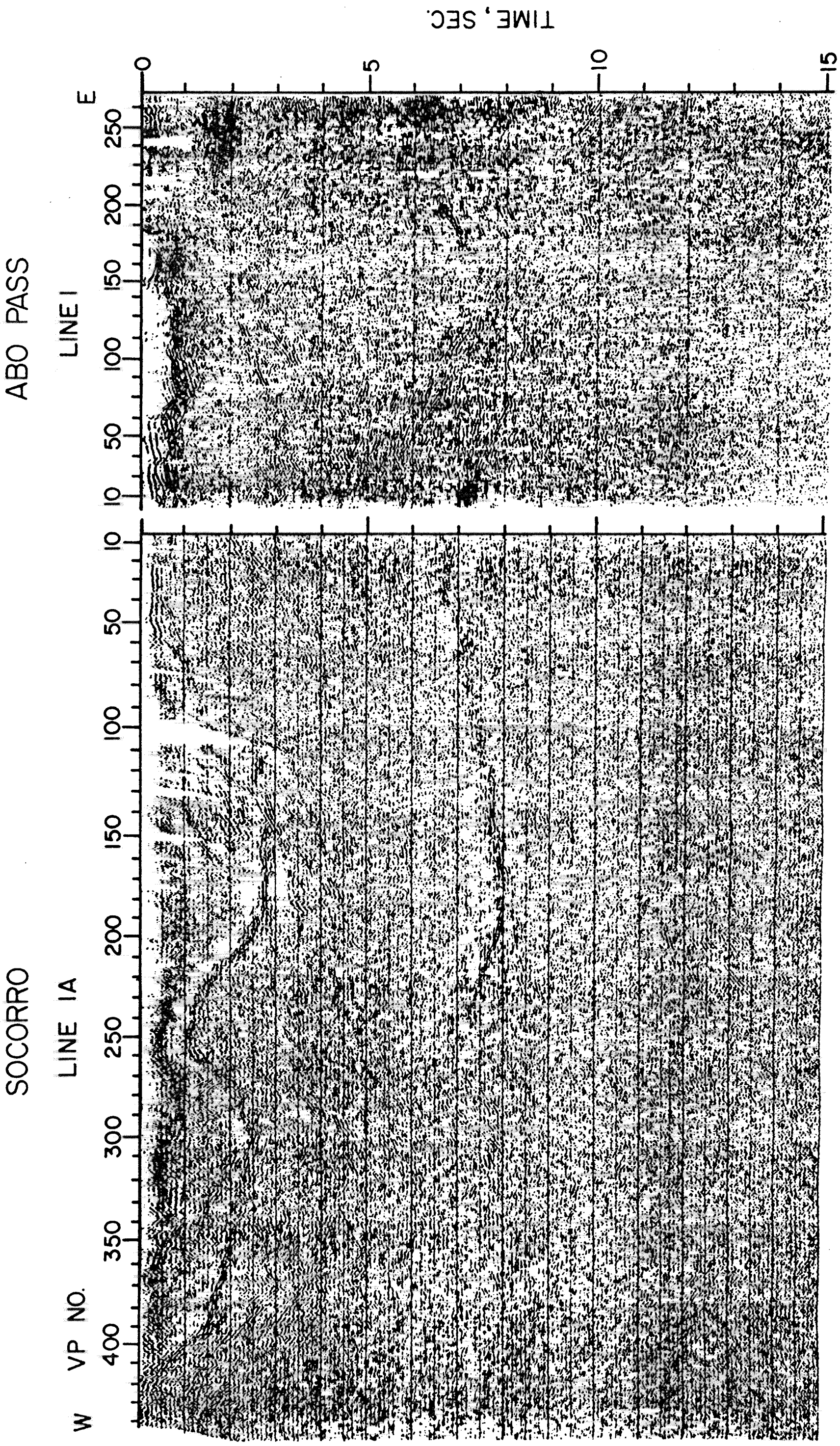

Fig. 2a. Seismic sections from lines 1 and 1A. VP spacing is 100 m for line 1 and 134 m for line 1A. Vertical scale in two-way travel time.

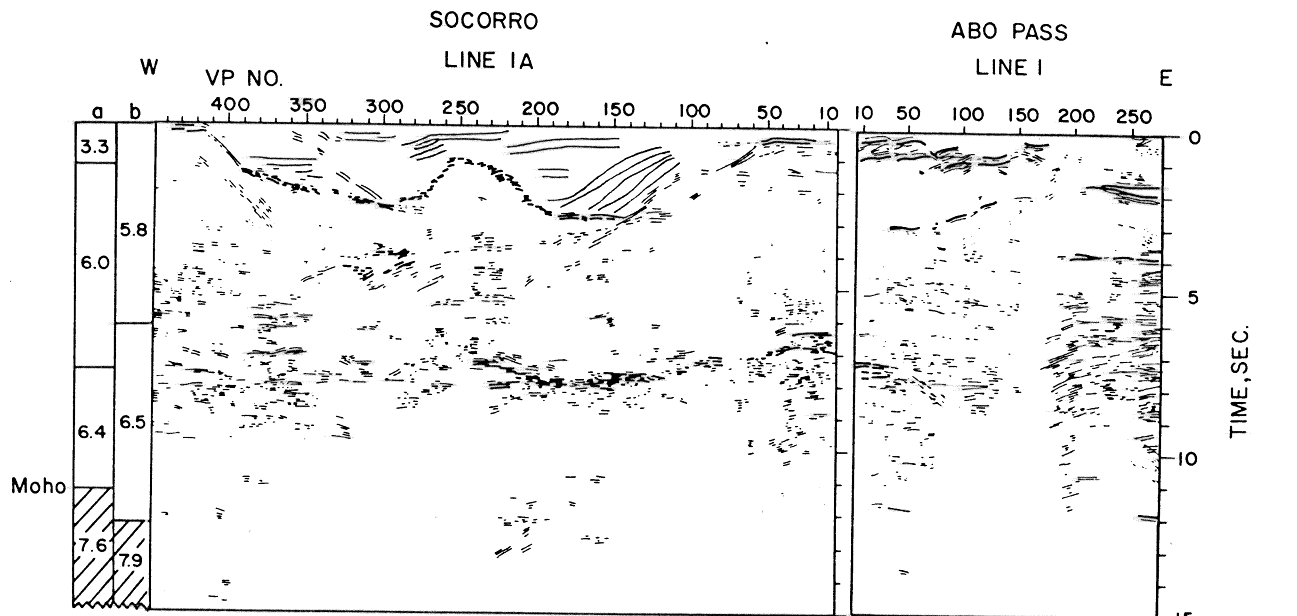

Fig. 2b. Line drawing emphasizing prominent features on the seismic sections from lines 1 and 1A. Refraction crustal sections a and b at left from Olsen et al. [1979] and Toppozada and Sanford [1976], respectively.

used during digital recording. For the Socorro surveys (lines 1A, 2A, 3, 4) the sweep was increased to 25 s with a total recording time of 50 s, resulting in 25 s of correlated reflection data. In both surveys, 16 separate sweeps were transmitted at each vibrating point (VP), and resulting recordings summed in the field. Vibrators were moved along 1/16 of a VP spacing after each sweep. The VP spacing was 100 m in the Abo Pass surveys and 134 m in the Socorro surveys. A 48 channel recording system was employed in both cases, and geophone spacing at each VP was coordinated with vibrator spacing to obtain optimum cancellation of horizontally propagating in-line noise (ground roll). Spacing and sweep parameters were determined by field tests carried out at the beginning of the Abo Pass work. Production profiling was executed in 24-fold common reflection point (CRP) mode with sources 'pushing' the recording spread. Sources were offset 600 m (6 VP) from the nearest geophone in the Abo Pass surveys, with a maximum source-receiver offset of 5.3 km (53 VP). In the Socorro work (lines 1A and 2A) a 670 m (5 VP)

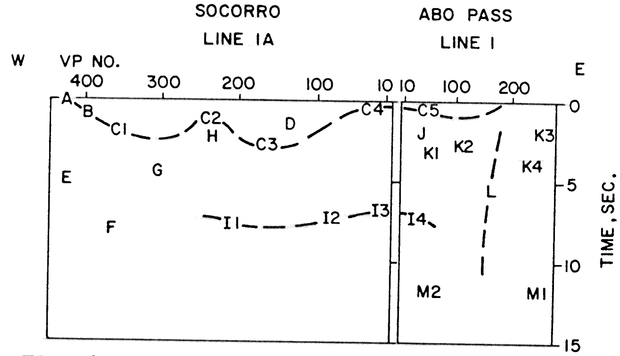

Fig. 2c. Index to features on lines 1 and 1A referred to in text.

offset was used, yielding a maximum source-receiver offset of 7 km (52 VP). Vibrating was done within the recording spread at the ends of Abo Pass lines 1 and 2 to increase multiplicity of coverage at those points. The primary departure from normal oil exploration methodology was in the use of longer recording times.

Once recorded, the reflection data were subjected to a standard sequence of signal enhancement procedures to produce the sections examined here. For the Abo Pass surveys the basic sequence included (1) demultiplex and trace edit with gain removal, (2) vibrator pilot sweep correlation and artificial gain restoration, (3) deconvolution, (4) brute stack, (5) field statics, (6) velocity analysis, (7) CRP sort, (8) application of dynamic corrections, (9) mutes application, (10) residual statics, (11) stack, and (12) display. The order of application of this sequence varied somewhat for the Socorro surveys. Unlike the treatment of the Abo Pass data, where gain removal was carried out in an early stage in the processing, the Socorro results were recorded and processed in true amplitude mode to facilitate amplitude analysis. Special statics analysis and migrations were carried out where required to aid interpretation of some of the data. Line 1A was subjected to additional processing by Digicon, Inc., to refine structural definition in the shallow part of the section.

The average velocities determined during the processing sequence were chosen by the contractor primarily to optimize the stacking process. Although these estimates render reliable interval velocities for the shallow sedimentary section, they provide little constraint for estimates of interval velocity at great depth. This is due not only to the extreme sensitivity of interval velocity estimates to small errors in stacking velocity measurements but to the relative lack of strong, continuous deep reflectors with which to define appropriate intervals, the short length of the recording spread relative to the greater depths of interest, and the extreme lateral variability of shallow structure as well. The lack of velocity control at great depth is compensated, at least as far as signal enhancement is concerned, by the relative insensitivity of the stack to velocity variations, given the length of the recording spread used. The more reliable velocity estimates from these analyses are cited (Table 1) as required by the interpretations.

Analysis and Interpretation

The geologic interpretations and inferences presented in this paper are based primarily on the seismic reflection sections shown in Figures 2 and 3 and the correlated field records from which they are derived. In spite of their similarity to geologic sections it must be remembered that they are unmigrated, two-dimensional time sections representing the seismic response of three-dimensional subsurface structure. As such, they may contain energy arriving from out of the plane of the survey, multiples, and converted phases, as well as exhibit distortion with respect to real earth geometries due to lateral velocity variations. In the particular case of the Rio Grande rift, 'statics' effects, i.e., differential travel time delays due to near surface structural complexity, are severe and have resulted in significant signal degradation in portions of the stacked sections. These various phenomena are known to cause ambiguity in the interpretation of normal reflection surveys of sedimentary sections [e.g. Fitch, 1976; Anstey, 1977] and are often even more problematical for deep surveys which must deal with lower signal to noise ratios, smaller receiver apertures, and, as will be argued, more complex reflector patterns [e.g. Brown et al., 1979]. Furthermore, it is not clear that the signal-processing techniques which have been highly refined by the petroleum exploration industry for use in sedimentary environments are optimum for application to deep seismic data, which are dominated by the metamorphic and igneous regimes of the lower crust and upper mantle. Undoubtedly, as the use of the seismic reflection technique for lithospheric studies becomes more widespread, new data acquisition and processing methods will be developed to enhance their particular characteristics [Phinney and Jurdy, 1979]. The horizontal, or distance, scales shown on the sections presented here are numbered in terms of vibrating points (VP's) along each profile. The vertical, or depth, coordinate is scaled in terms of two-way travel time. Time can be converted to an associated depth by multiplying by an appropriate 'average' velocity and dividing by two. Strict conversion was done here for only limited portions of the seismic sections in order to avoid introducing unnecessary inaccuracies from uncertainties in velocity estimates. For times greater than about 3 s, a suitable rule of thumb for conversion of time to depth is to multiply by a factor of 3, corresponding to an average velocity of 6 km/s. For the Tertiary sedimentary sections within the rift, velocities of 2.5-3.5 km/s are more appropriate. The scales used in Figures 2 and 3 were chosen to minimize horizontal or vertical exaggeration of the intrabasement portions of the

TABLE 1. Selected Deep Wells in the Albuquerque Basin

Well/Location/County/Offset	Formation	Depth, km	Velocity*, km/s	Depth, s	Thickness, km(s)
Humble 1, Santa Fe; 18-6N-1W; Valencia; 23 km NE from VP 360, line 1A	Santa Fe	surface	1.9	0	1.873(1.3)
	Baca	1.873	2.9	1.3	1.146(0.5)
	Cretaceous-Paleozoic	3.019	3.3	1.8	2.864(1.1)
	Precambrian (estimate)	5.883	4.1	2.9	
Shell 2, Santa Fe; 29-6N-1W; Valencia; 21 km NE from VP 360, line 1A	Santa Fe	surface	1.9	0	2.371(1.5)
	Baca	2.371	3.1	1.5	0.984(0.5)
	Cretaceous-Paleozoic	3.356	3.4	2.0	2.801(1.0)
	Precambrian	6.157	4.1	3.0	
Grober 1, Fuqua; 19-5N-3E; Valencia; 23 km N from VP 40, line 1	Qal/Santa Fe	surface	2.0	0	1.387(1.0)
	Tertiary, Baca	1.387	2.8	1.0	
Central New Mexico 1, Livingston; 16-3N-1E; Socorro; 7 km N from VP 130, line 1A	Santa Fe	surface	2.0	0.0	1.387(1.0)
	Cretaceous?	1.387?	2.8		

From Foster [1978]

*Root-mean-square velocity to top of formation, based on velocity analyses of seismic sections near the VP's indicated in the table.

