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**SEISMIC EXPLORATION FOR SHALLOW
MAGMA BODIES IN THE VICINITY OF
SOCORRO, NEW MEXICO**

MASTER

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SOCORRO, NEW MEXICO

Final Report
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SEISMIC EXPLORATION FOR SHALLOW MAGMA BODIES
IN THE VICINITY OF SOCORRO, NEW MEXICO

Technical Completion Report

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The second article, "Exploration Framework of the Socorro Geothermal Area New Mexico," was published in the New Mexico Geological Society's Special Publication No. 7.

The two appendices are Master's thesis produced by two students working on the project at NMIMT.

TABLE OF CONTENTS

	<u>Page</u>
Characteristics of Rio Grande Rift in Vicinity of Socorro, New Mexico from Geophysical Studies	1
Exploration Framework of the Socorro Geothermal Area, New Mexico.	7
Appendix I	
A study of Poisson's Ratio in the Upper Crust in the Socorro, New Mexico, Area Master's Thesis by John L. Fender	
Appendix II	
Microearthquake Frequency Attenuation of S Phases in the Rio Grande Rift near Socorro, New Mexico Master's Thesis by James A. Johnson	

CHARACTERISTICS OF RIO GRANDE RIFT IN VICINITY OF SOCORRO, NEW MEXICO, FROM GEOPHYSICAL STUDIES

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The Rio Grande rift has many unusual geological and geophysical characteristics in the area between Bernardo and San Marcial, New Mexico. In that area many geophysical investigations have been conducted. This paper summarizes the geophysical aspects of these studies that are significant to an understanding of this part of the rift.

Properties of the crust

GENERAL STRUCTURE—A north-south crustal profile of the Rio Grande rift through the Socorro region has been published by Sanford, Mott, and others (1977). Important features of the profile are 1) an upper crustal layer approximately 19 km thick with a velocity of 5.8 km/sec, 2) a lower crustal layer with a velocity of 6.5 km/sec and thickness of approximately 17 km beneath Socorro, and 3) a northward dip of about 2 degrees on the Moho, which overlies upper-mantle rock with a velocity of 7.9 km/sec. The most unusual characteristics of the general structure of the crust are the abnormally low upper-crust and upper-mantle velocities. The upper-crust velocity, 5.8 km/sec, is significantly lower than the 6.15 km/sec found beyond the margins of the rift. The low upper-crust velocity is probably caused by the pervasive fragmentation of the upper crust of the rift by normal faults (O'Connell and Budiansky, 1974). On the other hand, the low upper-mantle velocity is probably related to the mechanism that has produced crustal extension for the past 25-29 m.y. (Chapin and Seager, 1975).

Changes in crustal thickness to the east and west of Socorro

have been calculated from regional-gravity and seismic-refraction studies of mining explosions located at Tyrone, New Mexico, and Morenci, Arizona (Sanford, Mott, and others, 1977). Recently, the crustal structure east of the rift was determined from a seismic refraction profile extending northward from a large chemical explosion in the northern part of White Sands Missile Range (Olsen and others, 1977). These data indicate little change in Moho depth to the east of the Rio Grande rift but an increase in depth to the west. The gravity data (assuming the change in the Bouguer anomalies is totally the result of a 0.2 g/cm³ density contrast across the Moho) give an 8° dip S. 60° W., whereas the seismic data give 3.4° dip S. 40° W. and 1.1° dip S. 63° W. Regardless of which dip is used to the west, the geophysical data do not support the existence of a sharp, symmetrical mantle upwarp beneath the rift in the Socorro vicinity.

MAGMA BODIES IN THE CRUST—An unusual feature of the crust beneath the rift in the Socorro vicinity is seismic evidence for the existence of magma bodies. Two types of magma bodies have been detected. The first is an extensive (more than 1700 km²) sill-like body occurring at depths of 18-22 km. The second type of magma body is composed of small intrusives located above the southern end of the extensive magma layer (fig. 1).

Extensive mid-crustal magma body—The initial evidence for an extensive magma body at mid-crustal depths came from an analysis of the arrival times and amplitudes of two reflection phases on microearthquake seismograms (Sanford and Long, 1965; Sanford and others, 1973). These two reflections, S to P and S to S, are impulsive and have the same general frequency content as the direct S-phase. Recordings by instruments with a broad frequency response show that the S to S phase contains a wide range of frequencies—from 3 to 15 Hz (Rinehart, 1976). This observation indicates that the reflecting discontinuity is sharp and singular. In addition, the absolute strength of the reflections as well as the ratio of their amplitudes indicates that the discontinuity producing the reflections is underlain by magma (Sanford and others, 1973; Sanford, 1977b).

The minimum geographical extent of the magma has been determined from the geographical distribution of S to S reflections on microearthquake seismograms (Rinehart, 1976; Sanford, Mott, and others, 1977; Sanford, 1977b) and from exceptionally strong P-wave reflections on high-resolution seismic reflection profiles (Oliver and Kaufman, 1976; Brown and others, 1977). These latter reflections on the COCORP (Consortium on Continental Reflection Profiling) profiles have about the same depth and dip (6° northward) as the S-wave reflections from the top of the magma layer.

The available observational data indicate that the magma body is very thin relative to its extent. P-wave residuals for teleseisms and mining explosions (Yousef, 1977; Fischer, 1977; Tang, 1978) do not support the existence of a thick magma layer even after station corrections are applied. Analysis of the Gasbuggy refraction data (Topozada and Sanford, 1976) indicates that time delays for P_n arrivals passing through the magma layer cannot be much greater than 0.1 second. A layer of magma (full melt) 0.6 km thick would be sufficient to produce a 0.1-second time delay. Finally, no clearly defined reflection phases from the bottom of the magma layer have been identified on microearthquake seismograms. If such reflections exist, they must occur so closely in time to the reflections from the top of the magma body that they cannot be easily identified.

Small shallow magma bodies—Three types of observations suggest the existence of small magma bodies above the southern end of the mid-crustal magma layer: 1) the screening of SV waves (vertically polarized S waves), 2) the spatial distribution of Poisson's ratio, and 3) the spatial distribution of microearthquake hypocenters (Sanford, 1976).

For many microearthquakes in the Socorro vicinity, SV waves are absent or extremely weak on seismograms for one or more stations in the array (Shuleski, 1976; Sanford, Rinehart, and others, 1977). About 40 percent of these observations can be explained by the fault mechanism of the microearthquakes; the remaining 60 percent apparently result from SV screening by molten or partially molten rock bodies (Kubota and Berg, 1968; Matumoto, 1971). Four shallow magma bodies west and southwest of Socorro can account for most of the observed screening. Although the precise location, shape, and volume of the intrusives have not been determined, approximate locations for segments of the crust containing the shallow magma bodies are given in fig. 1.

From an analysis of S-P times and P travel times for microearthquakes, the spatial distribution of Poisson's ratio in the upper crust has been found (Caravella, 1976). Several pie-shaped segments in the upper crust with Poisson's ratios greater than 0.29 have been located near Socorro. Although these high Poisson's ratios could result from compositional changes in the crustal rocks, they can also be explained by the same general distribution of shallow magma bodies used to account for SV screening.

An analysis of the spatial distribution of microearthquake foci shows activity surrounding, but not within, the crustal segments containing the shallow magma bodies (Sanford, Rinehart, and others, 1977). Above these regions, microearth-

quake foci are never deeper than about 2½ km, whereas in adjacent areas, activity occurs to depths of 8 km or more. Earthquake activity is not expected in segments of the crust containing magma because high temperatures produce stable sliding rather than stick-slip movement along faults.

To date we at New Mexico Tech have only looked for shallow magma bodies above the southern end of the extensive magma body. Thus we cannot be certain that this region is the only one where magma may leak upward through the crust from the deeper magma body.

Natural seismicity

The pattern of natural seismic activity in space and time suggests recent intrusion of magma into the crust in the vicinity of Socorro. Elastic strain is being released in this region by large numbers of earthquakes, most occurring in swarms. Earthquake swarms are observed in the vicinity of active volcanoes and in regions that have had volcanic activity in recent geologic times (Richter, 1958).

Descriptions of the seismic activity in the Socorro vicinity since 1960 have appeared in a number of publications (Sanford and Holmes, 1961; Sanford and Holmes, 1962; Sanford, 1963; Sanford and Singh, 1968; Sanford and others, 1972; Singh and Sanford, 1972; Sanford, Mott, and others, 1977). The principal characteristics of the seismic activity are listed below:

- 1) Unlike the Albuquerque-Belen Basin to the north, elastic strain is being released in the Socorro area by hundreds of earthquakes, nearly all in the microearthquake range (local magnitudes less than 3.0).
- 2) The seismic activity is diffusely distributed over a 2,000 km² area roughly centered on the deep, extensive magma body.
- 3) The activity within this area does not correlate closely with the spatial distribution of known major faults.
- 4) A composite fault-plane solution for microearthquakes in the Socorro area indicates that the microearthquakes are the result of downdip movement on faults with near north-south strike.
- 5) Approximately 75 percent of the microearthquake activity is occurring in swarms (defined as six or more shocks from the same focal region in 24 hours).

A characteristic of historical as well as of recent seismic activity is the occurrence of a majority of earthquakes in swarms. One of the historical swarms from 1906 to 1907 was major, comparable in magnitude to the famous 1965-67 Matsushiro swarm, which Stuart and Johnston (1975) attribute to magmatic intrusion at shallow depth. During the 1906-07 Socorro swarm, shocks were felt almost every day and at times reached a frequency of one perceptible tremor an hour. Three shocks of this swarm reached a Rossi-Forel intensity of VIII at Socorro and were felt over areas of about 275,000 km² (Reid, 1911; Sanford, 1963). The local magnitudes of these quakes is estimated to be near 6.

Rift structure

A gravity survey (fig. 2) indicates that the long, narrow Socorro Basin is composed of three linked structural depressions. The two depressions adjacent to the Socorro-Lemitar Mountains are asymmetrical. Narrow fault zones with large displacements, perhaps as great as 3.5 km, border the western margins of these downdropped crustal blocks. From their deepest points west of the Rio Grande, these depressions rise

FIGURE 1—GEOGRAPHICAL EXTENT OF THE EXTENSIVE MAGMA BODY. A solid line indicates the boundary is closely defined; a dashed line denotes some uncertainty of its location. Positions of the COCORP lines and the shallow magma bodies (heavily hachured areas) are also shown.

FIGURE 2—BOUGUER GRAVITY MAP OF THE SOCORRO AREA; terrain and regional corrections have been applied (from Sanford, 1968).

fairly gradually to the east, probably by a combination of step faulting and tilting. Some interpretations of the gravity data also indicate the possibility of buried horst blocks beneath Socorro Basin.

La Jencia Basin is not as narrow as Socorro Basin nor as structurally asymmetrical. However, the total structural relief for La Jencia Basin is comparable to that for the structural depressions comprising the Socorro Basin—on the order of 3 km. Thus the geophysical data are compatible with geologic information indicating a broader basin in the Socorro area prior to uplift of the Socorro-Lemitar Mountains 9-10 m.y. ago (Denny, 1941; Chapin and Seager, 1975).