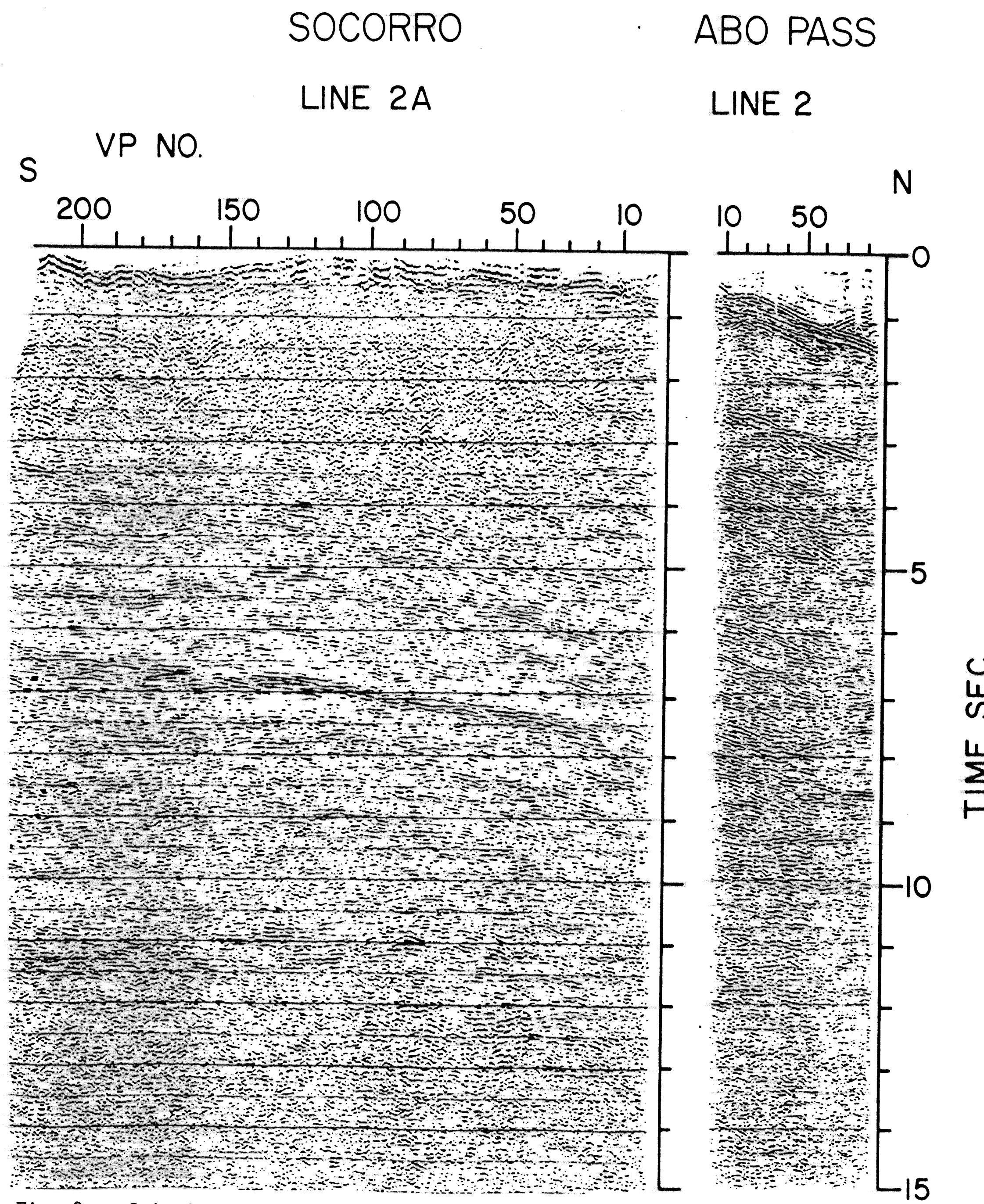

Fig. 3a. Seismic sections from lines 2 and 2A. VP spacing is 100 m for Line 2 and 134 m for line 2A. Vertical scale in two-way travel time.

sections. The sedimentary portions, with their lower velocities, therefore appear with a vertical exaggeration of between 1.5 and 2 to 1.

The Rift Boundaries

East

Line 1 (Figures 2 and 4) crosses the eastern boundary of the main body of the rift near Abo Pass (Figure 1). The Los Pinos fault, as described by Stark and Dapples [1946] and Kelley [1977], is taken to represent the edge of the central graben. Line 1 intersects this fault near VP 180. The character of the shallow portion of the seismic section changes dramatically as the line passes from the basin onto the exposed Precambrian rocks of the southern Manzano Mountains. West of VP 180 the shallow section is characterized by a well-defined, though pervasively faulted, set of correlatable reflectors (C5 in Figure 2; see also Figure 4). These reflectors abruptly end at the Los Pinos fault, with the shallow section beneath the uplifted flank relatively free of coherent reflectors. On the basis of their average reflection and refraction velocities [Jurdy and Brocher, 1980], their cumulative thickness (about

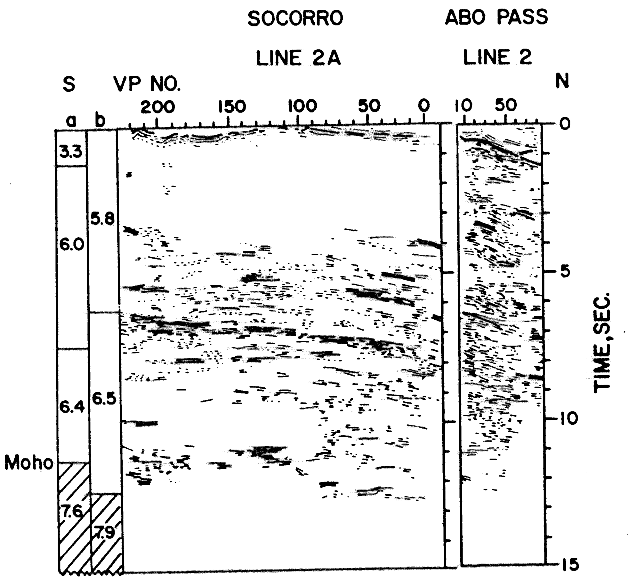

Fig. 3b. Line drawing based on seismic sections from 2 and 2A. Refraction sections on left same as in Fig. 2b.

1 km), and their correlation with events which crop out along line 2A, the shallow reflectors beneath the basin on line 1 are interpreted as part of the Mesozoic and Paleozoic section preserved under Tertiary fill (Santa Fe Group). Although there are no drill holes close enough to identify the top of the Precambrian basement, it is reasonable to infer that it lies at the base of this sequence of shallow reflectors (e.g., near 0.9 s at VP 10, line 1). Structure in the immediate vicinity of the Los Pinos fault is discussed by Stark and Dapples [1946] and Stark [1956] and summarized by Kelley [1977]. In the mountains east of this fault a sequence of Precambrian metasedimentary rocks, metarhyolite

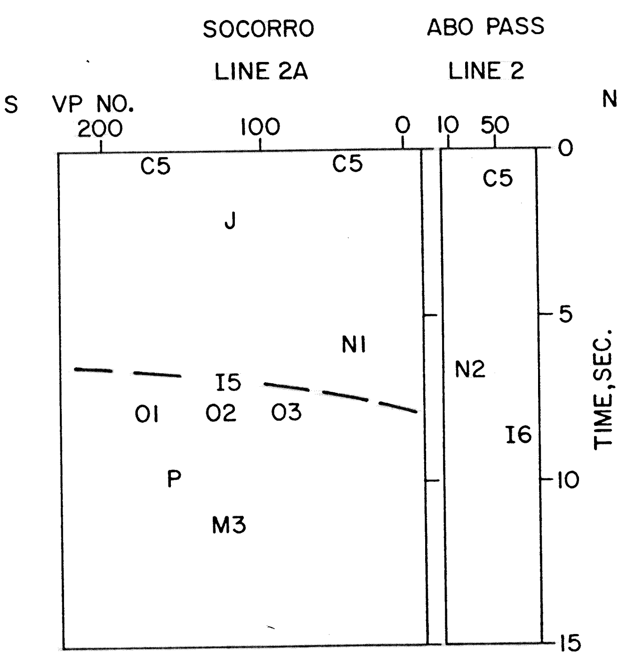

Fig. 3c. Index to features on lines 2 and 2A referred to in text.

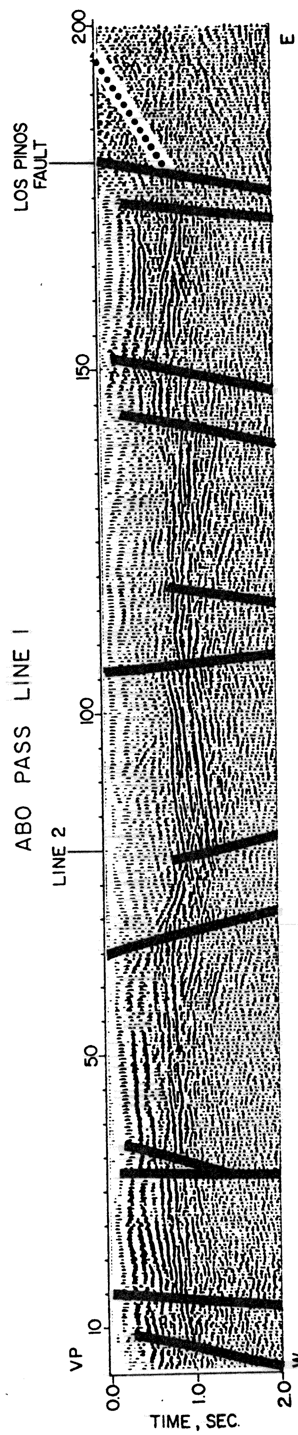

Fig. 4. Details of shallow section along line 1 showing faulted structure of the Hubbell bench. Solid lines denote inferred normal faults. Dotted line is the inferred position of Laramide thrust faults exposed in the Manzano and Los Pinos mountains.

flows, intercalated mafic sills and granitic stocks is upthrown along the Montosa thrust. The Montosa, and the related Paloma, thrusts are most likely Laramide. Both dip about 40-50°to the west at the surface, although Stark [1956] suggests that they may flatten at depth. Five mappable Precambrian units have been recognized in the Los Pinos and Manzano mountains: the basal Sais Quartzite, overlain by the Blue Springs Schist; the White Ridge Quartzite; the Sevilleta Metarhyolite; and the Priest Granite. Except for the Priest Granite these units also dip steeply to the west in the vicinity of the COCORP survey, although to the north in the Manzano Mountains they are folded into a tight syncline, possibly correlative with the concave reflector immediately beneath VP's 150-170 on line 1. Reflections from either the Laramide thrusts or contacts between the Precambrian units, both of which might be expected to be good reflectors, are not prominent on the stacked seismic sections. However, steeply dipping reflections are evident on unstacked records. Their absence on the stacked sections most probably results from their steep dips, which are preferentially discriminated against by stacking, and their shallow depths, where stacking redundancy is much reduced by trace muting. The inferred position of the Montosa thrust [Stark and Dapples, 1946] is indicated in Figure 4. It is possible that the dipping events K in Figure 2 represent the extension of the Montosa thrust beneath the basin, having been offset by the Los Pinos fault. This correlation would contrast with Kelley's [1977] speculation that the Montosa thrust steepens at depth, merging with the Los Pinos normal fault at depth. However, low-angle thrusting is inconsistent with the structural style of the southern Rocky Mountains, and there are alternative interpretations (discussed later) for events K.

Perhaps the most important feature of the eastern rift boundary is displayed on the seismic section immediately below the Los Pinos fault. As illustrated in Figure 2, this part of the record appears as a pronounced, narrow zone devoid of reflections (L), extending from near the surface down to times of at least 8 s. This near-vertical, seismically 'blank' zone dips very steeply to the west. The character of the seismic section on either side of this zone is very different. To the west beneath the rift basin, reflections at depth are generally short, subhorizontal, discontinuous, and for the most part relatively weak. In contrast, reflections beneath the rift flank to the east are stronger, more continuous, and to some degree exhibit more dip. Clearly, this zone is a major feature in the reflection record. However, as is often the case with seismic data, it is necessary to determine if this particular character is due to structural variations at depth, or if it is an artifact resulting from disruptive effects of the near surface regions through which the seismic waves must pass. Caution is especially important in this case since the zone is defined by recordings made while the seismic sources were passing from the basin floor up onto the uplifted flank--- a rapid change in near-surface lithology and structure, not to mention topography. However, the truncation of the seismic reflections east of this zone is well defined on unstacked field records as well as on the processed final sections, indicating that this zone is not an artifact of improper stacking (as might occur, for example, if statics effects were responsible). Although variations in source coupling and near-surface attenuation may have contributed to the lack of observed reflections , the fact that similar zones are not seen on other sections (e.g. line 1A) which also cross major lithologic boundaries at the surface argues against improper processing or near-surface effects being solely responsible. Therefore the most straightforward interpretation of zone L is that it represents a reflector- free zone in the crust.

Assuming that L is a real geologic structure, its correspondance with the subsurface extrapolation of the rift boundary (e.g., the Los Pinos fault) suggests several possible interpretations. The lack of reflections implies compositional and/or structural homogeneity on the scale of the seismic wavelengths used (i.e., a few hundred meters). Such homogeneity could be consistent with L being a fault zone within which intense deformation has destroyed reflector coherence (e.g., cataclasis in the upper, brittle part of the crust; plastic flow in the deeper, hotter regions). If so, this zone could be the deep expression of the central graben boundary fault, and its near-vertical orientation would argue strongly against listric models of graben formation (e.g., Moore, 1960). Alternatively, zone L could represent the deep extension of the Laramide faults exposed in the Los Pinos and Manzano Mountains, implying that these faults must steepen at depth. An important corollary implication of this interpretation is that Tertiary normal faulting (i.e. the Los Pinos fault) developed along the preexisting deep crustal flaws introduced by Laramide thrusting. In a variation on this theme, zone L could represent an even older crustal break along which both Laramide thrusting and Tertiary normal faulting developed. A fourth possibility is that this zone derives its seismic character not from fault-associated deformation but from emplacement of homogeneous intrusive material along a weakness introduced by faulting, perhaps in the manner envisioned by Thompson and Burke [1974].

West

Line 1A (Figure 2) extends westward from the end of line 1 across the Albuquerque Basin to the base of the Sierra Lucero, at the edge of the Colorado Plateau. As pointed out by Kelley [1977], the western border of the Albuquerque Basin, being more irregular and having less structural relief, differs markedly from the uplifted eastern margin. Except in the Sierra Ladron, the western flank is covered by Paleozoic and Mesozoic sedimentary rocks.

In view of the surficial differences in geology between the east and west rift margins, it is perhaps not surprising that the respective seismic sections exhibit pronounced differences. Line 1A intersects the Coyote fault, argued by Kelley [1977] to be the main western boundary fault of the rift, at VP 300. It is evident from Figure 2 that the edge of the Albuquerque Basin is not at the Coyote fault but at the western edge of the Monte Largo embayment near VP 420 (i.e., at the Comanche fault, inferred to be a Laramide thrust). Kelley's postulated Monte Largo bench does not exist. The very shallow west dipping reflectors west of VP 420 (A in Figure 2c) probably represent a thin section of Pennsylvanian limestones, sandstones, and shales on the western (upthrown) side of the Comanche thrust. The stronger, but more gently dipping, event C1 is inferred to represent the floor of the Albuquerque Basin.

An enlarged depth section across the western border is shown in Figure 5. The relatively shallow dip of the basin margin (about 40 degrees, migrated) and the simple character of the reflections from it were unexpected and have been the subject of some controversy. Unfortunately, interpretation of the west margin of the Albuquerque Basin is complicated by Laramide folding and thrusting, deflection of the margin where it crosses a major basement shear zone, and the sparseness of stratigraphic and structural data beneath the alluvial cover of the basin. A unique solution of the problem is not yet possible. The western margin of the Albuquerque Basin varies in structural style from an east facing, gentle downwarp cut by normal faults of small displacement in the northern part of the basin [Black and Hiss, 1974; Kelley, 1977] to an abrupt margin bordering a thick accumulation of basin-fill sediments in the southern part of the basin, as shown by steep gravity gradients and stratigraphic data from oil tests. The change in style occurs at about the latitude of Los Lunas (17 km NNE of Belen) and is conspicuous on the Bouguer gravity map of Cordell et al. [1978; Figure 6].

Additional complications arise because line 1A crosses the western margin within the Monte Largo embayment, which is formed by a southwestward deflection of about 10 km in the basin margin. Kelley [1977, 1979] has interpreted the Monte Largo embayment as a step-faulted bench

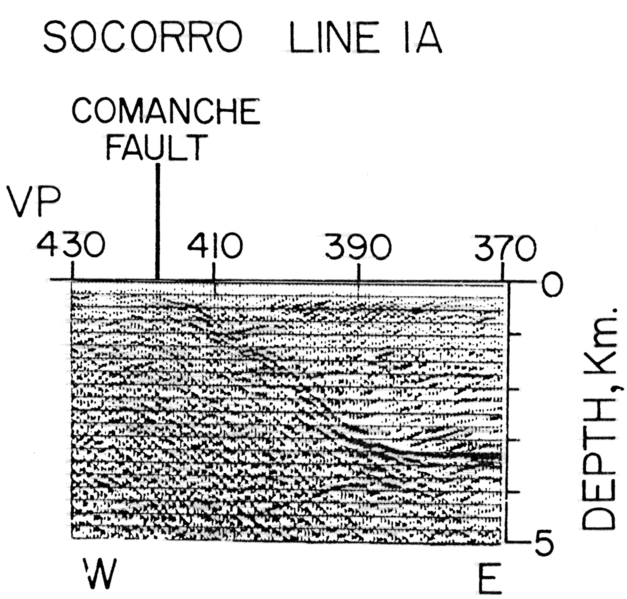

Fig. 5. Details of seismic depth section across the western rift boundary, from line 1A.

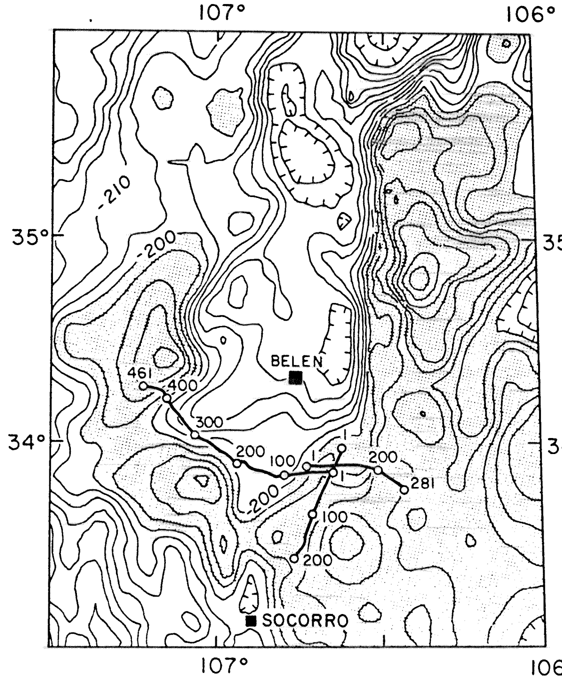

Fig. 6. Location of COCORP lines on Bouguer gravity map of Cordell et al. [1978]. Contour interval is 5 mGal; stippled pattern indicates areas having gravity values higher than 200 mGal.

covered by relatively thin Santa Fe alluvium. However, this interpretation does not fit either the Bouguer gravity map (Figure 6) or the thickness of Santa Fe fill indicated by line 1A (Figure 2), both of which show a deep Monte Largo embayment. The straight northeast trending edge of the embayment crossed by Line 1A is controlled by the Tijeras lineament [Chapin et al., 1979], a major northeast trending shear zone of Precambrian ancestry. A similar deflection of a graben margin occurs about 40 km to the southwest, where the Mulligan Gulch graben crosses a major fault zone of the Tijeras lineament [Chapin et al., 1979]. In many of the transverse zones along the Rio Grande rift the structure of embayments is that of a monoclinal downwarp, or basinward-plunging ramp, similar to that described by Kelley [1979]. Thus the gently dipping reflector forming the west margin of the Albuquerque Basin line 1A (Figure 2) may be downwarped prerift strata, probably the Permian Yeso Formation. Such a downwarp may be partly inherited from the flexure caused by the Laramide Comanche thrust, whose trace coincides with the present margin of the Monte Largo embayment [Callender and Zilinski, 1976].

The Bouguer gravity map of Cordell et al. [1978] indicates that the Monte Largo embayment is a northeast plunging trough (Figure 6). Line 1A crosses this structure at about a 50 angle to its axis and ascends its southeast dipping, northwest flank between VP 390 and VP 410, which is the line segment underlain by the shallow dipping reflector B. The south flank of the gravity trough is formed by the north flank of the Sierra Ladron, a spectacular late rift, west tilted, strongly uplifted horst. Thus the two flanks of the gravity trough contain structures of different ages and structural style. Their appearances on line 1A are also quite different, with the northwest flank being a simple, gently dipping reflector and the south flank showing a complex series of short reflections which are interpreted in the next section and in Figure 14 as representing a step-faulted margin. The very gently dipping floor of the Monte Largo embayment between VP 290 and VP 390 may be essentially a strike line on a strong reflector dipping northeastward down the plunge of the gravity trough.

Other possible interpretations for the structure of the west edge of the Monte Largo embayment are (1) that the gently dipping reflector B (Figure 2) is a low-angle, possibly listric normal fault and (2) that the reflector is a composite return from a series of closely spaced step faults, each of which could be relatively high angle. In the latter case the spacing of the individual faults must be of the order of tens to a few hundred meters if the resulting steps are to be below the resolution of the seismic wavelengths used. If it is assumed that reflector C1 was originally horizontal prior to being downfaulted, its present orientation argues against large-scale, asymmetrical, listric faulting at the western rift boundary. Had the basin floor been asymmetrically downfaulted as much as indicated (approximately 3.3 km at VP 380) along an east dipping fault with substantial upward concavity, C1 should be tilted strongly down to the west. However, as can be seen in Figure 2, this event tilts gently to the east. Slip of this magnitude along a very curved fault would therefore imply some form of compensatory slip along a complementary, west dipping fault to the east in order to rotate C1 back to its present position. Although other seismic sections collected nearby have been reported to show clearly listric faulting [Russell, 1978], there is no direct evidence of such faulting on these sections.

At depth in the vicinity of the western rift boundary the seismic response consists mostly of zones of discontinuous reflector segments (E and F in Figure 2c). In some cases these zones appear to define relatively distinct units which are traceable across the boundary with little indication of major disruption. This apparent zonal 'continuity' is even better displayed on unstacked field records than on the final sections shown in Figure 2. Unlike the eastern rift margin, there is no major throughgoing transparent zone which might correspond to a region of concentrated deformation associated with graben formation or a crustal flaw. Weak, east dipping events in the shallow basement beneath B and C in the vicinity of VP 390 may possibly be diffractions from a steeply dipping fault plane, although some may also be dipping multiples. There is some indication of higher frequency content in the reflections beneath the flank compared with those within the rift, probably because of greater attenuation in the graben fill. The seismic expression of the transition from rift to Colorado Plateau is thus significantly different from that between the rift and the High Plains on the east. Differences in the shallow section, through which deeply penetrating seismic rays must pass, cannot be ruled out as a cause of some of the contrasting appearance; however, it is interesting to note that the thermal regimes at these boundaries may also be quite different. No heat flow measurements have yet been made along lines 1 and 1A, but the distribution of hot springs, travertine deposits, and volcanic rocks, both to the north and south of these lines, indicate higher heat flow along the western edge of the rift. Reiter et al. [1975] have shown from regional heat flow studies that a major geothermal anomaly with values greater than 2.5 HFU (1 HFU = 1 μcal cm^{-2} s^{-1}) coincides with the western part of the Rio Grande rift.