The gravity data also support geologic mapping that indicates the existence of a large caldera south of Socorro Peak (Chapin and others, 1978). Gravity coverage in the Jornada del Muerto Basin east of the Rio Grande Valley is not great. Recently Schlue (1978) completed a gravity survey in this basin that extends detailed coverage northward from 33° 55' N. to 34° 00' N. between 106° 37' W. and 106° 42' W. His study indicates the basin is open further to the north than shown in fig. 2. The gravity anomalies north of Socorro Peak are substantially more positive than those to the south, although the topographic relief is similar. These observations can readily be explained by low-density deposits in the moat region of the caldera.

Fig. 3 shows a previously unpublished Bouguer gravity map covering a section of the rift that includes the transition from the Albuquerque-Belen Basin to the Socorro Basin. The map, based on work by Wongwiwat (1970), Oralratmanee (1972), and others, does not have the regional gravity removed. Inclusion of the regional gravity does not seriously affect a qualitative interpretation of near-surface geology from the map, because the regional variations in gravity are not great east of longitude 107° W. (Sanford, 1968). In addition, no terrain corrections were applied to the observations. With the exception of gravity stations in the northeast corner of the map, this correction is negligible.

The gravity map clearly shows the structural constriction between the Albuquerque-Belen Basin and the Socorro Basin. The Bouguer anomaly on the saddle between the two basins is about 15 milligals positive relative to the basins. The gravity map also indicates complex horst-and-graben structure beneath the southern end of the Albuquerque-Belen Basin. Details of the complex structure are particularly well shown on detailed seismic reflection profiles 1 and 1A (COCORP data). Two major horst structures revealed by these geophysical surveys appear to be subsurface extensions of the Ladron and Joyita uplifts.

Other geophysical observations

Additional geophysical observations of importance in the Socorro area are high temperature gradients and heat flows and surface uplift, all possibly related to the intrusion of magma into the crust. Temperature gradients as high as 241° C/km and heat flows as high as 11.7 HFU (heat-flow units) have been measured in the Socorro Mountain block (Reiter and Smith, 1977; Sanford, 1977a). Although movement of warm ground water upward along fracture zones is a possible explanation for the high temperature gradients (Reiter and others, 1978), two observations other than the seismic studies suggest these gradients are the result of magmatic intrusion into the upper crust. First, gradients as high as 184° C/km have been measured about 2 km west of the major frontal fault bordering the eastern margin of the Socorro Mountains

(Sanford, 1977a). Second, the residence time of the water passing through Socorro Peak is extremely short—approximately four years (Holmes, 1963); this residence time suggests that the water would not have had time to redistribute heat from great depths in the crust.

Uplift of the Socorro area in very recent time has been established by Reilinger and Oliver (1976) from an analysis of level-line data for the period 1909-1952. Their results for a single north-south line indicate surface uplift roughly coincident with the extent of the extensive magma body. The maximum average rate of uplift, about 6 mm/yr, has occurred approximately 20 km north of Socorro.

ACKNOWLEDGMENTS—Many of the studies described in this paper were conducted by undergraduate and graduate research assistants at New Mexico Tech. Where possible, I have acknowledged their work in the text; otherwise it appears in the references. The geophysical work in the Socorro vicinity has been supported by a number of organizations over the past 20 years but primarily by the State of New Mexico and the National Science Foundation.

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FIGURE 3—BOUGUER GRAVITY MAP OF THE SAN ACACIA-BERNARDO AREA; terrain and regional corrections have not been applied.

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EXPLORATION FRAMEWORK OF THE SOCORRO
GEOTHERMAL AREA, NEW MEXICO

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Figure 1. Generalized map of the Rio Grande rift (after Chapin, 1971; Chapin and Seager, 1975) and major crustal lineaments.

EXPLORATION FRAMEWORK OF THE SOCORRO GEOTHERMAL AREA, NEW MEXICO

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INTRODUCTION

The Socorro geothermal area is located about 75 miles (120 km) south of Albuquerque, New Mexico, at the town of Socorro (fig. 1). Geothermal activity has been known in the area for a long time. In fact, Socorro owes its existence to two warm springs on the southwest edge of town which have provided a reliable source of potable water since before the Spanish settlement of New Mexico. These springs (plus a third man-made spring) still supply a major portion of the town's water needs. Attention has been focused on the Socorro area in recent years because of studies of microearthquakes and magma bodies by Caravella (1976), Fischer (1977), Rinehart (1976), Sanford (1977a), Sanford and Long (1965), Sanford and others (1973, 1977a, 1977b), Shuleski (1976), Shuleski and others (1977); heat flow by Reiter and others (1975), Reiter and Smith (1977), Sanford (1977b); modern uplift by Reilinger and Oliver (1976); and deep crustal structure by Sanford (1968), Toppozada and Sanford (1976), Oliver and Kaufman (1976), and Brown and others (1977).

In April 1975, the New Mexico Bureau of Mines and Mineral Resources began a detailed geologic study of the Socorro area in anticipation of probable geothermal exploration and development. Since July 1976, the project has been funded by a grant from the New Mexico Energy Resources Board through the Energy Institute at New Mexico State University. The geophysical studies of Sanford have been supported by a series of grants from the New Mexico Energy Resources Board and the National Science Foundation. In 1976 the U.S. Geological Survey designated approximately 140 square miles (362 km²) in and around Socorro as the Socorro Peak Known Geothermal Resources Area (KGRA). The U.S. Bureau of Land Management designated a much larger area (624,814 acres) as the Socorro Peak Geothermal Leasing Area. The first competitive lease sale was held in November 1977 with nine tracts totaling 17,000 acres being leased within the Socorro Peak KGRA for \$275,411.46.

GEOLOGIC SETTING

The Socorro geothermal area is located within the Rio Grande rift (fig. 1), where the rift transects the north-eastern portion of the Datil-Mogollon volcanic field of Oligocene to early Miocene age. The north-trending fault-block ranges of the rift expose thick sequences of rhyolitic ash-flow tuffs overlain by, and interbedded with, basaltic-andesite flows (fig. 3). Beneath the ash-

flow tuffs are latitic conglomerates, mud-flow deposits, and sandstones representing the alluvial apron which surrounded the Datil-Mogollon field prior to its ignimbrite climax. The base of the volcanic pile rests unconformably upon rocks ranging from late Eocene to Precambrian in age. Most of the area west of the Rio Grande and south of San Acacia lies on the northeast flank of a major Laramide uplift from which the Mesozoic rocks were stripped by early Tertiary erosion. Basal volcanic rocks resting upon late Paleozoic limestones, quartzites, or shales on the uplift lap off the structure onto Eocene arkosic sediments which overlie Cretaceous sandstone and shales. Precambrian rocks in the Socorro area generally consist of low-grade metasedimentary rocks (especially quartzites and argillites), metavolcanic rocks, and intrusive rocks ranging from large granitic plutons to gabbroic stocks and diabase dikes. The high-potassium Precambrian granites (Condie, 1978) may be the main source of K₂O for the potassium anomaly described in a later section.

Two major crustal lineaments (fig. 1) intersect in the Socorro area. The Morenci lineament is one of a series of northeast-trending shear zones that dominate the structural grain of Precambrian rocks in the southern Rocky Mountains. Documentation for each of the lineaments labeled on Figure 1 will be presented in a later paper. They are included on Figure 1 to illustrate how structures in the Socorro area fit into the regional framework. The three northern lineaments are en echelon segments of the Colorado lineament of Warner (1978). Of these, the Silverton and Idaho Springs lineament are generally lumped as the well-known Colorado mineral belt (Tweto and Sims, 1963). The Capitan lineament parallels the better known (and highly controversial) Texas lineament (Albritton and Smith, 1956; Wertz, 1970) and may be related to it. The important point is that these lineaments are deeply penetrating flaws in the lithosphere that tend to "leak" magmas and to influence deformation in the brittle near-surface rocks. The Morenci lineament has had a major influence on tectonics and magmatism in the Socorro area for the past 32 m.y. The recognition that a transverse shear zone of the Morenci lineament separates fields of tilted blocks undergoing rotation and step faulting in opposite directions has been one of the most significant discoveries of our geologic research. This shear zone is acting as an incipient transform fault connecting en echelon segments of the Rio Grande rift. Figure 2 and Plate 1 (in back pocket) document the shear zone and illustrate its effect on magmatism, both past and present.

Figure 2. Overlapping and nested cauldrons and a transverse shear zone (double line) in the Socorro-Magdalena area. Base is composite of Socorro and Tularosa 2-degree quadrangles (Army Map Service plastic relief maps). From oldest to youngest, the cauldrons and their ash-flow tuff sheets are: 1) Nogal Canyon—Vicks Peak, 2) North Baldy—Hells Mesa, 3) Mt. Withington—A-L Peak Tuff, gray-massive member, 4) Magdalena—A-L Peak Tuff, flow-banded member, 5) Sawmill Canyon—A-L Peak Tuff, pinnacles member, 6) Mt. Withington—Potato Canyon Tuff, 7) Socorro—tuff of Lemitar Mountains, 8) Hop Canyon—tuff of South Canyon. Nogal Canyon and Mt. Withington cauldrons after Deal and Rhodes (1976).

Post Santa Fe Alluvium: (0-80', 15 m) ALLUVIUM, SLUMP BLOCKS, and COLLUVIUM: inset deposits of gravel, sand, & mud along the Rio Grande & its tributaries. Fan, piedmont-slope and playa deposits of La Jencia basin. Extensive slump & colluvial deposits on slopes around Socorro Pk., Strawberry Pk., & Black Mesa mantle mudstones of upper Popotosa Fm.

Sierra Ludrones Formation: (0-1000', 305 m) PIEDMONT-SLOPE, RIVER CHANNEL and FLOOD PLAIN DEPOSITS, and BASALT FLOWS: buff to pale-rd., poorly indur., fangl. w/ mixed clasts of all major rock types shed off present highlands. Fangl. intertongue w/ lt.-gry., friable, well-sorted, fn.-med.-gnd. sands of ancestral Rio Grande. Flood plain deposits are rd. & grn. mudst. & siltst. Basalt of Sedillo Hill (4 m.y.) underlies piedmont-slope facies & overlies ancestral Rio Grande sands & piedmont facies near Grefco mine.

Rhyolite of Socorro Peak: 7-12 m.y. (0-800', 244 m) SILICIC DOMES, FLOWS, and TUFFS: pale-rd.-gry. to lt.-gry. rhyolite & rd.-brn. to lt.-gry. rhyodacite flow-banded lavas, domes, necks, & tuffs. Tuffs interbedded w/ clayst. of upper Popotosa at Socorro Spring, N. of Blue Canyon, Kelly Ranch (12 m.y.), and N. of Sedillo Hill.