Intrarift Faulting

The most outstanding feature of the shallow section across the rift is the extreme relief on the floor of the Albuquerque Basin, as defined by reflectors C1-C5. Most, if not all, of this relief appears to result from normal faulting accompanying rift formation. Reflector C is a convenient and well-defined marker event which can be traced eastward on line 1A (Figure 2) from the western rift boundary (B) to define a series of structural highs and intervening basins. The most pronounced of these is the large horst H (Figures 2 and 7) separating a major structural depression beneath the Monte Largo embayment from a similar one to the east. Between VP 150 and VP 50 on line 1A the floor of the central graben rises, coming to within 1.0 km of the surface near VP 55. Between VP 55 (line 1A) and the exposed Precambrian of the Los Pinos-Manzano Mountains, the seismic section defines a shallow, faulted ledge (VP 60 to VP 180, line 1) correlative with, and confirming Kelley's [1977] hypothesized Hubbell-Joyita bench. This bench is also conspicuous on the Bouguer gravity map (Figure 6).

In the absence of direct evidence from drilling or outcrop, the identity of the events inferred to represent the floor of the Tertiary graben on lines 1 and 1A (C1-C5) cannot be unequivocally established. The nearest drill holes (Table 1) which penetrate to depths of interest are more than 20 km away, and correlation over that distance in complexly faulted terrain can be a dubious undertaking. Three possible interpretations of C seem worthy of consideration:

1. C is a volcanic unit, possibly correlative with the late Oligocene-early Miocene La Jara Peak Basaltic Andesite [Chapin and Seager, 1975]. The strong amplitude, uniform appearance, and simple seismic character are all consistent with C being an areally extensive extrusive layer. The La Jara Peak Basaltic Andesite consists of thick series of mafic lava flows interbedded both with the upper ash flow sheets of the Oligocene Datil-Mogollon volcanic field and with the basal sediments of the Santa Fe Group. The La Jara Peak is about 360 m thick in the Bear Mountains, 34 km southwest of line 1A, and is nearly as thick in the Lemitar and Joyita Hills, 30 and 15 km south of line 1A, respectively. Also, erosional remnants of mafic flows equivalent in age and stratigraphic position to the La Jara Peak Basaltic Andesite are present along the east flank of the Ladron

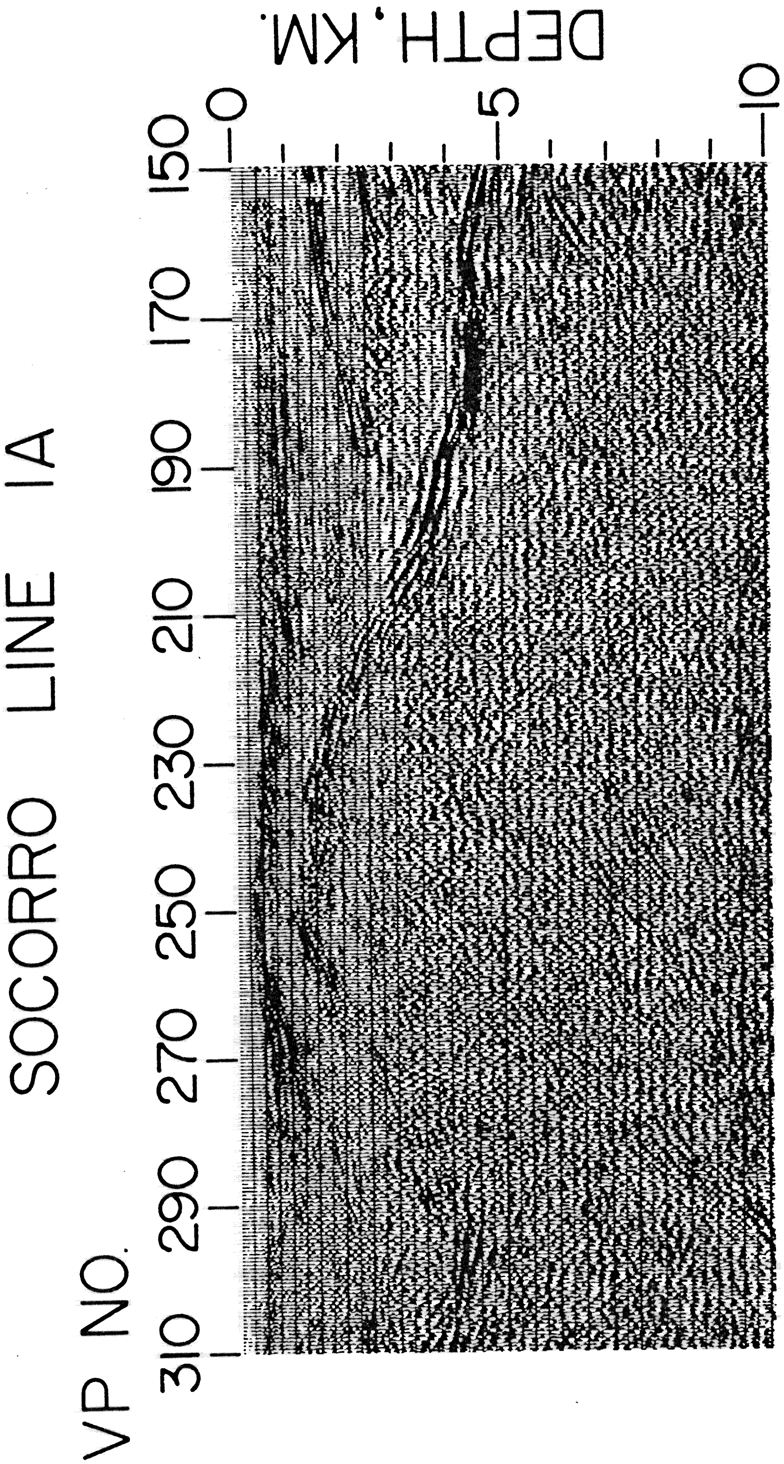

Fig. 7. Details of seismic depth section across an intragraben horst on line 1A.

Mountains (M. N. Machette, written communication, 1979), on Turututu (Black) Butte (basaltic andesite flow capping Oligocene ash flow tuffs and dated at 24.3 + 1.5 m.y. by Bachman and Mehnert [1978]), and along the west flank of the Manzano Mountains about 30 km north of line 1 (mafic flow interbedded with lower Santa Fe alluvium and dated at 21.2 + 0.8 m.y. by Bachman and Mehnert [1978]). Thus the COCORP seismic reflection lines cross an area of the Rio Grande rift known to have accumulated mafic lava flows during the early stages of subsidence. These flows should provide a very strong reflector near the base of the Santa Fe alluvium. This interpretation would imply a substantially thicker post- Oligocene (Sante Fe) unit beneath the Monte Largo embayment than is found in the deep wells to the north (e.g., Humble 1 or Sante Fe 2).

2. C is an Upper Cretaceous horizon, an identification suggested by the deep drill holes which place the top of the Cretaceous section at about the same depth as C (Table 1). However, the relative simplicity of reflector C, with its lack of a well-developed layered character, seems unusual for a Cretaceous section which is about 700 m thick in this area and overlies an additional 2.5 km of Triassic and Paleozoic limestones, sandstones, and shales.

3. C is the top of the Precambrian basement, or a basal Paleozoic layer. This identification correlates C with the base of the shallow layered sequence on line 1 (C5), inferred to be Paleozoic on the basis of refraction velocities and outcrop of the shallow reflectors on line 2A. While this inference is a plausible, in fact probable, correlation for C4, it requires that the post-Paleozoic section above C1 and C3 in the deep basins on line 1A be substantially thinner than that found in the deep drill holes. It also begs the question as to why the reflection character of C1-C3 is so simple, especially when compared with the multiple layering evident in the shallow events (C5) on line 1.

At present, the first two alternatives seems most likely, with hypothesis 1 preferred, at least for C1 and C3. The lack of layered events below C, corresponding to the Mesozoic/Paleozoic section, remains a problem. There is some evidence of weak, layered reflections (G in Figure 2c) near 4 s beneath the Monte Largo embayment which may correspond to these units; however, they may also be multiples of reflector C1 or side reflections from the Sierra Ladron block to the south. Whether C1-C4 all correspond to the same reflector is open to question. For example, C4 is most likely part of the Paleozoic section inferred for nearby line 1, which overlaps line 1A at its eastern end. In any case, C1-C4 appear to accurately depict the scale of vertical displacement induced during rifting and are so interpreted. Depth to C3 at VP 170 is 4.3 km (Figure 8), making the Albuquerque Basin one of the deepest of the rift. Although C1, C3, and C4 are remarkably coherent, many small scale disruptions of this marker are likely due to numerous small-displacement faults.

The midgraben horst H (Figures 2 and 7) rises to about 1.0 km beneath the surface and has more than 3 km of structural relief above the floors of the adjacent basins. Gravity data (Figure 6) and the proximity of H to the nearby Sierra Ladron (Figure 1) suggest that it is the northeast plunging nose of the uplift. The complex reflections defining the sides of the horst exhibit sufficient disruptions to be interpreted as a series of closely spaced step faults. Curved events, interpreted to be diffraction hyperbolae, are especially evident on the top of the horst (Figure 7) and are inferred to be direct evidence of pervasive faulting [Fitch, 1976]. The faults probably vary in trend as Line 1A crosses from the east dipping to north dipping flanks, and some may be arcuate in plan view as shown by surface mapping [Kelley, 1977]. Some of the major longitudinal faults bounding the east and northeast flanks of the Ladron uplift are low-angle normal faults and the strata they cut dip westward into the uplift at angles of 20-60′ . Such strong tilts associated with low-angle normal faults can be interpreted as being caused by either rotation of beds above concave-upward listric faults or progressive rotation of both the fault planes and the strata as the crust is increasingly attenuated during rifting [Morton and Black, 1975]. Chamberlin [1978] has shown that the Morton and Black model adequately explains the complex rift faulting in the area south of the Ladron Mountains and has aptly referred to it as ′domino style′ faulting. The Ladron Mountains are actually the north end, and most strongly uplifted part, of an 80-km-long, north-northwest trending intrarift horst [Chapin and Seager, 1975] that formed during the fragmentation of an early rift basin bordering the Albuquerque Basin on the south. From Socorro north the combination of early rift, domino style faulting and late rift, fault block uplift has tilted Miocene and older strata 30-60°to the west and produced 4-7 km of structural relief along the east side of the horst. Precambrian rocks crop out to elevations nearly 3 km above sea level in the Ladron Mountains and have been downfaulted to elevations about 4.5 km below sea level in the adjacent basins. Thus the subsurface nose of the Ladron uplift crossed by line 1A between VP 190 and VP 290 is a feature with great structural relief and complexity. The seismic section contains little evidence concerning the attitude of the constituent faults at depth. However, the essential horizontality of reflectors C1 and C3 on either side of the horst argues against the master faults for H being listric.

Between VP′s 140 and 55 on line 1A, the central basin shallows from 2.7 s (approximately 4.3 km) to 0.4 s (approximately 0.7 km). Although events C3 and C4 may not be strictly correlative, they clearly indicate a major structural drop-off within the Albuquerque Basin at the edge of the Hubbell-Joyita bench. Again strain seems to have accumulated along a series of westward downstepping faults (Figure 8). Perhaps most striking about the edge of this bench is the great thickness of west dipping sediments which fill the graben out to the central horst (Figure 8; D in Figure 2c). The character of this unit suggests an extremely large alluvial fan complex, probably formed from sediments shed westward off the Hubbell-Joyita bench and Manzano-Los Pinos uplifts. It is likely that there has been