Upper Popotosa Formation: (0-2500', 762 m) CLAYSTONES, MUDSTONES, SILTSTONES, SANDSTONES, CONGLOMERATES, and BASALT FLOWS: Dk.-rd.-maroon gyp. playa clayst. & mudst. w/ thin. bds. of yel.-brn. to grn.-gry. clayst. North of Strawberry Pk., playa deposits intertongue w/ buff. cgl. ss. w/ hydro. altered clasts derived from Magdalena area. At Grefco mine, mudst. intertongue w/ pale-rd. cgl. ss. derived from E. of Rio Grande and w/ perlitic-rich cgl. shed from Grefco dome. Basalt flows interbedded w/ playa deposits near Pound Ranch, Strawberry Pk., & Hwy. 60.

Lower Popotosa Formation: (0-1500', 457 m) MUDFLOW DEPOSITS, FANGLOMERATES, and minor LACUSTRINE DEPOSITS: rd.-brn., v. well-indur., heterolithic, crs. cgl. w/ abun. clasts of underlying volc. units & locally grading downward into ancient colluvial breccias. Gry. siltst., mudst., & thin. fresh-water ls. intertongue w/ lt.-gry. to prpl.-gry. cgl. ss. in Socorro Pk.-Blue Canyon area. Facies relationships reflect late-stage topography of Socorro caldera.

Lavas of Water Canyon Mesa: 20 m.y. (0-600', 183 m) INTERMEDIATE to SILICIC LAVAS and TUFFS: rd. silicic lavas & tuffs underlain by intermed., dk.-gry. to dk.-rd.-brn., dense lavas; intruded and overlain by flow-banded wht. rhyolite domes & flows.

Tuff of South Canyon: 26 m.y. (0-600', 183 m) ASH-FLOW TUFFS: multiple-flow, simple cooling unit of rhyolite ash-flow tuffs. Lower zone of lt.-gry., poorly to densely welded, xl.-poor tuff w/ "botryoidal" pumice & lithic-rich lenses. Upper zone of med.-gry. to prpl.-gry., mod. to densely welded, mod. xl.-rich tuff w/ chatoyant sandstone & euhedral qtz.

Unit of Luis Lopez: (0-2000', 610 m) ASH-FLOW TUFFS, ANDESITE LAVAS, RHYOLITE DOMES and TUFFS, minor LANDSLIDE DEPOSITS and TUFFACEOUS SANDSTONES. Moat fill of Socorro cauldron. Lt.-gry., pumiceous, lithic-rich, poorly welded, rhyolite ash-flow tuffs intertongue w/ andesite flows, vent agglomerates, & minor andesitic laharic breccias. Pale-rd., org.-rd., & lt.-gry. flow-banded rhyolite lavas, domes and tuffs cap sequence. Contemporaneous tongue of La Jara Peak Basaltic Andesite (0-800', 244 m) in Lemitar Mts.

Tuff of Lemitar Mountains: 27 m.y. (0-400', 122 m outflow; 700-2900', 213-884 m cauldron) ASH-FLOW TUFFS: multiple-flow, comp. zoned, simple to compound cooling unit of densely welded tuff. Lt.-gry. to pale-rd., xl.-poor lower member 200-800 ft. (61-244 m) thk. in Socorro cauldron but only locally present as thin. lenses on outflow sheet. Md.-rd., xl.-rich, qtz. & biotite-rich, 2-feld. upper member 0-400 ft (122 m) outflow facies. 350-2900 ft (107-884 m) cauldron facies. Contact sharply gradational & welded. Xl.-rich clots common at top. Lithic-rich zones w/ Precambrian & andesite clasts widespread S. of Tower mine.

Unit of Sixmile Canyon: (0-2000', 610 m) ANDESITE to BASALTIC ANDESITE LAVAS, RHYOLITE LAVAS and DOMES, ASH-FLOW TUFFS, LAHARIC BRECCIAS, and SANDSTONES: post-collapse fill of Sawmill Canyon cauldron. Dk.-gry. mafic lavas intertongue w/ th. & surround other rock types. Heterolithic breccias along cauldron walls. Contemporaneous tongue of La Jara Peak Basaltic Andesite (0-500', 152 m) in Lemitar Mts.

A-L Peak Tuff: 32 m.y. (0-700', 213 m outflow; 2000', 610 m cauldron) ASH-FLOW TUFFS: composite sheet w/ basal gray-massive, middle flow-banded, & upper pinnacles members. All densely welded, xl.-poor, l-feld., rhyolite, ash-flow tuffs. Sources are Mt. Withington, Magdalena, & Sawmill Canyon cauldrons. Tongue of La Jara Peak Basaltic Andesite (0-200', 61 m) usually separates flow-banded & pinnacles members.

Lava Flows: (0-300', 91 m) ANDESITE to BASALTIC ANDESITE: unnamed tongue of dense prpl.-rd. flows locally present in SW Lemitar Mts.

Hells Mesa Tuff: 32-33 m.y. (0-500', 152 m outflow; 3000', 915 m cauldron) ASH-FLOW TUFFS: rhyolite, multiple-flow, simple cooling unit, densely welded, xl.-rich, qtz.-rich, 2-feld., mass. tuffs. Pk. to rd.-brn. when fresh, gry. when propyl. altered. Weathers to blocky bldrs. Basal tuffs similar to tuff of Granite Mtn.; abrupt increase in qtz. 10-25 ft above base.

Spears Formation - tuff of Granite Mountain: (0-100', 30 m) ASH-FLOW TUFFS: qtz.-latite, multiple-flow, simple cooling unit, densely welded, xl.-rich, lithic-rich, qtz.-poor, mass. tuffs; rd.-brn. when fresh, dk.-grn.-gry. when propyl. altered.

Spears Formation - volcaniclastic unit: 37-33 m.y. (0-1500', 457 m) CONGLOMERATES, MUDFLOW DEPOSITS, SANDSTONES, LAVAS, and ASH-FLOW TUFFS: volcaniclastic apron of latite to andesite comp., becomes coarser & contains more volc. units up & to south.

Madera Limestone: (500-1500', 152-457 m) LIMESTONES: thk., homo. sequence of gry. to blk. micrites w/ a few thin. bds. of med.-crs.-gnd. qtz. ss. Up. 200-300 ft (61-91 m) is rd., grn., & gry. micrites grading up. into arkosic bds. of Abo Fm. Nodular micrites abun. throughout.

Sandia Formation: (550-650', 168-198 m) SHALES, QUARTZITES, and LIMESTONES: gry. to blk., sdv., carb. sh. and siltst. w/ thin. bds. of gry. to brn., med.-crs.-gnd. qtz. ss. (abun. near base) & gry., med.-gnd. micritic ls. (gradational w/ Madera).

Kelly Limestone: (0-90', 27 m) LIMESTONE: lt.-gry., med.-crs.-gnd., thk.-bdd. crinoidal biosparrites.

Caloso Formation: (0-30', 9 m) LIMESTONES and CONGLOMERATES: basal arkosic cgl. overlain by gry., mass., sdv. or pluv. micrites.

Basement: ARGILLITES, QUARTZITES, VOLCANIC and PLUTONIC ROCKS: thk. sequence of low-grade metased. & metavolc. rocks intruded by granitic & gabbroic plutons, & diabase dikes.

Figure 3. Composite stratigraphic column of the Socorro area. Thickness of units not to scale.

In addition to exposing a good cross section of volcanic and pre-volcanic rocks, the north-trending, tilted, fault-block uplifts also transect several cauldrons and related vent areas. The area from Socorro southwestward to beyond the San Mateo Range, a distance of about 50 miles (80 km), consists of at least seven overlapping cauldrons (fig. 2). These circular to elliptical subsidence structures formed by collapse of the roofs of large, shallow magma chambers as a consequence of the voluminous ash-flow eruptions which deposited the widespread tuff sheets that cap the Datil-Mogollon volcanic pile. The north margin of one of the youngest of these cauldrons is exposed on Socorro Peak; the ring-fracture zone and moat of this cauldron partly control the location and plumbing system of the Socorro geothermal area. The Socorro cauldron formed about 27 m.y. ago during onset of the extensional stress field which pulled apart the earth's crust along the Rio Grande rift. As a result, fault patterns within the caldera interrelate in a complex manner with those outside the structure.

After collapse, the floors of many cauldrons are domed upward by renewed magma pressure to form a central resurgent dome separated from the cauldron walls by a trough-like annulus called a moat (Smith and Bailey, 1968). Because the moat is underlain by the deeply penetrating ring fractures of the cauldron, post-collapse magmatism usually fills it with lava flows, lava domes, and local tuffs which are interbedded with sediments washed in from the resurgent dome and from the outer cauldron walls. Volcanism may continue along the moat for many millions of years after cauldron collapse because the underlying ring fractures provide the easiest route of ascent for magmas (Lipman and others, 1976). Likewise, the ring-fracture zone often controls geothermal systems by providing magma bodies and highly fractured rock beneath a permeable sequence of moat deposits. This idealized picture is true in a general way for the Socorro cauldron. Cauldron collapse was followed by resurgent doming and accumulation of moat volcanics and sediments. After a long lull in magmatism, during which the cauldron was largely buried by sediments of the Popotosa basin, renewed magmatism occurred during the interval 12 to 7 m.y. ago. After another quiet spell, basaltic lava flows were emplaced at about 4 m.y. ago (Bachman and Mehnert, 1978). Volcanism again became dormant, but the geophysical studies of A. R. Sanford and others indicate that magma bodies are again rising along, and near, the ring fracture zone (pl. 1, map 4).

Another important element in the geologic setting is the formation and disruption of a broad sedimentary basin which spanned the Rio Grande rift in the Socorro area during the interval from about 26 m.y. to 7 m.y. ago. The basin extended from the Gallinas Range on the west to the high mesas east of the Rio Grande and from the Ladron Mountains on the north to the eastern Magdalena and central Chupadera mountains on the south. Within the basin accumulated several thousand feet of alluvial-fan, piedmont-slope, fluvial, and playa deposits now known as the Popotosa Formation (Denny, 1940; Bruning, 1973). The lowest portions of the basin floor persistently occupied a north-trending area along the present site of the Chupadera, Socorro, and Lemitar

mountains. Alluvial fans graded downslope from the bordering uplifts to playa lakes along this depression. At least 800 to as much as 2500 feet (244-762 m) of relatively uniform, gypsiferous clays were deposited from the playas. During the interval 12 m.y. to 7 m.y. ago, rhyolitic magmas rose along the northern moat of the buried Socorro caldera, spread over the thick clay deposits, and lapped onto bedrock highs along the east side of Socorro Peak and in the northern Chupadera Mountains and eastern Magdalena Mountains. Chemical and petrographic data suggest that a very large geothermal system existed at the time.

The Popotosa basin was disrupted sometime after about 7 m.y. and before 4 m.y. by extensional forces which broke the basin into tilted fault-block uplifts and grabens. The reason for this disruption is not known. It coincides in time with a pulse of uplift and faulting throughout the Rio Grande rift, but it may also be related, in part, to intrusion of a large sill-like magma chamber similar to the one now present at a depth of 18 to 20 km (Sanford and others, 1973, 1977a). The area occupied by the playa lakes was uplifted to form the Chupadera-Socorro-Lemitar mountains. Many slopes in the Socorro Peak, Black Mesa, and Strawberry Peak areas are underlain by red Popotosa claystones; slump blocks of the overlying volcanic rocks are riding downslope on the claystones and have masked much of the underlying bedrock (pl. 1, map 1). Drill holes along the canyon bottoms indicate that several hundred feet of claystone is yet to be exposed.

Extensive pediment surfaces were carved across uplifted and tilted Popotosa beds and older rocks on the Chupadera-Socorro-Lemitar horst during the Pliocene and Pleistocene periods. The rock removed was dumped mainly into the Socorro basin on the east. Sometime during the breakup of the Popotosa basin, the regional drainage was integrated to form an ancestral Rio Grande. Extensive deposits of light-colored, well-sorted arkosic sands were laid down by this river and some of its larger tributaries. These sands underlie much of the Rio Grande valley east of the frontal faults of the Chupadera-Socorro-Lemitar uplift and provide an important aquifer for New Mexico Institute of Mining and Technology and the Town of Socorro. Broad, gently sloping piedmont plains graded to the ancestral Rio Grande and on these accumulated gravels, sands, and silts which interfinger with the clean sands and muddy overbank deposits of the trunk stream. Both the piedmont and main-stem fluvial deposits have been named the Sierra Ladrones Formation (Machette, 1977). These deposits locally overlie the Popotosa Formation with angular unconformity and are lighter in color, less indurated, and contain a larger portion of clasts derived from prevolcanic rocks. At about 4 m.y. ago, basalt flows were erupted in the Sedillo Hill area, south of Socorro Peak. These lavas flowed eastward down a pediment surface, across truncated Popotosa claystones and silicic domes and flows of the rhyolite of Socorro Peak, and onto fluvial sands of the ancestral Rio Grande.

Capture of the ancestral Rio Grande at El Paso in middle Pleistocene time (between 300,000 and 500,000 years ago; Kottowski, 1958; Hawley, 1975), and its integration to the sea, lowered its base level and caused it

MAPS*

*The following few pages are black and white photoreduced copies of maps from this article. The originals in full color may be obtained from the New Mexico Geological Society Special Publication No. 7.

EXPLANATION

Volcanic vents:

- basalt of Sedillo Hill
- rhyolitic lavas of Socorro Peak
- rhyolite } related to the Socorro cauldron
- ▲ andesite }
- inferred structural margin of the Socorro cauldron (mostly buried)

- normal fault
- low-angle normal fault
- dike
- spring
- thermal spring
- 20° strike and dip of strata

Santa Fe Group	Qa	Upper Quaternary alluvium; gravel, sand and mud of major arroyos and the Rio Grande Valley, local alluvium and colluvium.
	QTa	Plio-Pleistocene basin and valley fill (largely equivalent to Sierra Ladrones Fm.); poorly consolidated piedmont-slope fangls. intertonguing with friable ancestral Rio Grande sandst. and flood-plain siltst. and mudst., includes some post-Santa Fe Group terrace deposits.
	Tb	Basalt flows of Sedillo Hill (4 m.y.); interbedded in QTa.
	Ts	Rhyolite to rhyodacite domes, flows, necks and tuffs of Socorro Peak (12-7 m.y.); tuffs interbedded in upper Popotosa included in Tpu.
	Tpu	Upper Popotosa Fm.; gypsiferous playa clayst. and mudst. with minor intertonguing fangls. and channel sandst., and several interbedded basalt flows (largely masked by landslides and colluvium).
	Tpl	Lower Popotosa Fm.; well-indurated red mud-flow deposits and fangls. intertonguing with minor prpl.-gry. fangls. and lacustrine siltst. and mudst.
	Ti	Intrusive rocks; stocks and dikes of silicic to andesitic composition (Oligocene-Miocene).
	Tv	Volcanic rocks post-dating tuff of Lemitar Mts. (26-20 m.y.); lithic-rich tuffs, andesite flows and rhyolite domes of Socorro cauldron moat, local intermediate lavas and rhyolite domes on Water Canyon Mesa, includes undiff. regional ash-flow tuffs and basaltic-andesite lavas in Lemitar and Magdalena Mts.
	Tl	Tuff of Lemitar Mountains (27 m.y.) compositionally zoned, densely welded, rhyolite ash-flow tuff erupted from the Socorro cauldron.
	To	Older volcanic rocks (37-32 m.y.) predating tuff of Lemitar Mts.; latitic-andesitic cgl. capped by thick sequence of densely welded, xl.-rich and xl.-poor rhyolite ash-flow tuffs and some associated moat-fill deposits.
	P	Paleozoic limestones, shales and sandstones (Mississippian-Pennsylvanian)
	pG	Precambrian granites, gabbros, diabase dikes, metavolcanic and metasedimentary rocks.

GEOLOGIC CROSS SECTIONS

GENERALIZED GEOLOGIC MAP OF THE SOCORRO GEOTHERMAL
AREA, NEW MEXICO

SCALE 1:62,500

geology by R.M. Chamberlin and G.R. Osburn, 1976-77

MAP OF MAGMA BODIES, THERMAL-GRADIENT MEASUREMENTS, AND MICROEARTHQUAKE EPICENTERS FOR THE SOCORRO GEOTHERMAL AREA, NEW MEXICO

SCALE 1:62,500

outline of deep magma body (depth: 18-22 km)

outline of shallow intrusive zone (minimum depth: 5 km)

△ seismograph station

× microearthquake epicenter

● boreholes for thermal gradient measurements

● 54.7 measured gradient, °C/km } data from A.R. 15-76 depth interval, meters } Sanford, 1977

● 159 measured gradient, °C/km } data from: M. Reiter 23-40 depth interval, meters } and R. Smith, 1977

● 9.7 mWU calculated heat flow (μ call cm²-sec) 11.7 mWU

◆ borehole in mine, collapsed 125 m below surface

• • • inferred structural margin of the Socorro cauldron

magma bodies and microearthquake data from: A.R. Sanford, E.J. Rinehart, P.J. Shuleski and J.A. Johnston, 1977

18

MAP 4

to entrench deeply into its earlier deposits. The highest outcrops of ancestral Rio Grande deposits in the Socorro area are about 600 feet (180 m) above its present floodplain; about half the elevation difference is due to entrenchment and half to uplift along the east flank of the Chupadera-Socorro-Lemitar horst. The present landscape consists of four elements: 1) the mountain ranges, 2) the present floodplain of the Rio Grande, 3) broad piedmont surfaces of Pleistocene to Holocene age, and 4) arroyos which grade to the present Rio Grande and are dissecting the piedmont slopes. The largest of the arroyo systems, such as Socorro Canyon and Nogal Canyon, head in the Magdalena Range and cross the Chupadera-Socorro-Lemitar uplift. This is an anomalous situation which probably occurs because the Chupadera-Lemitar block was uplifted within a broad basin and beveled by a Pliocene pediment. During subsequent uplift streams have been able to maintain their courses across the upthrown block.

The geophysical characteristics of the Socorro area and their spatial relationships to the Socorro cauldron and the transverse shear zone are shown on a series of maps accompanying the generalized geologic map (pl. 1, maps 1-4 in pocket). Map 2 is a residual Bouguer gravity map (Sanford, 1968); Map 3 is an aeromagnetic map prepared by the U.S. Geological Survey. Map 4 shows the extent of a deep magma body at about 18 km depth and several shallow magma bodies inferred by Sanford and others from seismic studies. The distributions of microearthquakes (Sanford and others, 1977a, b) and heat flow measurements (Reiter and Smith, 1977; Sanford, 1977) are also shown on this map. In summary, the magma bodies, seismic activity, high heat flow, geothermal activity, and modern uplift of the Socorro area occur in a geologic setting which has experienced repeated magmatism in the past and is likely to again in the future.

STRUCTURAL CONTROLS

Structural controls of the Socorro geothermal area may be divided into those that: 1) control the ascent of magma bodies, i.e. the location of heat sources; 2) control the location and thickness of permeable reservoir rocks and impermeable cap rocks, hence, the lateral migration and storage of geothermal fluids; 3) control the upward migration of geothermal fluids; and 4) control the fracture trends and fracture spacing in rocks whose reservoir characteristics are dependent upon fracture permeability. These controls are interrelated; their relative importance to discovery and development of a commercial geothermal reservoir in the Socorro area is yet to be tested.

Structural Controls of Magmatism

The major structural controls of the Socorro geothermal area are the Rio Grande rift, a transverse shear zone of the Morenci lineament, and the Socorro cauldron. The ring fracture zone of the Socorro cauldron has influenced the ascent of magmas at intermittent periods since cauldron formation. K-Ar dates on volcanic rocks indicate periods of magmatism at 27-20 m.y., 12-7 m.y. and 4 m.y. Geophysical data indicates that magma bodies are again present in the plumbing system (pl. 1, map 4). Reiter and others (1978) have recently pointed

out the association of several heat-flow anomalies along the Rio Grande rift with cauldron margins.

Until very recently, we thought that the Socorro cauldron margin was the most important structural control of magmatism other than the Rio Grande rift. However, when the detailed geologic maps of the eastern Magdalena Mountains (Osburn, 1978) and the Socorro Peak-Lemitar Mountains area (Chamberlin, 1978) were completed, a remarkable pattern began to emerge. To the north of a line extending southwestward from Socorro through the South Baldy area of the Magdalena Mountains, formations are tilted to the west and stepped down to the east by north-trending normal faults (pl. 1, map 1, cross section A-A'). To the south of this line, formations are tilted to the east and stepped down to the west (cross section B-B'). Examination of other geologic maps, both completed and in progress, revealed that this pattern persists southwestward at least as far as Mount Withington in the San Mateo Range and northeastward to the Mesa del Yeso area. The rotation and step faulting of rocks at the surface in opposite directions across this line must be accompanied by shearing in the underlying lithosphere. The sense of the shearing is left lateral, yet the shear zone connects en echelon segments of the Rio Grande rift which have a right-lateral apparent offset (Albuquerque graben to the northeast, Mulligan Gulch and Winston grabens to the southwest). Hence, the shear zone is acting as a transform fault connecting en echelon axes of extension. The rates of shearing and extension, however, are very slow compared to ocean ridges and the transform is in a very early stage of development. Barberi and Varet (1977) have recently described transverse structures, some of which are transform faults, in the Afar depression of East Africa. They note that the transverse structures connecting offsets of spreading centers tend to occur along prerift lines of tectonic weakness in the continental block; that transform faults in the initial stages of development are somewhat obscure and may be represented at the surface by zones of en echelon faults whose trends may diverge by 15° to 30° from the trend of the transform and whose motions are partly shear and partly extensional; that the common occurrence of volcanism along transverse structures suggests a tendency to "leak" magmas; that large central volcanoes tend to occur at the intersections of transverse structures with faults bordering depressions; and that magmas erupted along transverse structures are alkalic and contain peridotite xenoliths. All but the last of these characteristics are also true of the transverse shear zone at Socorro. The Socorro structure occurs along the Morenci lineament, a major northeast-trending flaw of probable Precambrian age; its surface expression is marked by a zone of jostled fault blocks a mile (1.5 km) or more in width across which the tilt of beds and sense of extension changes markedly; it has leaked magmas at irregular intervals since late Oligocene time and is leaking them today.