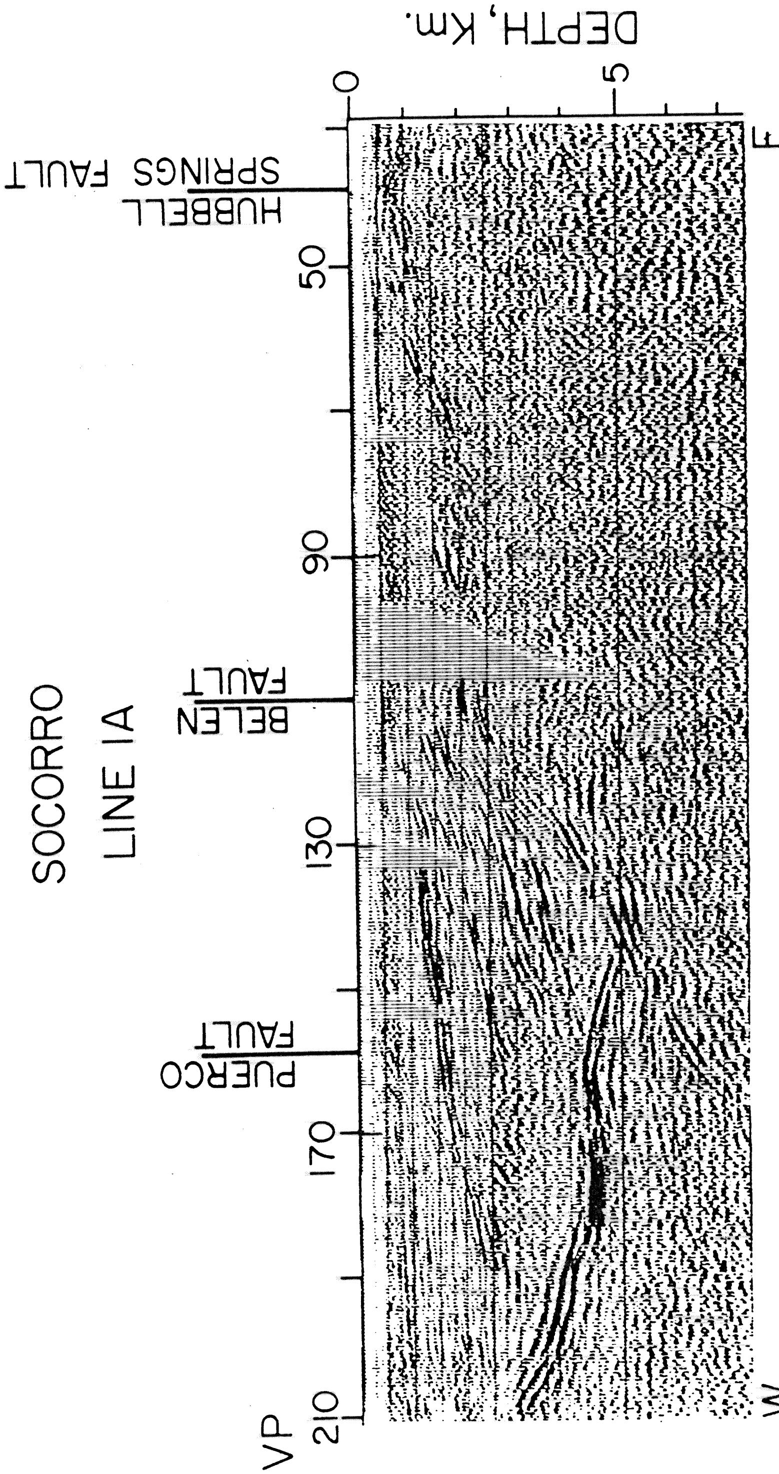

Fig. 8. Details of the western edge of the Hubbell-Joyita bench, showing reflections from a west dipping alluvial fan deposit and/or tilted strata of older basin fill deformed by progressive differential subsidence across the fault zone bounding the bench. Surface position of previously inferred [Kelley, 1977] normal faults shown at top. Depth section from line 1A.

significant rotation of these basin-fill strata by downfaulting at the edge of the bench. This section strongly supports the suggestion by Cordell [1979] that gravity gradients measured across such faults are due both to faulting and lateral density changes associated, here at least, with sorting of clastic sediments in the blanketing alluvial fan [Bull, 1972].

The nature of intrarift faulting near the surface is best displayed on line 1 (Figure 4), which shows the Hubbell-Joyita bench to be pervasively disrupted by normal faults. Although fault plane reflections are difficult to discern, offsets of the sedimentary layers suggest that most of the faults are steeply dipping near the surface. Some of these offsets correspond to mapped faults (i.e., near VP 30--an extension or splay of the West Joyita/Hubbell Springs fault; VP 60--East Joyita fault, and VP 105--unnamed fault shown by Kelley [1977]; see also Sanford et al. [1972]. Most of the blocks defined by faulting show minor tilting, for example, between VP 10 and 50. This tilting is relatively small and could result from either listric [e.g., Moore, 1960] or nonlistric modes of deformation [e.g., Morton and Black, 1975]. The faulting between VP 60 and 80 on line 1 is also evident on line 2 (Figure 3), the two lines intersecting at VP 80 (line 1)/ VP 43 (line 2). The complexity and interfering dips in the shallow section on line 2 suggest sideswipe (i.e., reflections from out of the plane of the profile) and emphasize that these lines undoubtedly cross some structures obliquely.

Further evidence of shallow faulting can be seen in the Tertiary fill along line 1A. A steeply west dipping event which projects to the surface near VP 150 (Figure 8) is probably a fault plane reflection or reflected refraction from the Puerco (or related) fault [Kelley, 1977]. Offsets and large scale slumping in Santa Fe reflectors between VP 190 and 300 may be related to uplift of horst H (Figure 7). Faults in the section beneath the Monte Largo embayment which extrapolate to the surface near VP 210 (west dipping), 300 (east dipping) and 360 (east dipping) may correspond to the Huning, Coyote, and Sante Fe (?) faults, respectively (Kelley [1977]; see Figures 2 and 7). The very shallow dips on some of these may only be apparent, since the profile crosses the surface traces at oblique angles.

The degree to which major normal faults flatten at depth has been a fundamental and disputed aspect of rift tectonics. A review of representative geometries proposed for rift faulting is given by Wright and Troxel [1973]. On the one hand are proposals that the main rift faults extend well into or through the crust at high angle with relatively little decrease in dip [e.g., Thompson, 1966, 1971; Illies, 1970; Stewart, 1971; Thompson and Burke, 1974]. Others cite observed tilting of near surface strata as evidence of substantial rotation of these units along curved normal faults, high angle at the surface but becoming near horizontal along some decollement zone at depth [e.g., Longwell, 1945; Moore, 1960; Mackin, 1960; Anderson, 1971a, b). The listric fault model, cited as particulary appropriate for parts of the southwestern Great Basin [Wright and Troxel, 1973] has been applied by Woodward [1977] to faults in the Rio Grande rift north of Albuquerque, New Mexico. More complex, multistage models have also been proposed, such as the 'domino' mechanism of Morton and Black [1975], which has been recognized in the Lemitar Mountains northwest of Socorro [Chamberlin, 1978]. Evidence on the COCORP sections relevant to this problem is limited. Except perhaps at the eastern rift boundary (where the transparent zone L may mark a throughgoing fault), shallow faults cannot be traced into the basement with confidence. However, there is no reflection that can be identified with the subhorizontal decollement zone of a strongly listric fault. Perhaps the most important piece of evidence bearing on this question is the lack of tilting of the floor of the Albuquerque Basin. Had this basin formed by asymmetric slip along strongly concave master faults, the floor most likely would have be strongly tilted. While far from conclusive, the flat floors argue against listric or domino faulting along deep-seated master faults.

Deep Structure

To a large extent the seismic character of the subsedimentary crust consists of (1) relatively long, ordered, correlatable reflections, (2) zones of numerous commonly subhorizontal and relatively dispersed short reflector segments, and (3) zones totally devoid of identifiable coherent reflections. The first type is exemplified by the reflectors in the sedimentary section and will be referred to as type C (for correlatable or continuous). Examples of such features in the basement include events K1-K4 (Figure 2). The second type, D for discontinuous, is ubiquitous to deep seismic reflection sections [Dohr and Meissner, 1975; Oliver et al., 1976; Oliver and Kaufman, 1977]. Examples include zones E and F (Figure 2c). There is obviously a degree of subjectivity and ambiguity of scale involved in discriminating between these two categories, but the concept of continuity (or its lack thereof) is extremely useful for mapping deep seismic reflection patterns. The third type of seismic character, lack of coherent reflections, suggests either the corresponding lack of reflectors in the earth or the inability of seismic energy to penetrate to reflectors which may exist. Lack of reflectors at a given depth is indicated when reflected energy can be seen to have successfully penetrated to greater depths. Such zones may be classified as type T (for transparent). Zone J is a good example (Figure 3). Without later reflections as a guide it is difficult to establish whether a reflection-free portion of the record is due to a transparent zone or lack of penetration of seismic energy. Small reflection-free zones can also be generated as an artifact of gain processing near very strong events. The gray 'corona' associated with the basin floor events on line 1A is a good example of this effect. Reflection-free zones which cannot be confirmed as transparent will be referred to as type G (for their characteristic 'gray' appearance; not to be confused with the zone labeled G in Figure 2c). The lowest portions of the sections, to which seismic energy has apparently not penetrated (and returned), are examples of type G zones. In addition to these broad classes of events, the seismic records also exhibit numerous hyperbolic and steeply dipping events, often cross-cutting themselves as well as other reflections (e.g., on the east end of line 1). Their curvature and other attributes usually suggest that they are diffractions rather than true reflections.

The classifications and generalizations outlined above pertain only to the character of the seismic record with no explicit petrologic or structural connotations. They are presented primarily as a convenience for discussing the particulars of individual sections, where geologic implications can be more thoroughly explored. Yet in general terms, these types of seismic response suggest a transition from ordered (C zones) to disrupted (D zones) to totally disordered (T zones) states, probably corresponding to a similar variation in the geologic section. The considerable lateral as well as vertical variation on the seismic record of these deep seismic 'facies' is persuasive evidence of the degree and scale of inhomogeneity in the continental lithosphere of this region. It must be remembered, however, that similar characteristics can sometimes be generated as artifacts introduced by improper processing or induced by disruptive effects in the overlying geologic section. The following brief description of the intrabasement portions of the seismic sections precedes an attempt to associate the observed seismic characteristics with plausible, if not specifically verifiable (i.e., by drilling or tracing to outcrop), geologic structural models.

The Precambrian basement immediately below the sedimentary rocks on both the east and west ends of lines 1 and 1A is mostly transparent. Below and to the east of J, separating it from an underlying zone of discontinuous reflectors (i.e. type D), is a distinctive set of relatively strong reflection segments (K1 and K2). The easternmost of these (K2) defines a reflector dipping about 25° to the west. Whether the dipping portion correlates with or truncates the flat-lying segment (K1) is difficult to ascertain. The subparallel events beneath K2 may or may not be multiples. Event K2 is interesting because (1) it is one of the few reflections in the upper part of the basement within the rift which appears to be traceable for any substantial distance, (2) it is one of the relatively few strongly dipping events within the crystalline basement on these sections, and (3) it may correlate with certain neighboring elements of the eastern rift boundary. When K2 is migrated and its displacement due to Tertiary faulting is restored, it stands at the same depth as event K3 beneath the rift flank. Hence it may be the down-faulted and rotated western extension of the layered events K3. Differences in reflection character between K2 and K3 could be attributed to disruption of K2 by faulting. Alternatively, K1 and K2 could be interpreted as a fault plane reflection, possibly a shallow dipping extension of the Paloma and Montosa thrusts exposed in the Los Pinos and Manzano mountains to the east (Figure 4). Recent COCORP profiling in the southern Appalachians has emphasized that such thrusting involving crystalline basement can be an important mode of orogenic deformation [Cook et al., 1979b]. However, evidence of 'sled-runner' style Laramide thrusting, while abundant in the northern Rockies and Arizona [e.g., Bally et al., 1966; Drewes, 1978], is generally lacking in the Southern Rocky Mountains of Colorado and New Mexico. A third hypothesis is that these events represent the lower portion of the Los Pinos normal fault, implying that the rift boundary at Abo Pass is extremely listric; however, the geometry required by this interpretation seems highly unrealistic and inconsistent with the faulting observed in the overlying basin sediments.

The basement beneath the Hubbell-Joyita bench includes discontinuous reflector zones, transparent zones, an extremely strong reflection (I4) at 7 s on the west end of the line 1, and a relatively weak but nonetheless distinctive reflection (M2) at 11 s. The last two, which also appear as unique features on the other record sections, are inferred to correspond to a magma chamber and the Moho, respectively, and are treated later. The discontinous reflector zones in some cases seem to define elongate bands, while in other cases they occur as irregularly shaped patches of more or less uniform appearance. In contrast to the apparently discontinuous nature of the basement reflections west of the transparent zone L, the seismic section beneath the eastern flank is dominated by stronger and generally more continuous reflectors, perhaps indicating less disruption beneath the flank than beneath the central graben. The section is complicated by at least 3 cross-cutting sets of coherent reflected energy: (1) an east dipping ensemble with relatively weak amplitudes, (2) a much stronger west dipping ensemble, and (3) an equally strong set of nearly horizontal reflections. This pattern extends from 4 s down to over 10 s (Figure 2). At least two of these sets are likely to be diffractions from inhomogeneities located either out of the plane of the section or off the ends of the line. The best intrabasement reflectors beneath the eastern flank are those at K3, which show an apparent onlap more than 3 km beneath the top of the Precambrian basement. Another type C event at about 4 s (K4) seems to bifurcate to the west. Below K4 and above M1 is a thick sequence of reflections beginning at about 6 s. The last clearly identifiable event on the section is M1 at 12 s, inferred to be a Moho reflection and discussed in a later section.

The character of the deep crust within the rift is best displayed by the seismic section obtained along line 2A (Figure 3). Because line 2A is oriented along rather than across the structural grain of the rift and crosses a thinner attenuating sedimentary section, the disruptive effects of faulting and basement topography are much less pronounced. Line 2A indicates a thin sedimentary section (C5) overlying a pronounced transparent zone in the upper basement (J). The shallow reflectors can be traced to an outcrop near VP 120, where Paleozoic rocks are exposed [Spiegel, 1955]. Transparent region J extends from about 1 to 4 s across the entire section, representing a seismically uniform body at least 30 km by 12 km in cross sectional area. Its extent perpendicular to line

2 is difficult to ascertain, although a possibly correlative zone can be seen on line 1 (J in Figure 2). Below J the section is again characterized by numerous discontinuous zones of short reflector segments, better defined and more numerous here than on lines 1 and 1A, probably because of the greater simplicity of the overlying section. On the north end of the line, several events exhibit a uniform northward dip (e.g., N). Some of the reflector groups between 4 and 7 s are quite strong. As on lines 1 and 1A, the most pronounced deep reflections come from a zone at about 7 s (I5, discussed in detail later). Below the 7 s event are numerous other reflections. Some, like O, appear to be subhorizontal and discordant with the overlying north dipping events, while others also dip to the north. Some regions, such as P, appear to be true transparent zones since they are floored by relatively strong reflections. Between about 11 and 12 s on line 2A is another pronounced band of discontinous and complex reflections (M3). Correlation of this ensemble with the standard crustal model derived from refraction in this area (Figure 3) suggests that these are reflections from the Moho.

Although the identity of certain events on deep seismic records (e.g., the magma body I or the Moho M, discussed in following sections) can sometimes be inferred from relevant geological and geophysical information, deep reflections which cannot be traced to the surface still remain the subject of considerable speculation. Yet useful generalizations are possible. Transparent zones, for example, imply homogeneity and have often been interpreted as igneous plutons [e.g., Oliver et al., 1976; Smithson et al., 1977]. This interpretation is viable for these data as well--event J on line 2A could easily be a large granitic pluton, perhaps correlative with the Priest and Los Pinos granites exposed in the nearby Manzano and Los Pinos mountains [Stark, 1956]. At the west end of line 1A, where the shallow basement also appears transparent, a granitic pluton at the base of the overlying sedimentary strata is well exposed in the nearby Ladron Mountains. Large volumes of granite in the upper crust would also help explain the relatively low upper crustal velocities observed by Toppozada and Sanford [1976]. However, igneous plutons are not the only possible explanations for a transparent seismic response. For example, zone L at the eastern rift boundary could be due to intense structural disruption along a wide fault zone. Likewise, relicts of shear zones from preceding orogenies could be scattered throughout the crust.

The range of structures that could reasonably produce a discontinuous seismic response is very broad, and many possible geologic analogues have been discussed in the literature [Smithson and Brown, 1977; Smithson et al., 1977; Smithson, 1979], including the following: (1) folding with wavelengths of the order of, or smaller than, the depth of burial, resulting in quasi-periodic focusing of reflected energy or isoclinal folding of a pervasive but intermittent nature (Figure 9), (2) lamellar or podiform structures, such as intermittent gneissic banding, boudinage, metamorphic mineral segregations, etc., (3) pervasive faulting, (4) multiple, small scale intrusions, (5) lateral variations in layer thicknesses, causing alternating constructive and destructive interference of reflections, (6) near-surface 'statics' effects, (7) shallow generators of complex path multiples, (8) sideswipe from shallow complexities. The last three possibilities represent coherent 'noise' and contain no information about the deep crust in the plane of the section corresponding to the times at which they are recorded. They are listed here as a reminder that care is required in translating a deep seismic time section into structure. While processing and recording techniques are designed to minimize or eliminate such contributions, much remains to be learned about the extent of their possible interference.

The first five structural types are common features exposed in the metamorphic (granulite) terrains often cited as analogues for the deep crust (e.g., the Ivrea zone of the Alps [Mehnert, 1975]. Xenoliths found in New Mexico diatremes suggest similar rock types and, by association, structures in the lower crust beneath the rift; for example, Padovani and Carter (1977) report anhydrous quartzo-feldspathic garnet granulites, two pyroxene granulites, charnockite, and anorthosites from Kilbourne Hole (see also Warren et al. [1979]). It is reasonable therefore to postulate that some, if not the majority, of the discontinuous reflector patterns defined on these sections correspond to structural variations in moderate- to high-grade metamorphic rocks. If such a generalization is valid, the volumetric proportion of discontinuous (metamorphic) to transparent (igneous) appearing zones on the seismic sections may be a crude guide to the relative importance of these two processes in crustal evolution; here the former dominates. The first and fifth models above are particularly attractive as explanations for deep reflections because they involve interference effects (such as focusing and tuning) which serve to enhance net reflectivity, an important mechanism considering the relatively moderate velocity contrasts which are probably commonplace in the lower crust [Smithson and Brown, 1977; Noponen et al., 1979].

The apparent subhorizontal attitude of the majority of the short reflector segments on these (and other) sections seems inconsistent with hypotheses for a highly deformed and intruded metamorphic crustal model, where steep dips due to folding would be expected to be commonplace [Smithson, 1978]. Neither is it clear that the biases against dipping events inherent in the data acquisition and processing techniques are an adequate explanation for this preferred orientation. It is possible, though speculative, that there is actually a subhorizontal 'fabric' in the lower crust. Subhorizontal structural styles have been reported for certain Archean terrains (Figure 9 and Bridgwater et al., 1974), and may represent a fundamental mode of structural alignment in the lower, and presumably more plastic, crust. Shear deformations along sub-horizontal planes or creep under the influence of gravity have been suggested as mechanisms for generating such a fabric [Phinney, 1978; Berry and Mair, 1977; Smithson, 1978]. In this case, perhaps extensional stresses due to

Fig. 9. Isoclinal folding in the Precambrian of Greenland, a possible analogue for subhorizontal, discontinuous reflectors in the lower crust beneath the Rio Grande rift ? [Escher and Pulvercroft, 1976].

rifting have served to impress a preferred orientation, although the ubiquitous nature of these zones argues for a more general mechanism, or at least a multiplicity of possible causes. Until more specialized processing and analysis is carried out, the significance, if any, of this preferred reflector attitude will remain uncertain.

The Magma Body

The most distinctive set of deep reflections observed on COCORP profiles within the rift is a relatively coherent, high amplitude event I between 6 and 8 s, (18-24 km) on lines 1, 1A, and 2A. This complex compressional wave reflector corresponds in depth to a strong shear wave reflector inferred to represent the top of a magma body beneath the rift in this area [Sanford et al., 1973, 1977]. Releveling results indicate that it is still active [Reilinger and Oliver, 1976]. The existence of magma bodies in the crust is consistent with the extensional tectonic setting, late Cenozoic volcanism [Chapin and Seager, 1975], and a variety of geophysical observations, including the regionally high heat flow [Reiter et al., 1979], and swarmlike seismicity [Sanford et al., 1979]. P wave delay studies suggest that the magma body is very thin in relation to its areal extent--less than 1 km thick assuming a 100% partial melt [Rinehart et al., 1979]. While several geomagnetic studies have suggested molten material at relatively shallow depths beneath parts of the rift [e.g., Hermance and Pedersen, 1978], magnetotelluric soundings by Jiracek et al. [1979] in the vicinity of the COCORP profiles indicate a high conductivity layer at a much shallower depth (about 10 km) than the reflector discussed here. Because of possible effects of shallow lateral conductivity variations, the implications of these measurements are not yet clear.

The magma body event is most clearly seen on line 2A (I5 in Figure 3c). The reflector appears as a complex zone of events as much as 0.25 s wide, distinguished primarily by its coherence across the section and high amplitude relative to nearby events. Figure 10 shows the average and maximum signal amplitude for the stacked traces on line 2A. The magma body event is the pronounced high amplitude return at 7 s, averaging 6 db above background level. The arrival time of about 7 s (midsection) matches the estimated 20 km depth of the S wave reflector nicely. This event cannot be a multiple because there is no appropriate primary, and it cannot be a side reflection because it is simultaneously recorded on lines oriented at high angle to each other (1A and 2A). It is possible that the P wave reflections are coming from an effective reflector associated with, but distinct from, the S wave reflector, although the close spatial correlation renders such an intepretation ad hoc at best.

The COCORP surveys delineate a laterally varying reflection character, commonly consisting of several cycles in the reflected waveform. In some places the event consists of cross-cutting reflector segments, while in others it is defined by a series of parallel segments. Migration of the seismic section for line 2A eliminates some but not all of this complexity. Although side reflections from the same event (if it has substantial topography out of the plane of the section) or pegleg multiples generated in the overlying sedimentary rocks may also contribute to the character of the reflection, the simplest and preferred interpretation is that the reflector itself has a complex structure, possibly consisting of closely spaced layers varying in number and thickness.

Figure 11 shows the reflection character of the magma body along line 2A in more detail. The section has been passed through a 20 to 30 Hz high-pass filter to enhance contrast. The upper surface of the event is irregular, with an

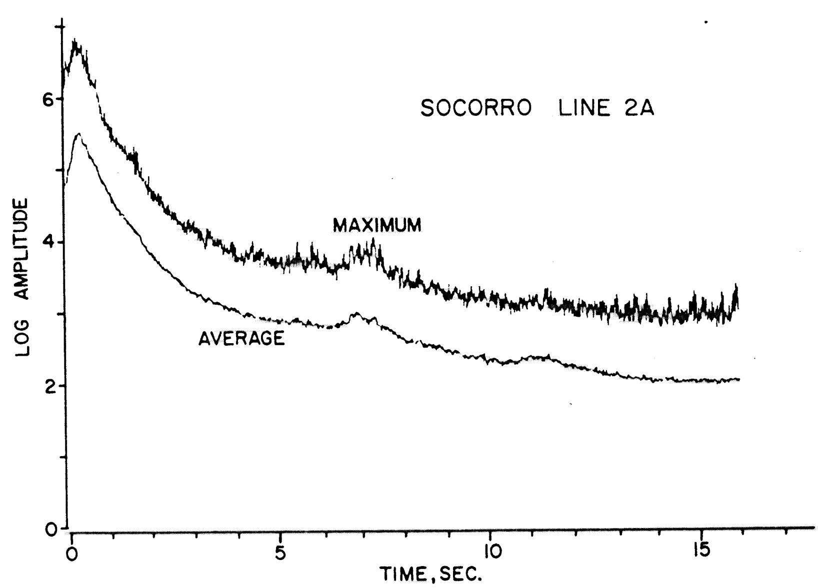

Fig. 10. Average and maximum amplitude recorded as a function of travel time for all traces along line 2A. The amplitude decay curve exhibits local maxima corresponding to energy reflected from a midcrustal magma body and the Moho.

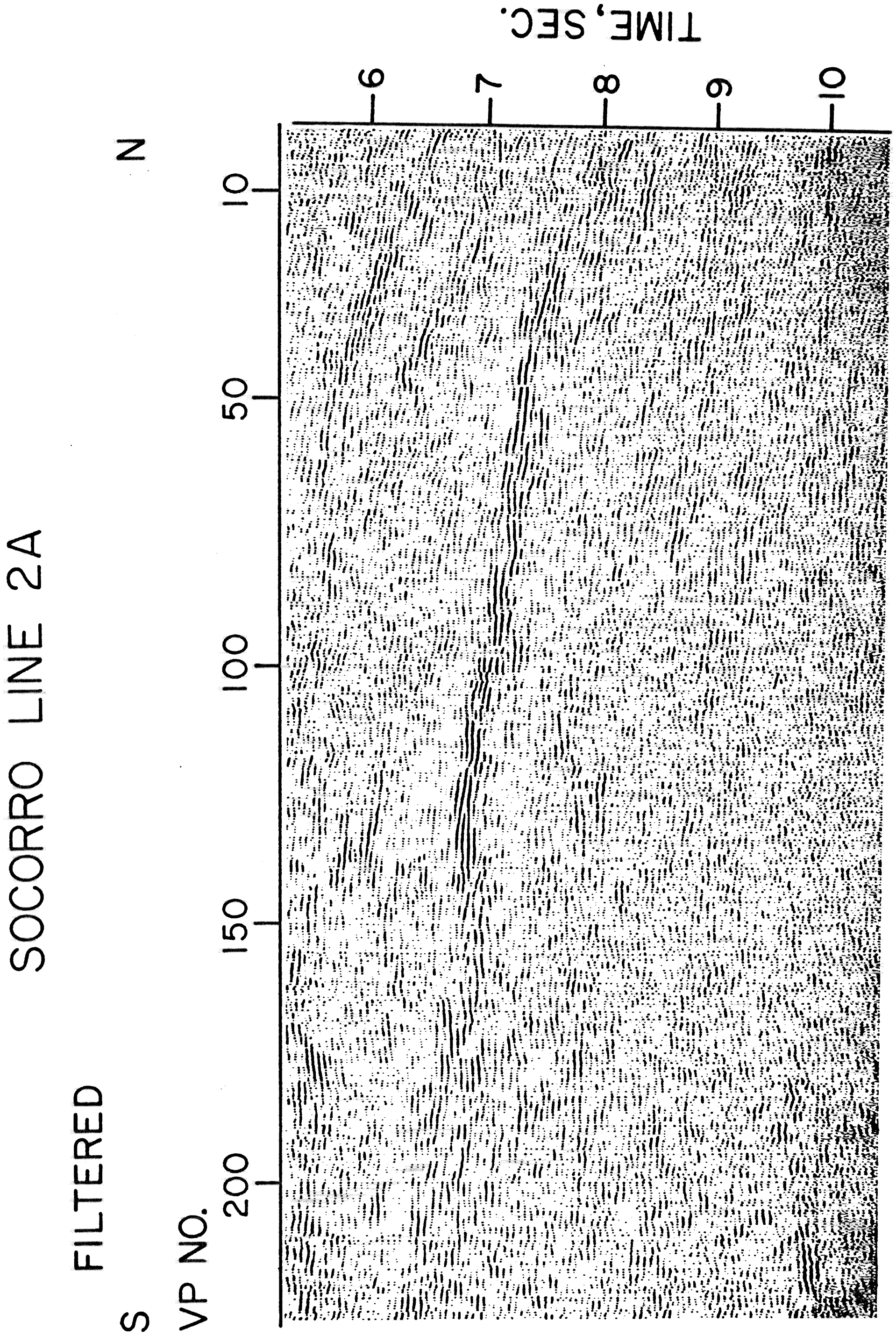

Fig. 11. Details of the midcrustal magma body from line 2A. Filtered with 20 to 30 Hz bandpass filter (input sweep is 10-32 Hz).