The deep magma body at about 18 km, outlined by Sanford and others (1977a) using reflections from microearthquakes, terminates in an irregular manner (pl. 1, map 4) against the transverse shear zone. The five shallow magma bodies inferred by Sanford and others (1977b) from changes in Poisson's ratio, screening of S

waves, and distribution of microearthquake foci, are distributed along the shear zone and lie within 3 or 4 miles (5-6 km) of either side of it. Thus, the transverse shear zone seems to influence both the lateral and vertical movement of magma. The deep magma body may terminate against the shear zone because of a structural "curtain" of sheared and recrystallized rock, and/or the compositions and physical properties of basement rocks may be different on opposite sides of the shear zone. Note on the aeromagnetic map (pl. 1, map 3) the change in trend of magnetic anomalies across the shear zone. Magmas may "bleed" upward along the shear zone and accumulate in shallow reservoirs. The north-south orientation of the shallow magma bodies (pl. 1, map 4) probably reflects the dominance of north-trending, rift-related fractures in the shallow crust and the favorability of fractures of this orientation for dilation and magma injection. The fact that the shallow magma bodies lie within or close to the Socorro cauldron could be fortuitous, but the well-documented relationship of recurrent magmatism to cauldrons elsewhere suggests that the cauldron ring fractures provide additional vertical permeability to the magma plumbing system. In the last several kilometers of ascent, the combination of north-trending rift fractures and intersecting cauldron ring fractures may tend to draw the magmas towards the cauldron by providing paths of least resistance.

Control of past magmatism by the transverse shear zone is evident from Figure 2 and Map 1 (pl. 1, in pocket). Going back in time, the vent for the 4 m.y. old basalt of Sedillo Hill lies on the shear zone; the two youngest silicic domes (Grefco dome, 7 m.y.; unnamed dome between Black Mesa and Socorro Spring, 8 m.y.) lie on the shear zone, the area of 7-12 m.y. old silicic volcanism is approximately bisected by the shear zone and all known vents for this period of magmatism lie within 3 or 4 miles (5-6 km) of it (pl. 1, map 1); and, finally, the seven overlapping and nested cauldrons of the Socorro-Magdalena area are clustered along the shear zone.

On the aeromagnetic map (pl. 1, map 3), the transverse shear zone is marked by a "quiet" zone which separates an area of north-trending magnetic anomalies to the north from an area of northeast-trending magnetic anomalies to the south. The transverse shear zone seems to be following a major discontinuity in the continental crust across which the structural grain of basement rocks changes from northerly to northeasterly. Along a zone about 8 miles (13 km) wide and extending from Socorro southwestward across Socorro Peak and the northern Chupadera Mountains through the Magdalena Range, the aeromagnetic anomalies are muted. In spite of rugged topography and exposure of Precambrian rocks at the surface, Socorro Peak is a relatively featureless area on the aeromagnetic map. The "quiet" zone extends across the highest and most rugged part of the Magdalena Range with bold aeromagnetic anomalies to the north and south but greatly subdued anomalies within it. The "quiet" zone seems independent of rock types and terrain. The most logical explanation is that high heat flow along the transverse shear zone has raised the Curie temperature isotherm to shallower depths and thus canceled the aeromagnetic signature of

the deeper crust. Considering the rather striking control of volcanism exerted by the transverse shear zone in the past, high heat flow along this zone seems a reasonable possibility.

Structural Controls of Reservoir Rocks and Cap Rocks

High heat flow above the shallow magma bodies may lead to development of geothermal reservoirs if: 1) permeable water-bearing reservoir rocks are present, and 2) an impermeable cap exists above the reservoir to prevent loss of fluid and dissipation of heat. Sanford and others (1977b) estimate depths of 4 to 5 km to the tops of the shallow magma bodies. At these depths, the magma bodies are within either Precambrian basement rocks or crystalline rocks of the late Cenozoic batholith of which the Socorro cauldron is a surface manifestation. Rocks at these depths are probably dry, but envelopes of high heat flow above the shallow magma bodies may intersect permeable water-bearing horizons at shallower depths. The depth of rock units with good aquifer characteristics is apt to be greatest within the Socorro cauldron because this subsidence structure has dropped pre-cauldron rocks an undetermined distance below the elevations at which they are normally found. The Socorro and La Jencia grabens (pl. 1, maps 1, 2) were superimposed across the Socorro cauldron during and after its collapse. Where these grabens overlap the cauldron, subsidence effects are additive so that potential reservoir rocks are depressed to the greatest depths. Where these zones of additive subsidence overlie shallow magma bodies, temperatures in potential reservoir rocks should be highest.

The Socorro cauldron overlaps and buries the eastern half of the older Sawmill Canyon cauldron (fig. 2). The buried ring fracture zone of this older cauldron is difficult to locate; however, the stratigraphic sequence in a drill hole at the Tower mine (pl. 1, map 1) suggests that it lies east of the drill hole. An unusual feature of the Socorro cauldron is the fact that moat deposits are very thick in the northern Chupadera Mountains, northeast of the Tower mine, and along the east-facing escarpment of Socorro Peak; whereas, from the Tower mine west, moat deposits are thin to absent and the tuff of Lemitar Mountains is only about one-third as thick as it is to the east. It seems possible that during subsidence of the Socorro cauldron the buried Sawmill Canyon cauldron behaved as a rigid, buoyant block which did not subside as much as the rest of the Socorro cauldron. Another possible explanation is that subsidence was much greater east of a major north-trending contemporaneous rift fault. A north-trending rift fault zone is known to control the thickness of the outflow sheet of the tuff of Lemitar Mountains north of the Socorro cauldron (pl. 1, map 1, cross section A-A'). Whatever the cause, or combination of causes, the eastern third of the Socorro cauldron seems to have subsided considerably more than the rest of the cauldron and accumulated as much as 3,000 feet (900 m) of cauldron facies tuff of Lemitar Mountains and as much as 2,000 feet (610 m) of moat deposits. Thus, the depth of pre-cauldron rocks may be as much as 5,000 feet (1.5 km) greater east of a line trending north-northwest from a point about 0.6 miles (1 km) east of the Tower mine to

the north edge of the Socorro cauldron. Alternatively, this buried escarpment may bend northwestward towards the mouth of South Canyon to follow the inferred margin of the buried Sawmill Canyon cauldron. However, gravity data (pl. 1, map 2) favors the first alternative.

Structural Controls of the Ascent of Geothermal Fluids

The location of the two warm springs southwest of Socorro seems to be controlled by the intersection of the moat of the Socorro cauldron with the major fault zone bounding Socorro Peak on the east. Downfaulting of the thick claystone interval in the upper Popotosa Formation (fig. 3) may have blocked the eastward migration of heated ground waters, thus forcing them to the surface. Alternatively, these hot waters may be leaking upward along the range-bounding fault from a deep reservoir heated by a shallow magma body. The thermal springs lie directly on the transverse shear zone across which the rotation of tilted fault blocks changes from easterly to westerly and the sense of extension changes 180 degrees. These changes must be accompanied at depth by shearing and at shallow depths by a twisting motion. Beds which have a northerly strike, but opposing dip directions, to the north and south of the shear zone generally become parallel to the shear zone in crossing it. These changes in tilt, accompanied by breaking of strata in new directions, may provide escape paths for geothermal fluids leaking from reservoir rocks at depth. Whatever the heat source, the purity and relatively low temperature of the waters (Hall, 1963) issuing from the Socorro thermal springs indicates major dilution by ground water. The tritium studies of Holmes (1963) indicate a transit time of about 4.3 years between recharge in the eastern Magdalena Mountains and discharge at the Socorro springs. This relatively rapid flow also suggests that dilution by ground water has a major effect on any geothermal fluids leaking from depth.

Structural Controls of Fracture Permeability

All rocks in the Socorro geothermal area have abundant fractures and joints, except for the claystone interval of the Popotosa Formation and the younger, less-indurated sedimentary deposits. Primary cooling joints in lava flows and ash-flow tuffs provide excellent permeability at right angles to bedding. The interconnection of these joints provides good permeability parallel to bedding. The long history of faulting has superimposed abundant tectonic fractures on the primary joint patterns. Thus, fracture permeability should be good in all of the brittle volcanic rocks and well-indurated sedimentary rocks. Fracture permeability is probably highest in the brittle, densely welded tuffs, such as the A-L Peak Tuff and the tuff of Lemitar Mountains. The distribution of potassium enrichment in reservoir rocks of an ancient geothermal system indicates that the well-jointed and fractured ash-flow tuffs did have the greatest permeability in that system.

STRATIGRAPHIC CONTROLS

Possible stratigraphic controls of the Socorro geothermal area are the sequence, thickness, facies changes, and lateral and vertical permeability of geologic forma-

tions which comprise the rock column in the area. For purposes of discussion, the stratigraphic units may be divided into: 1) Paleozoic rocks, 2) volcanic and volcanoclastic rocks older than the Socorro cauldron, 3) rocks of the Socorro cauldron and its moat, 4) volcanic rocks younger than the moat deposits, 5) sedimentary rocks of the Santa Fe Group (alluvial fill of the Rio Grande rift).

Paleozoic Rocks

Paleozoic rocks (fig. 3) consist of the Caloso Formation and Kelly Limestone of Mississippian age (Armstrong, 1962) and the Sandia Formation and Madera Limestone of Pennsylvanian age (Kottlowski, 1960; Siemers, 1978). These rocks vary from 1050 to 2270 feet (320-690 m) in thickness in the Socorro area because of varying degrees of removal during Eocene erosion of the Laramide uplift and, to a minor degree, because of differences in the thicknesses deposited. The Kelly and Madera limestones are well-jointed, relatively brittle rocks which have good fracture permeability at shallow depths. The new municipal well for the town of Magdalena produces 100 gallons/minute from fractured Paleozoic limestones along a major fault about 1.5 miles (2.4 km) northeast of Magdalena. The Kelly Limestone was highly permeable to hydrothermal fluids during stock intrusion and ore deposition in the Kelly mining district. The Sandia Formation, however, is largely shale and formed a very effective impermeable cap above the Kelly Limestone during circulation of hydrothermal fluids. Both the Kelly and Madera limestones could provide deep reservoir rocks within the Socorro geothermal system. The Sandia shales would provide an impermeable cap above the Kelly and the volcanoclastic rocks of the Spears Formation would provide a relatively impermeable cap above the Madera Limestone.