undulating appearance; the event is strongest and best defined beneath VP 50 and 130. To the south the ensemble dip appears to decrease, and the event diffuses into a broader, less well defined zone. At the north end of 2A the event also appears to lose its identify with respect to surrounding reflector segments.

The frequency, amplitude, and waveform characteristics of a portion of the magma body event are shown in Figure 12, where the constitutent recordings of the stacked trace at VP 132.5 on line 2A (i.e., those traces which correspond to seismic rays which would reflect from the same subsurface point in a homogeneous earth), are displayed at a fixed but arbitrary gain. This gather has been corrected for normal moveout at a velocity of 6.1 km/s and filtered at successively higher frequency passbands. Frequency spectra for 0.5 s windows averaged over 24 adjacent stacked traces, including VP 132.5, are also shown. From these unstacked, filtered displays it can be seen that (1) the effective reflector is sufficiently sharp to return substantial high frequency energy (25-30 Hz, corresponding to wavelengths of 240-200 m), (2) the first and strongest component in the reflection waveform (see, for example, the 15 to 20 Hz panel) appears to have reversed polarity, consistent with but not unique to a simple change from high to low velocity, (3) other high-amplitude reflections are also apparent (e.g., near 7.8 s at the low frequencies), although none that are as consistently strong as the 6.8 s event, and (4) reflections from segments beneath the 6.8 s event appear in at least some cases to have a lower frequency content than those above it (see spectra at left in Figure 12).

Although these characteristics are consistent with the existence of partially molten rock in the middle crust, alternative models are possible [Brocher, 1979; Krumhansl, 1980]. For example, the apparent polarity reversal and

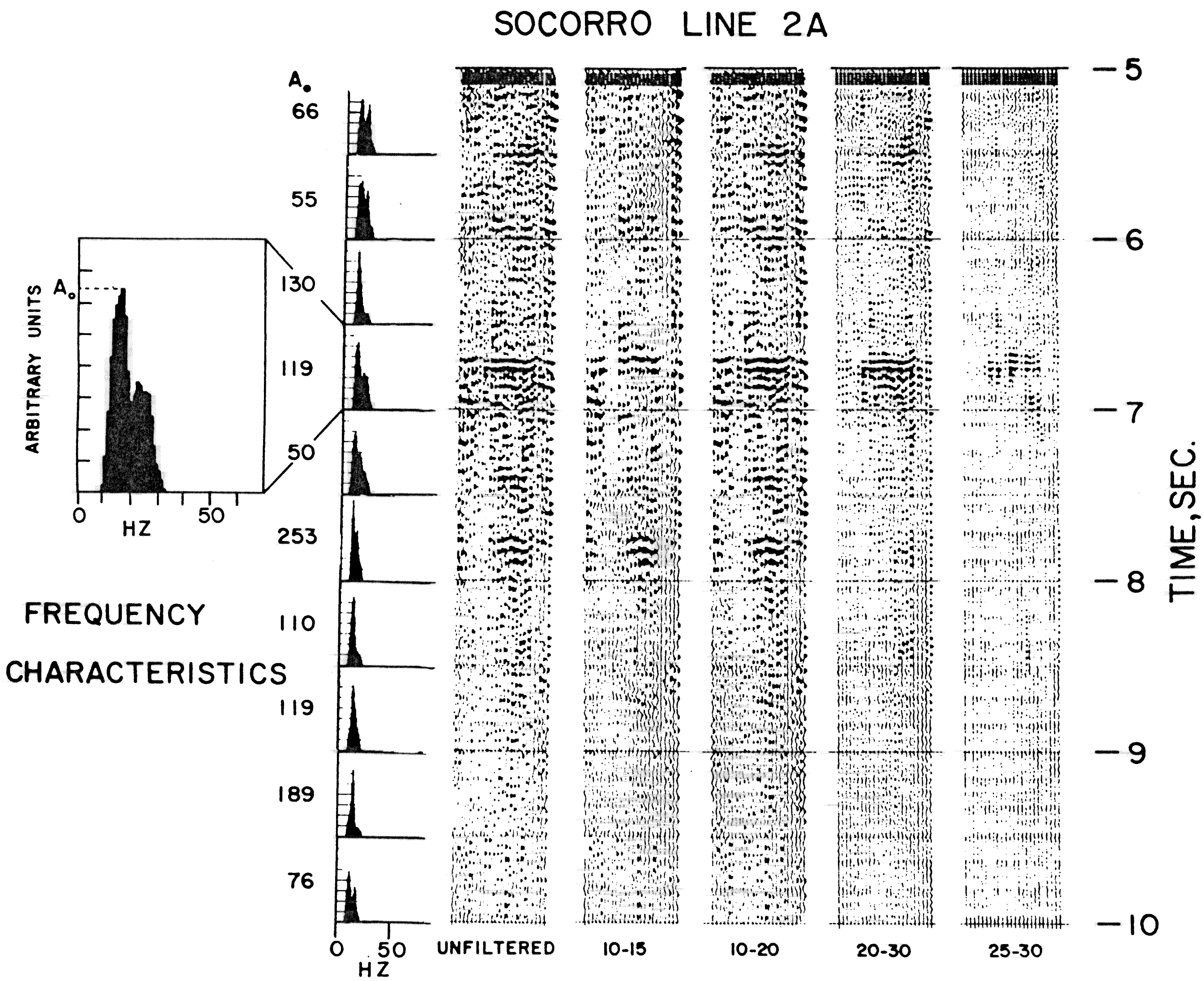

Fig. 12. Frequency characteristics of the magma body reflector. Fixed gain display of individual traces of one depth point near the center of line 2A. The gather panels have been filtered with bandpass filters using corner frequencies indicated at bottom. Spectral estimates at left based on 0.5 s data windows down trace, averaged over 24 consecutive stacked traces. Details of spectrum for window embracing the magma body reflection at far left.

relatively high amplitudes might also be obtained by moderate contrast layering with fortuitous spacing. Furthermore, other strong events, such as those at 7.4 and 7.8 s, suggest that the conditions giving rise to the high-amplitude reflections at 6.8 s are not necessarily unique to that particular event, although these other reflections do not appear to be as coherent over long distances.

If the complex 7 s event in Figures 2 and 3 corresponds, like the shear wave reflector, to the top of a magma body, an equally pronounced reflection from its bottom is not obvious. Either the base of the magma chamber is not a strong, consistent reflector (e.g., a crystal 'mush' zone), or the effective thickness of the body is small, and the observed reflection is actually a composite waveform made up of reflections from the top and bottom (and intermediate?) parts of the chamber. The latter interpretation is consistent with P delay results [Rinehart et al., 1979] and spectral analyses [Brocher, 1979] which indicate a thin body. Distinct, though weaker and less coherent, reflection segments are evident beneath the 7 s event on line 2A (such as O in Figure 3c), which could be interpreted as pieces of a 'base' for the magma body. Yet, the complexity of the waveform strongly suggests the interference effect of closely spaced reflectors, such as might be expected for thin intercalations of molten material within solid rock. Lateral variations in the number and thickness of interspersed magma 'lenses' could account for the variability in seismic character.

The extent of the magma body event on COCORP sections appears to be large but limited. Reflector I4 on line 1 seems to end, at least as a distinctive high-amplitude event, near VP 50, although a much lower amplitude band of reflections can be discerned on unstacked records as far east as the transparent zone L. East of L, there is no single distinctive reflection which can be correlated with I, although a broader band of strong reflections is evident at 6 s on unstacked records. Olsen et al. [1979] report strong compressional wave arrivals on wide-angle recordings east of the rift boundary, but extrapolating the 7 s event across the rift edge at Abo Pass on the basis of the COCORP data alone is not justified. The 7 s event appears to become a diffuse set of reflections and lose its identity at both the north and south ends of line 2A. A distinctive reflection corresponding to this event is difficult to pick on line 2, although the strong event I6 lies along the northward extrapolation of I5 on line 2A.

The apparent western termination of the magma body event is particularly interesting. Rather than simply dying out in amplitude, it appears to splay into three fairly distinct branches. The disruptive effects of overlying basement topography appear inadequate to account for this splaying. Rinehart et al. [1979] point out that if the velocity pulldown effect of the graben fill is taken into account, the faulting responsible for uplifting the horst H in relation to the basin floor seems to have had no effect on the underlying, relatively flat 7 s event. This suggests that either the magma body was emplaced after the formation of the horst or the faulting associated with the horst was accommodated (by plastic flow?) above 20 km. Both continuous and discontinuous reflector zones exist at similar depths west of VP 250, but they do not appear to be sufficiently pronounced or distinctive to be identified with the magma body.

Superficially, the 7 s reflections suggest something resembling the top of the magma chamber in the model of Wyllie [1970] or the clay models of Ramberg [1970] (Figure 13). Although a connected sill model is plausible, a perhaps more realistic interpretation of these new data is that the reflections are from interfingered magma lenses. If such lenses or pockets are sufficiently closely spaced, they would appear as essentially a single reflector to the sparse (relative to multichannel coverage) shear wave observations. If the thin body, or multiple lens, interpretation is correct, the question arises as to why this material accumulated where it did instead of continuing on to the surface. The relative concordance between the 7 s event and shallower, north dipping reflections (e.g., N in Figure 3c, see also Figure 11) suggests that magma may have accumulated along some prexisting barrier to migration [Rinehart et al., 1979]. Only part of this northward dip can be attributed to velocity pulldown of the northward thickening sedimentary section. The barrier may correspond to the refractor found by Toppozada and Sanford [1976] near this depth (perhaps corresponding to the Conrad, although Oliver [1978] has questioned the significance of this discontinuity), and there is some evidence on the seismic sections for an increased density of reflection segments in the 6 to 8 s range beyond the limits of the magma body event (e.g., F in Figure 2c). These admittedly tenous correlations indicate a midcrustal discontinuity, possibly of local extent, serving as a barrier to upward migration of melt, a kind of 'cap rock' for magma accumulation [Elder, 1978]. A similar 'barrier' model has been proposed for magma chambers beneath the mid-Atlantic ridge by Nisbet and Fowler [1978]. If geothermal gradients are sufficiently high, and heat flow measurements indicate they may be [Reiter et al., 1979], the lower crust could be sufficiently plastic to allow magma to migrate upward toward this postulated barrier as droplets [Turcotte and Ahern, 1978]. Alternatively, magma may have intruded along localized weaknesses or propagating cracks [e.g., Anderson and Grew, 1977]. It is unlikely that it migrated upward for any great distance as a very large body [Ahern et al., 1979].