The Paleozoic rocks have been downfaulted to depths of 4,000 to 12,000 feet (1220-3660 m) within the Socorro cauldron and possibly to depths as great as 17,000 feet (5180 m) where the Socorro graben overlaps the eastern edge of the Socorro cauldron. The reservoir characteristics of the Kelly and Madera limestones at these depths and temperatures are unknown. However, they provide potential reservoir rocks relatively close to the tops of the shallow magma bodies delineated by Sanford and others (pl. 1, map 4) and may be worth testing if shallower reservoirs prove inadequate.

Pre-cauldron Volcanic and Volcanoclastic Rocks

Volcanic and volcanoclastic rocks older than the Socorro cauldron are, in ascending order, the Spears Formation, Hells Mesa Tuff, A-L Peak Tuff, and unit of Sixmile Canyon (fig. 3). The Spears Formation consists mainly of volcanic conglomerates with thin interbeds of volcanoclastic sandstones. Mudflow deposits become increasingly abundant upward in the Spears; ash-flow tuffs and basaltic andesite flows are common near the top. The muddy matrix and high degree of induration of the sedimentary rocks makes them relatively impermeable. The poor permeability and great thickness (0 to 1500 ft, 0-457 m) of the Spears Formation makes it a possible cap rock for the underlying Madera Limestone.

Outflow facies Hells Mesa Tuff consists of 0 to 500 feet (0-152 m) of densely welded, massive rhyolitic ash-flow tuffs; the cauldron facies is as much as 3,000 feet (915 m) thick. The brittle character and good jointing of these rocks, together with their lateral continuity and substantial thickness suggests good potential as a reservoir. The Hells Mesa is often separated from the overlying A-L Peak Tuff by 0 to 300 feet (0-90 m) of basaltic-andesite flow rocks. These flows are probably too thin and permeable to provide an effective cap. The overlying A-L Peak Tuff consists of 0 to 700 feet (0-213 m) of moderately to densely welded, rhyolitic ash-flow tuffs. A thin tongue of basaltic andesite flows of the La Jara Peak Formation generally separates the A-L Peak into an upper and lower cooling unit. The tuffs are generally well-jointed and brittle and should have good fracture permeability. Anomalously high K_2O values in the flow-banded and pinnacles members of the A-L Peak in the Lemitar Mountains (table 1) indicate that these units provided reservoir rocks for an ancient geothermal system.

The A-L Peak Tuff interfingers to the northwest with basaltic-andesite lavas of the La Jara Peak Formation. In the Lemitar Mountains, the A-L Peak is overlain by 0 to 500 feet (0-152 m) of basaltic-andesite flows. These lavas accumulated on downthrown blocks during early rift faulting; on upthrown blocks they are generally thinner and may be missing entirely so that the tuff of Lemitar Mountains rests directly on A-L Peak Tuff. In the Magdalena Mountains, mafic lavas vary from 0 to 600 feet (0-183 m) on the outflow sheet and to as much as 2,000 feet (610 m) within the Sawmill Canyon cauldron. The basaltic-andesite flows are well jointed but not as brittle as the densely welded tuffs. Their lateral and vertical permeability is probably too good to be an effective cap rock but relatively poor for a reservoir rock. However, these mafic flows are receptive to deposition of calcium carbonate and silica and they may provide a cap rock in areas of secondary cementation.

The Socorro cauldron overlapped and buried the eastern half of the Sawmill Canyon cauldron from which the pinnacles member of the A-L Peak Tuff was erupted. At least 2,000 feet (610 m) of densely welded A-L Peak Tuff accumulated within the Sawmill Canyon cauldron. Where exposed in the Magdalena Range, this tuff is pervasively jointed with closely spaced (5 to 15 cm) sheet joints that are approximately perpendicular to the foliation. This thick puddle of A-L Peak Tuff may provide one of the best reservoir rocks in the area. A drill hole near the Tower mine in the northern Chupadera Range penetrated about 300 feet (92 m) of hydrothermally altered and pervasively sheared A-L Peak Tuff and was bottomed in it.

Overlying the A-L Peak Tuff in the Sawmill Canyon cauldron is a thick series of cauldron-fill or moat deposits consisting of andesitic flows, talus breccias, sedimentary rocks, rhyolitic domes and flows, and local ash-flow tuffs. These deposits have been termed the unit of Sixmile Canyon (Osburn, 1978) for good exposures in the amphitheater at the head of this canyon. The unit of Sixmile Canyon ranges from 0 to 2000 feet (0-610 m) in thickness with andesitic lavas comprising most of the thickness at the type locality. A drill hole near the Tower mine in the northern Chupadera Mountains in-

tercepted a 213-foot (65-m) interval of rhyolitic to andesitic sandstones and conglomerates above the A-L Peak and below the tuff of Lemitar Mountains. Both the stratigraphic position and the rock types present in this interval are similar to the unit of Sixmile Canyon. If this correlation is valid, then the eastern edge of the Sawmill Canyon cauldron may extend as far east as the Tower mine. The area of overlap of the Sawmill Canyon and Socorro cauldrons seems to have acted as a buoyant, resurgent block following collapse of the Sawmill Canyon cauldron and during collapse of the Socorro cauldron. Both the moat deposits of the Sawmill Canyon cauldron (unit of Sixmile Canyon) and the caldera-facies tuff of Lemitar Mountains of the Socorro cauldron are anomalously thin on this block. Structural and stratigraphic traps may be present in tilted formations along the eastern and northern boundaries of the resurgent block; structural closure may exist on the resurgent block for reservoir rocks from the tuff of Lemitar Mountains through Paleozoic rocks.

Rocks of the Socorro Cauldron and Its Moat

Collapse of the Socorro cauldron occurred during, and as a result of, eruption of the tuff of Lemitar Mountains. The lower crystal-poor member of the tuff of Lemitar Mountains accumulated to thicknesses in excess of 800 feet (244 m) within the cauldron (the base of the lower member is rarely exposed within the cauldron so the maximum thickness may be much greater). Outside the cauldron, the lower tuff of Lemitar Mountains is discontinuous and thin (0 to 100 feet, 0-30 m). The upper, crystal-rich member is also thicker within the cauldron (350 to 3,000 feet, 107-884 m), but its outflow facies is relatively thick (0 to 400 feet, 0-122 m) and areally extensive. Anomalously high K_2O values and altered plagioclase feldspar within both the upper and lower members, both inside and outside the cauldron, indicate that the tuff of Lemitar Mountains may have been an important reservoir rock during an ancient geothermal system. Both members are densely welded and brittle with well-developed jointing and good fracture permeability.

During and after eruption of the tuff of Lemitar Mountains, a north-trending zone of differential subsidence existed in what is now the northern Chupadera Range. At the Tower mine in the western Chupadera Mountains, 690 feet (210 m) of the tuff of Lemitar Mountains have been intercepted in a drill hole; whereas, about 0.6 mile (1 km) to the southeast, a minimum of 2900 feet (884 m) of the upper member of the tuff of Lemitar Mountains is exposed in continuous section on the resurgent dome. Elevation of the resurgent dome formed a structural moat between its northern and eastern sides and the outer, topographic wall of the Socorro cauldron. Within this moat accumulated as much as 2,000 feet (610 m) of landslide deposits, local lithic-rich ash-flow tuffs, and andesite flows capped by rhyolite domes and tuffs. Collectively, these moat deposits have been termed the unit of Luis Lopez (Chamberlin, 1978). Their reservoir characteristics are quite varied but, in general, they form a relatively permeable unit with abundant joints, fractures and unconformities. The occurrence of the Socorro thermal

springs where the frontal fault of the Socorro Range cuts the moat deposits may be partly related to this permeability.

Volcanic Rocks Younger than Moat Deposits

The outflow sheet of another major ash-flow tuff unit was emplaced across the Socorro cauldron and surrounding areas about 26 m.y. ago, shortly after the cauldron had filled. This tuff, informally termed the tuff of South Canyon (Osburn, 1978), is a multiple-flow simple cooling unit generally ranging from 0 to 600 feet (0-183 m) in thickness. The tuff of South Canyon is so high in the stratigraphic column that it probably has no chance of being a reservoir rock except where down-faulted beneath the Socorro or La Jencia grabens. In these basins, it may occur at depths of 1,000 to 5,000 feet (305-1524 m) and, because of the high thermal gradient, could have reservoir potential. The lower, poorly welded zone and the overlying lithophysal zone are moderately porous and total about 180 feet (55 m) in thickness at the type locality. The upper 440 feet (134 m) of the tuff at the type locality is moderately to densely welded and brittle, with well-developed jointing and good fracture permeability. The tuff of South Canyon was a reservoir rock in an ancient geothermal system as evidenced by anomalous K_2O values and highly altered plagioclase phenocrysts.

The tuff of South Canyon is generally separated from the underlying Lemitar Tuff by basaltic-andesite flows which range in thickness from about 100 feet (30 m) in the Socorro cauldron to as much as 800 feet (244 m) in the Lemitar Mountains. This basaltic-andesite interval is probably a tongue of La Jara Peak Basaltic Andesite; the flows are also very similar to basaltic andesites interbedded in the A-L Peak Tuff and between the A-L Peak and the tuff of Lemitar Mountains. A typical basaltic-andesite flow is about 15 feet thick with autobrecciated zones above and below a thin, massive core. Thin andesitic sandstones and conglomerates are occasionally interbedded between flows. The entire basaltic andesite interval initially had high lateral and vertical permeability but much of the primary permeability has been lost by deposition of calcite and silica in void spaces. The basaltic andesites are not as brittle as the densely welded tuffs and are probably now a relatively poor reservoir rock.

The basal fanglomerates and mud-flow deposits of the rift-fill sediments (lower Popotosa Formation) rest directly on the tuff of South Canyon in the Lemitar Mountains, on Socorro Peak, and in the northern Chupadera Mountains. In the Water Canyon Mesa area of the eastern Magdalena Range, rhyolitic domes and flows, local rhyolitic tuffs, and a few intermediate lavas separate the tuff of South Canyon from the overlying Popotosa Formation. However, this seems to be a local occurrence.

The most extensive volcanic rocks younger than the moat deposits of the Socorro cauldron are the silicic flows and domes of Socorro Peak, Strawberry Peak, and Pound Ranch. These rocks were erupted between 12 and 7 m.y. ago and locally rest on top of all but the very youngest Popotosa Formation; however, they also lap unconformably onto the lower Popotosa Formation

and tuff of South Canyon in the Blue Canyon, Tower mine, and Pound Ranch areas. They are as much as 800 feet (244 m) thick but are so high in the stratigraphic column that they have virtually no chance of being reservoir rocks. They were not reservoir rocks in the ancient geothermal system as evidenced by their normal K_2O values and relatively fresh plagioclase. They overlie the thick playa claystone section of the Popotosa Formation which may have provided an impermeable cap for the ancient geothermal system. These claystones would locally provide a similar cap to a modern geothermal system so that, even where downfaulted beneath the Socorro and La Jencia grabens, the late Miocene silicic domes and flows would probably be above the geothermal systems.