The geotherm estimates of Decker and Smithson [1975] for the southern Rio Grande rift (heat flow = 2.4 HFU) intersect the wet granite solidus at about 20 km. If these geotherms are applicable to the Socorro area, it is possible that a wet granitic fraction at midcrustal depths would have developed into a zone of partial melt. Lateral changes in reflection character could result from variations in the local amount and distribution of melt, which is in turn dependent upon water content and bulk composition. However, in situ melting is an unlikely explanation for the inferred magma chamber reflector. The anhydrous nature of lower crustal xenoliths in New Mexico [Padovani and Carter, 1977], the composition of synrift volcanic rocks [Chapin,

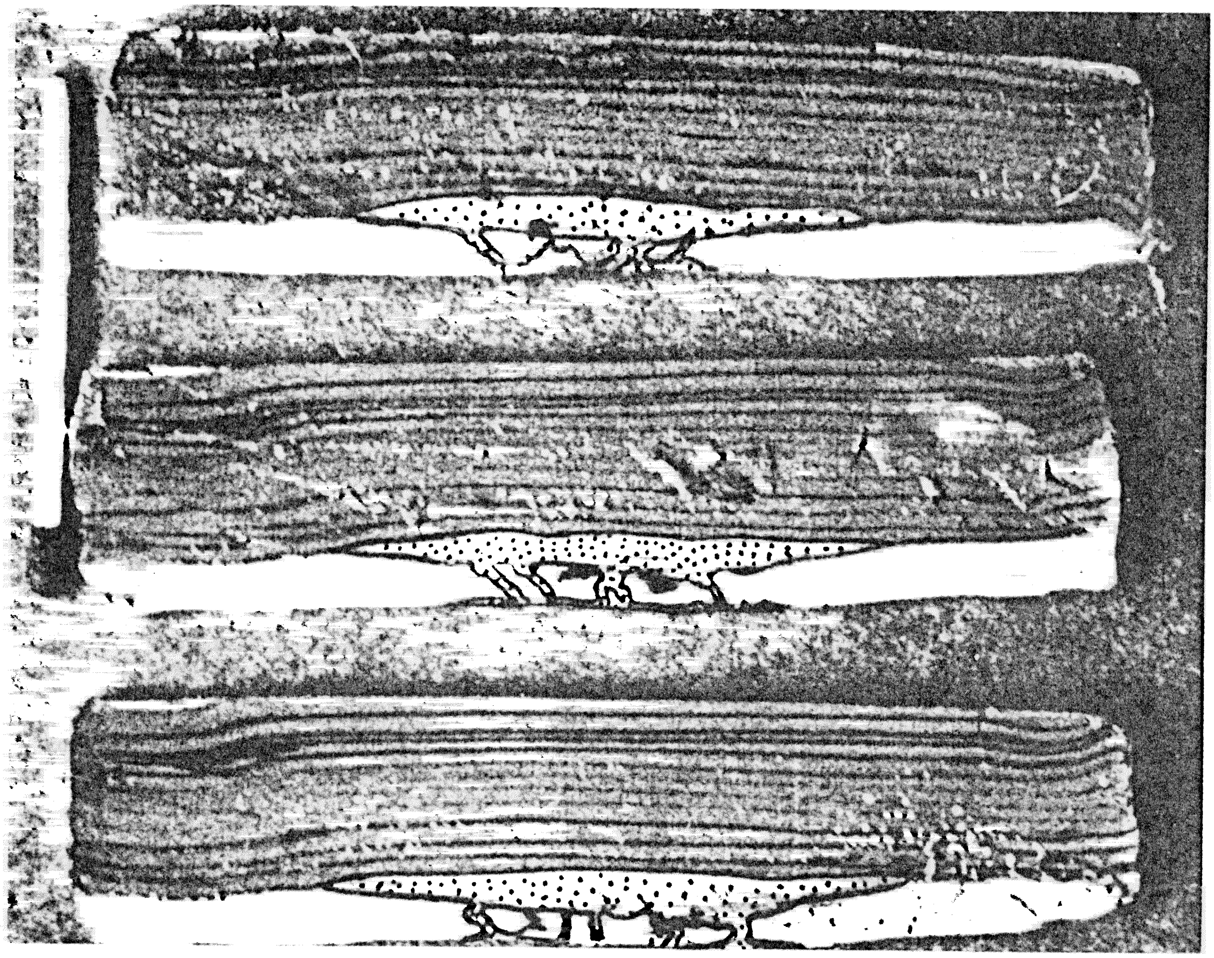

Fig. 13. Laboratory clay model of a sill, a possible analogue for the Socorro magma body. [Ramberg, 1970].

1979], and the isotopic characteristics of surface volcanics [e.g., Zimmerman and Kudo, 1979] argue against widespread partial melting at midcrustal depths. Furthermore, partial melting is more likely to form a diffuse transition zone than a sharp reflector such as that observed.

The Moho

The amplitude decay curve in Figure 10 shows a local peak near 11 s on line 2A, corresponding to a series of laminated reflectors between 10 and 12 s on the seismic section (M3). These events at 30-36 km are interpreted to be Moho reflections in accordance with nearby refraction surveys [Toppozada and Sanford, 1976; Olsen et al., 1979]. This transition from crust to mantle appears on the seismic section as a complex zone of reflector segments varying in number and amplitude, in contrast to the relatively simple discontinuity usually associated with more classical refraction models. Because reflections were returned at 10-15 Hz (Figure 12), the seismic discontinuities involved must be defined on the scale of a few hundred meters or less. The laminated character of this zone is similar to that observed in other areas [Meissner, 1973; Clowes et al., 1968] and has been interpreted in terms of interfingered zones of partial melt, preexisting gneissic banding [Meissner, 1973] or subhorizontal tectonism [Berry and Mair, 1977]. In view of the high heat flow and recent volcanism in the rift, the first hypothesis would seem an attractive candidate. The relatively strong reflection character of the Moho on line 2A (Figure 10) is quite possibly due, at least in part, to constructive interference from layering. Events M1 and M2 on line 1 (Figure 2c) are also likely to be reflections from the base of the crust, although they are distinctly not laminated. The eastward dip of M1 and M2 is consistent with the expected eastward increase in crustal thickness. There are some weak events in the 11 to 13 s range on line 1A, but a coherent zone cannot be traced. The lack of more pronounced Moho reflections on lines 1 and 1A may be due to the more severe statics problems caused by near-surface structure on these profiles. The observed complexity of the crust-mantle transition argues against the subrift Moho being a simple phase change or a chemical discontinuity but is consistent with the concepts of lower crustal heterogeneity summarized above.

Summary and Conclusions

The Rio Grande rift which emerges from examination of these seismic sections is an asymmetric, composite graben formed predominantly by normal faulting within a complex, heterogeneous crust. Figure 14 is an interpretive, and very simplified, east-west cross section summarizing its salient characteristics as deduced from COCORP seismic data. Although based primarily on lines 1 and 1A, relevant features on the cross lines 2 and 2A, for example, the strong Moho reflections, are projected or extrapolated onto this view. Among the observations and speculations emphasized in this composite are the following:

1. The shallow basement beneath the Albuquerque Basin is pervasively disrupted by Tertiary normal faulting.

2. The rift at the latitude of Abo Pass is an asymmetric graben; the existence of an extensive shallow bench (Hubbell-Joyita) beneath the southeastern portion of the Albuquerque basin, and the absence of a similar feature in the southwest portion, is confirmed. The western edge of this bench is bounded by step faults and overlain by a large, west dipping alluvial fan, probably tectonically rotated. Although not evident on the seismic sections, substantial thicknesses (perhaps several kilometers) of Mesozoic and Paleozoic sedimentary rocks may underlie the deep basins.

3. A large, buried intragraben horst projects northward beneath the Albuquerue Basin from the nearby Sierra Ladron. Faulting associated with the flanks of this horst does not appear to have offset deeper reflectors at 20 km. Normal faults bounding this horst are shown here as steeply dipping step faults, but other geometries are also possible.

4. The eastern rift boundary at Abo Pass appears to be defined by a seismically transparent, high-angle zone which extends through the entire crust. This zone may represent a plane of intense deformation and/or intrusion associated with rifting, earlier Laramide thrusting, or an even earlier crustal flaw.

5. The western rift boundary at the Sierra Lucero is represented by an anomalously low angle event which may represent a structural downwarp, or ramp, of prerift strata or, alternatively, a composite reflection from a series of closely spaced step faults.

6. A strong, complex reflector at about 20 km depth corresponds with an inferred magma body beneath the rift. This reflection zone is relatively flat and may represent the accumulation of pockets of molten material at the base of a midcrustal discontinuity.

7. The crust-mantle transition (Moho) beneath the rift is a laterally varying zone of discontinuous, sometimes layered, reflector units.

8. Numerous seismically transparent zones on the section may indicate homogeneous plutons emplaced in the crust. The best defined of these, beneath the Hubbell-Joyita bench, is about 9 km thick and may correlate with nearby granitic intrusions exposed in the flanking Precambrian uplifts.

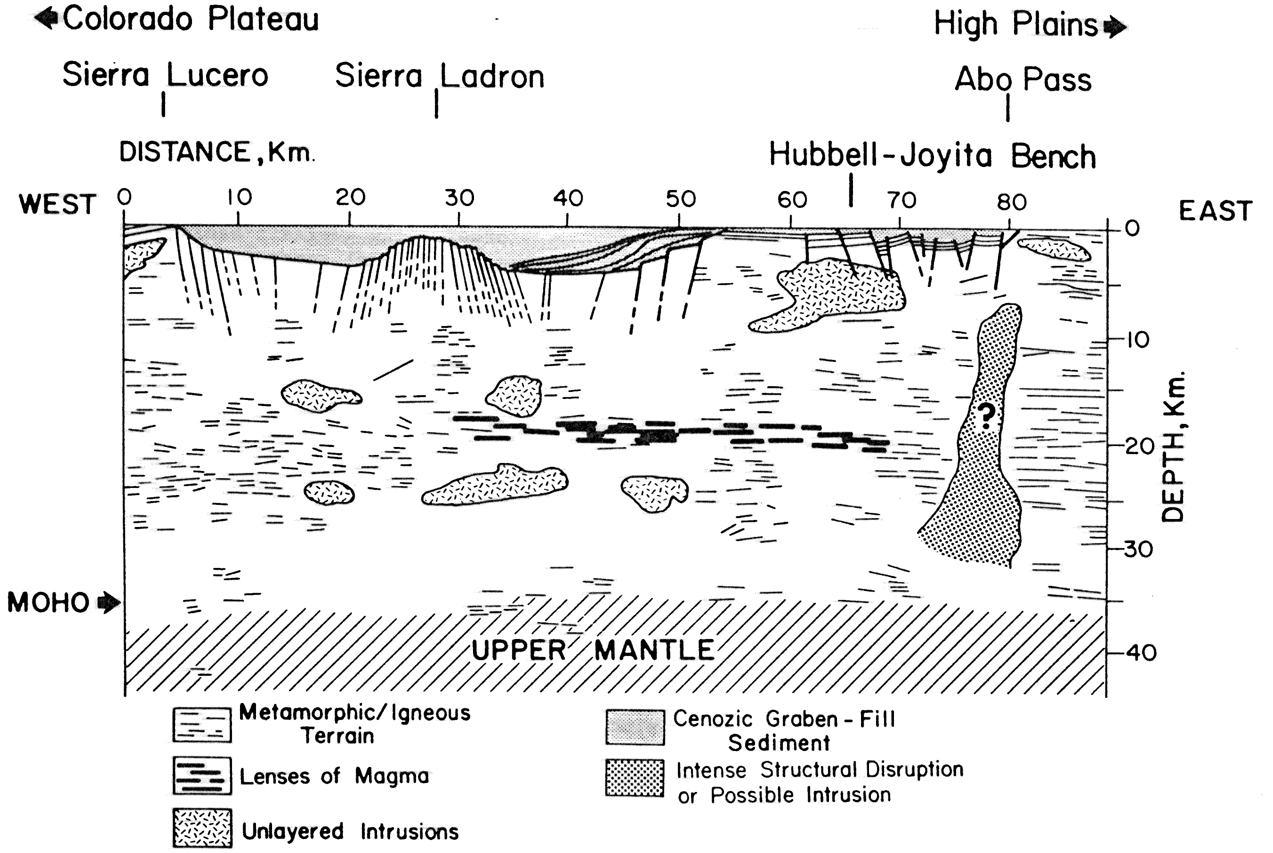

Fig. 14. Schematic cross-section summarizing salient features observed on COCORP sections across the Rio Grande rift. Based primarily on lines 1 and 1A, but with important features of lines 2 and 2A projected or extrapolated as necessary (see text).

9. The majority of the crustal section is characterized by zones of discontinuous reflection character, probably corresponding to a deformed and intruded metamorphic terrain. These zones suggest a preferred horizontal fabric in the middle and lower crust.

Undoubtedly, as more deep reflection data become available, a better perspective with which to interpret these results will develop. New approaches and refinements in data acquisition and processing techniques, designed especially for application to the deep crust also promise to enhance the information carried in existing and future deep reflection records. For these reasons, alternatives to some of the interpretations favored here are not ruled out and in most cases have been presented for evaluation. In spite of the questions still outstanding these seismic sections provide a substantial new look at structural variations within continental lithosphere in the early stages of continental breakup.

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