Sedimentary Rocks of the Santa Fe Group

Following eruption of the tuff of South Canyon, the Socorro area subsided as part of a broad, early-rift basin in which as much as 5,000 feet (1524 m) of sediments of the Popotosa Formation accumulated. The sediments were deposited as coalescing alluvial fans which graded down broad piedmont slopes to playa lakes on the floor of the basin. The present site of the Lemitar-Socorro-Chupadera uplift was a persistent north-trending low on the basin floor. At least 800 feet (244 m) of red, gypsiferous clays accumulated along this axis. When volcanism again became active in the Socorro area about 12 m.y. ago, the silicic magmas intruded through the basin-fill alluvium and spilled out onto the playa deposits. The silicic lavas also lapped onto older bedrock along the southern margin of the basin in the Pound Ranch-Tower mine area. An unconformity which locally separates the lower and upper members of the Popotosa Formation is a reflection of this basin margin. Sedimentation continued during the silicic volcanism; consequently stratigraphic relationships are complex between contemporaneous volcanic and sedimentary rocks in the uppermost Popotosa Formation. An ancient geothermal system may have been established during this interval with the playa claystones locally forming an impermeable cap.

The coarser sediments of the lower Popotosa Formation (mud-flow deposits, fanglomerates and sandstones) may have been cemented and partially oxidized during the ancient geothermal activity. They are for the most part anomalously well-indurated and red for rift sediments. East of Socorro Peak and in the central Lemitar Mountains, the red, well-indurated facies intertongues and grades abruptly to light-gray, moderately indurated conglomeratic sandstones with opposing paleocurrent directions. How these primary facies relationships control the apparent secondary coloration and cementation is as yet unknown. The red facies of the lower Popotosa is now hard and brittle with little primary permeability left. However, these rocks are moderately well-jointed and brittle enough to develop good fracture permeability in fault zones, as evidenced by the fact that Socorro Spring, Sedillo Spring, and an unnamed cold spring northeast of Strawberry Peak issue from this unit.

Sometime after emplacement of the 7 to 12 m.y. old silicic domes and flows, the Popotosa basin was disrupted by uplift of tilted fault blocks to form the

Lemitar-Socorro-Chupadera mountain chain. The Popotosa sedimentary rocks were tilted and beveled by pediment surfaces on top of which a new cycle of rift sedimentation began. The younger sediments were derived from the uplifted blocks and contain significant quantities of clasts of Paleozoic and Precambrian rocks. These upper Santa Fe sedimentary rocks are generally lighter in color, less indurated, and less deformed than rocks of the underlying Popotosa Formation. These younger rift sediments have been termed the Sierra Ladrones Formation by Machette (1977). Beginning sometime prior to eruption of the 4 m.y. old basalt of Sedillo Hill, the drainage was integrated to form an ancestral Rio Grande. The extensive fluvial sands deposited by this trunk stream interfinger with piedmont-slope deposits of the Sierra Ladrones Formation and were included by Machette as a fluvial facies of that formation. The ancestral Rio Grande sands form the major fresh-water aquifer of the Rio Grande Valley in the Socorro area.

AN ANCIENT GEOTHERMAL SYSTEM

Preliminary chemical analysis and microscopic examination of thin sections of volcanic rocks in the Socorro area has revealed a pervasive addition of potassium to several of the major ash-flow tuff sheets. Tuffs which normally contain 4 to 5% K_2O were found to contain 6 to 11.5% K_2O . Sodium was found to be

depleted in rocks enriched in potassium, thus suggesting an alkali exchange phenomenon. Rocks with moderately elevated K_2O values were found in thin section to have plagioclase feldspar partly replaced by potassium feldspar. In rocks with higher K_2O values, the plagioclase feldspar has been completely altered to a white, fine-grained aggregate of potassium-rich, "clay-like" material which has not yet been identified. Upon staining with sodium cobaltinitrite, everything in the thin sections, except for quartz phenocrysts, gave a yellow stain indicative of the presence of potassium. The pervasive potassium stain and the very high K_2O values indicate that, in addition to replacement of sodium by potassium, there is a net addition of potassium to some of these rocks. There also appears to be significant addition of iron to the rocks and subtle reddening of their color, which probably indicates an increase in the ferric-ferrous ratio. Table 1 compares relatively fresh and altered samples of the A-L Peak Tuff and the tuff of Lemitar Mountains. The tuff of Lemitar Mountains samples are all altered to some extent, but samples LM-6-3 and PR-1-77L are comparatively fresh. Note the dramatic increase in the K_2O/Na_2O ratio, increase in total alkalis, and increase in total iron in the altered samples. Yet none of these rocks appear altered in outcrop; the potassium metasomatism is a subtle, easily overlooked alteration.

In addition to the above chemical and mineralogical

Table 1. Comparison of Preliminary Chemical Analyses of Fresh and Potassium-Metasomatized Samples of the A-L Peak Tuff and Tuff of Lemitar Mountains.

Sample #	Location	Total Fe_2O_3	Na_2O	K_2O	K_2O/Na_2O	Total Alkalies
<i>A-L Peak Tuff</i>						
<i>Flow-banded Member</i>						
76-6-7	Lemitar Mts.	Altered 2.03	3.39	6.61	1.95	10.00
<i>Pinnacles Member</i>						
76-1-11	Lemitar Mts.	2.46	1.46	9.48	6.49	10.94
<i>Basal vitrophyre</i>						
72-12	San Mateo Mts.	Fresh 1.28	4.64	3.95	0.85	8.59
72-2	San Mateo Mts.	1.26	3.85	4.68	1.22	8.53
<i>Tuff of Lemitar Mountains</i>						
<i>Lower Member</i>						
LM-6-2	Lemitar Mts.	Altered 2.02	1.98	7.94	4.01	9.92
LM-6-2b	Lemitar Mts.	1.01	2.22	8.12	3.66	10.34
<i>Relatively Fresh</i>						
LM-6-3	Lemitar Mts.	2.42	2.63	5.80	2.21	8.43
<i>Upper Member</i>						
LM-6-5	Lemitar Mts.	Altered 3.11	2.75	10.63	3.87	13.38
LM-6-6b	Lemitar Mts.	2.78	1.79	9.55	5.34	11.34
LM-6-8	Lemitar Mts.	2.26	1.54	8.67	5.63	10.21
LM-6-8b	Lemitar Mts.	2.71	1.97	9.46	4.80	11.43
76-1-10	Lemitar Mts.	2.87	1.97	8.37	4.25	10.34
LM-6-8a	Lemitar Mts.	3.88	1.57	11.57	7.37	13.14
PR-1-77d	East Magdalena Mts.	2.70	3.65	8.45	2.32	12.10
<i>Relatively Fresh</i>						
PR-1-77L	East Magdalena Mts.	1.88	3.63	5.97	1.64	9.60

changes in some of the ash-flow tuff sheets, the sedimentary rocks of the Popotosa Formation are anomalously red and very well-indurated in the Socorro area. Reconnaissance of other graben-fill sedimentary rocks along the Rio Grande rift indicates that the normal color is buff to gray or cream and that the rocks are normally only slightly to moderately indurated. Only one other locality along the rift was found to have such red and highly indurated early-rift sedimentary rocks—the San Diego Mountain-Rincon area near Hatch, New Mexico. Detailed chemical studies have not been done in this area, but a chemical analysis of a rhyolite sill in the Robledo Mountains (Seager and Clemons, 1975, p. 11) gave 7.04% K_2O and only 0.68% Na_2O —values similar to those of potassium metasomatized rocks in the Socorro area.

The potassium anomaly in the Socorro area is very extensive both laterally and vertically. Our chemical coverage of the anomaly is still very incomplete, but thin-section examination of samples for plagioclase alteration and staining with sodium cobaltinitrite has helped to establish the general extent of the anomaly. It appears that the more permeable volcanic and sedimentary rock units older than, or interbedded with, the Popotosa claystones have been affected from South Baldy in the central Magdalena Range on the southwest to the Ladron Mountains on the north. Using Socorro Peak as a center, these two lines form radii of about 14 and 20 miles (22.5 and 32.2 km), respectively. The extent of the anomaly to the south is not known. In vertical section, ash-flow tuff units aggregating 1,300 to 3,000 feet (396–914 m) in thickness are known to be affected; a minimum of 800 feet (244 m) of Popotosa fan-glomerates are inferred to have been involved because of their anomalously red color and high degree of induration.

If we assume a rectangular block of altered rock 15 miles wide by 25 miles long $\times$ 1 mile thick ($24.1 \times 40.2 \times 1.6$ km), an average increase in K_2O content of 3 percent, and an average density of 2.7, the net addition of K_2O is approximately 11×10^9 tons. The only reasonable mechanism for deriving such a large quantity of K_2O seems to be the leaching of potassium from granitic rocks in the Precambrian basement. Orville (1963) demonstrated experimentally that if a temperature gradient and a pervasive vapor phase exists in a two-feldspar rock, alkali ions will diffuse through the vapor in a reciprocal transfer process that depletes the hotter rock in potassium and enriches the cooler rock. He pointed out that hot spring waters are generally depleted in K relative to Na and cited evidence by White (1955) and Fenner (1936) that potassium is replacing sodium in plagioclase in rocks of the Wairakei, New Zealand, and Yellowstone geothermal systems. Battey (1955) demonstrated that alkali-rich volcanic rocks (called keratophyres) in New Zealand were normal rhyolite flows originally, but during deep burial they were held for a long time at moderately elevated temperatures during which alkali exchange took place by diffusion in pore fluids. In these rocks, the total alkali content remained constant but some rocks became enriched in potassium while others nearby were enriched in sodium.

Orville (1963) also demonstrated that an increase of calcium in plagioclase of the two-feldspar-plus-vapor

assemblage affects the vapor-feldspar equilibrium by increasing the K/Na ratio in the vapor. He postulated that original inhomogeneities in the calcium content of rocks will result in compositional gradients with respect to alkalis in the vapor phase and hence to alkali transfer. Rocks originally rich in calcium will tend to be depleted in potassium while those originally poor in calcium will tend to be enriched. Preliminary data indicates that mafic flows interbedded with potassium-enriched ash-flow sheets in the Socorro area are enriched in sodium relative to potassium. The data is as yet too sketchy to present; however, sodium enrichment of these mafic flows would fit Orville's experimental results and would explain where some of the sodium goes after it is replaced by potassium in the tuffs. This would also fit Battey's observations that adjacent rock masses in the New Zealand keratophyres are enriched either in potassium or in sodium.

Ratté and Steven (1967) discovered a similar potassium anomaly in the Bachelor Mountain cauldron of the San Juan Mountains (see also Steven and Lipman, 1976, p. 20). The potassium-enriched Bachelor Mountain Rhyolite contains 6 to 11.4% K_2O compared to about 4.7% K_2O in the unaltered rock. As the K_2O content of these samples increases, the Na_2O content decreases. Ratté (personal commun., 1977) was able to partially outline the potassium anomaly by contouring the K_2O values of about 30 samples. The exposed portion of the anomaly is about 6 miles wide by 10 miles long (9.7×16.1 km). Much of the anomaly, however, has been downfaulted beneath the younger Creede cauldron. We are especially grateful to Jim Ratté for helping us to interpret the Socorro anomaly.

The potassium anomaly in the Socorro area is a very interesting discovery which may be quite useful in evaluating the present geothermal system. It indicates, first of all, that a very large geothermal system existed in the Socorro area in the past; probably in late Miocene time (about 7 to 12 m.y. ago) during intrusion of rhyolitic magmas along the northern margin of the Socorro cauldron, but possibly also in early Miocene time during moat volcanism. Geophysical data indicates that magmas are again being intruded in this area and that the heat flow is very high. An ion exchange process is currently taking place in waters emerging from the Socorro thermal springs; waters in the recharge area are calcium bicarbonate types whereas those discharging at the springs are sodium bicarbonate types (Hall, 1963). Secondly, the distribution of potassium metasomatized rocks in the stratigraphic column can be used to determine which formations were the reservoir rocks of the ancient geothermal system and, by analogy, which formations may be the best reservoir rocks in the present system. And, finally, the lack of potassium enrichment in rocks above the Popotosa claystone horizon suggests that the claystones probably provided an impermeable caprock for a late Miocene geothermal system and are likely to do so again.

MODERN MAGMA BODIES

Two types of magma bodies have been detected in the vicinity of Socorro. The first is an extensive body, shown in red on Map 4 (pl. 1), which covers a minimum area of 1700 km². Depths to this body range from 18 km

to 22 km and the evidence to date suggests it has a thin, sill-like shape.

The second type of magma body is small intrusives located in the regions shown in yellow on Map 4 (pl. 1). These anomalous crustal segments appear to extend from the deep magma layer to within about 4 km of the surface.

Extensive Mid-Crustal Magma Body

The initial evidence for an extensive magma body at mid-crustal depths came from an analysis of the arrival times and amplitudes of two reflection phases on microearthquake seismograms (Sanford and Long, 1965; Sanford, Alptekin and Topozada, 1973). These two reflections, S to P and S to S, are impulsive and have the same general frequency content as the direct S-phase. Recordings by instruments with a broad frequency response show that the S to S phase contains a wide range of frequencies—from 3 to 15 Hz (Rinehart, 1976). This observation indicates that the reflecting discontinuity is sharp and singular. In addition, the absolute strength of the reflections as well as the ratio of their amplitudes indicates that the discontinuity producing the reflections is underlain by magma (Sanford, Alptekin, and Topozada, 1973; Sanford, 1977a).

The geographical extent of the magma body is being determined from the presence and absence of reflections on the microearthquake seismograms (Sanford et al., 1977a; Sanford, 1977a). Beyond the boundaries of the magma body no S to S reflections have been observed to date although the instruments have been positioned to record such reflections. Northward, beyond the limits of Map 4 (pl. 1), the boundaries are not closely defined by the S-phase reflection data. However, good control on the extent of the magma body in this region has been obtained from high-resolution seismic reflection profiles (Oliver and Kaufman, 1976; Brown and others, 1977). These crustal profiles show very strong P-wave reflections from approximately the same depth as the S wave reflections. In addition, the P-wave reflections have a northward dip of about 6 degrees which is in agreement with the S-wave reflection data from the top of the magma layer.

The available observational data indicate that the magma body is very thin relative to its extent. P-wave residuals for teleseisms (Fischer, 1977) do not support the existence of a thick magma layer even after station corrections are applied. Analysis of the Gasbuggy refraction data (Topozada and Sanford, 1976) indicates that time delays for P_n arrivals passing through the magma layer cannot be much greater than 0.1 second. A layer of magma (full-melt) 0.6 km thick would be sufficient to produce a 0.1 second time delay. Finally, no clearly defined reflection phases from the bottom of the magma layer have been identified on microearthquake seismograms. If such reflections exist, they must occur so closely in time to the reflections from the top of the magma body that they cannot be easily identified.

Small Shallow Magma Bodies

Three types of observations suggest the existence of small magma bodies above the southern end of the mid-crustal magma layer: 1) the screening of SV waves, 2)

the spatial distribution of Poisson's ratio, and 3) the spatial distribution of microearthquake hypocenters.

(1) Screening of SV Waves. For many microearthquakes in the vicinity of Socorro, SV waves are absent or extremely weak on seismograms for one or more stations in the array (Shuleski, 1976; Sanford and others, 1977a). About 40 percent of these observations can be explained by the fault mechanism of the microearthquakes; the remaining 60 percent apparently result from SV screening by molten or partially molten rock bodies (Kubota and Berg, 1968; Matumoto, 1971). Magma bodies located within the regions shown in yellow on Map 4 (pl. 1) can explain the observed SV screening in the Socorro area.

(2) Spatial Distribution of Poisson's Ratio. From an analysis of S-P times and P travel times for microearthquakes, the spatial distribution of Poisson's ratio in the upper crust has been determined (Caravella, 1976). Several segments of the crust with average Poisson's ratios greater than 0.29 have been mapped in the Socorro region. These anomalously high values of Poisson's ratio can be explained by the same distribution of shallow magma bodies that is used to account for the SV screening.

(3) Spatial Distribution of Microearthquake Hypocenters. A detailed three-dimensional analysis of the spatial distribution of microearthquake foci shows activity surrounding, but not within, the anomalous crustal segments mapped in yellow on Map 4 (Shuleski and others, 1977; Sanford and others, 1977a). Above these regions, microearthquake hypocenters are never deeper than about 2½ km, whereas adjacent to those regions, activity occurs to depths of 8 km or greater. Earthquake activity is not expected in segments of the crust containing magma inasmuch as high temperatures prevent accumulation of elastic strain energy.

Configuration and Distribution of Shallow Magma Bodies

The techniques employed in detecting shallow magma bodies do not permit precise determination of the shape or volume of magma injected into the crust. However, an acceptable model at this time is injection of thin, discontinuous dikes of magma such that the fractional volume of molten material in the yellow regions of Map 4 is quite small.

To date we have only looked for shallow magma bodies above the southern end of the extensive magma body. Thus we cannot be certain this is the only region where magma leaks upward through the crust from the deeper magma body.

Other Geophysical Evidence for Magma Bodies

An independent geophysical observation suggesting magma bodies at shallow depths in the Socorro area is the high temperature gradients (maximum, 241°C/km) and heat-flows (maximum, 11.7 HFU) measured in boreholes within the Socorro Mountain block (Reiter and Smith, 1977; Sanford, 1977b). Values of measured temperature gradients and heat-flow are shown on Map 4. The low temperature gradients for stations in the Rio

Grande Valley are most likely the result of southward flow of groundwater within the basin (Bushman, 1963).

Two observations suggest that magma may have been injected into the crust in very recent times. The first is the discovery by Reillinger and Oliver (1976) of historical uplift in the Socorro area roughly coincident with the spatial extent of the extensive mid-crustal magma body. The second is the pattern of seismic activity in space and time (Sanford and others, 1977a). The microearthquake activity is diffusely distributed over an area of 2000 km² that is roughly centered on the extensive magma layer as presently mapped. Most earthquakes in the Socorro area have occurred in swarms, which is the characteristic seismic behavior for volcanic regions. One of the Socorro earthquake swarms (from 1906-1907) appears to have been comparable to the Matsushiro swarm (from 1965-1967) which Stuart and Johnston (1975) believe was caused by magmatic intrusion.

GEOHERMAL POTENTIAL OF THE SOCORRO AREA

High heat flow, the presence of shallow magma bodies, the existence of both reservoir rocks and cap rocks in the rock column, and the downfaulting of potential reservoir rocks to considerable depths makes the Socorro area an attractive target for geothermal exploration. Recognition of the transverse shear zone, its long history of "leaking" magmas, and the probable upwarp of the Curie temperature isotherm along it reinforces this conclusion. Exploration, however, must take into account the rapid flow of ground water from recharge areas in the Magdalena Mountains towards the Rio Grande. The travel path for most of this water is beneath the thick Popotosa claystone section; hence, temperature measurements made in shallow holes will not be an effective guide to temperatures at depth.

THE SOCORRO SHEAR ZONE—A POSSIBLE MODEL FOR GEOTHERMAL EXPLORATION ELSEWHERE ALONG THE RIO GRANDE RIFT

Recognition of the Socorro transverse shear zone and its role in controlling magmatism and geothermal activity in the Socorro area provides a possible model for exploration elsewhere along the Rio Grande rift. Several major northeast-trending, subparallel lineaments transect the rift (fig. 1); however, different lineaments have "leaked" magmas at different times. The Jemez lineament has been the most active lineament magmatically during the past 5 million years. The Kremmling lineament was active between 24 and 20 m.y. ago and 14 to 8 m.y. ago, with minor activity of Pleistocene and Holocene age (Larson and others, 1975). The Idaho Springs and Silverton lineaments were the most active of the northeast-trending lineaments during the Laramide orogeny (75-50 m.y. ago). The reasons for this behavior are as yet unknown. In fact, the subject of lineament tectonics and magmatism has been shrouded in mystery and controversy for decades and remains so today (see

Gilluly, 1976, 1977; Walker, 1977; Warner, 1978). Apparently, when a regional stress field is applied to a flawed plate, the stress is relieved along different flaws at different times. Why the flaws leak magmas is another, largely unresolved, question.

Basins of the Rio Grande rift are often hinged in opposite directions on opposite sides of transverse lineaments. For example, the upper Arkansas basin is hinged down from the east side with the deepest fill on the west; the San Luis basin is hinged down from the west with the deepest fill on the east. The transition between opposing structural styles occurs where the Salida lineament transects the rift. The Salida area is also the site of several major hot springs. A similar twist in basin geometry occurs across the Jemez lineament. To the north, the southern extension of the San Luis basin is hinged down from the west with the deepest fill to the east; to the south, the Espanola basin is hinged down from the east with the deepest fill to the west. The transition occurs across the Jemez lineament which is marked in this area by the Jemez volcanic field and present day geothermal activity.

Transverse horsts often jut into basins where lineaments cross the Rio Grande rift. Examples are the Browns Canyon horst and the northern tip of the Sangre de Cristo Range in the Salida area, the Picuris Range in the Taos area, and the Mud Springs Mountains near Truth or Consequences. Hot springs are often present near these horsts. It is also common for one side of a lineament to be elevated relative to the other. In the Socorro area, the La Jencia and Mulligan Gulch grabens shallow southward across the Morenci lineament. Recurring uplift of the south side is reflected in unconformities and facies changes in the basin-fill sediments and in the geomorphology. However, the north side is now being uplifted (Reillinger and Oliver, 1976) because of inflation and thermal expansion related to intrusion of a sill-like magma body at 18 to 22 km depth. In the Espanola basin, the north side of the Jemez lineament seems to have been persistently high. In the upper Arkansas basin, the south side of the Salida lineament has been recurrently high.

Where lineaments cross the Rio Grande rift, deeply penetrating flaws in the continental plate produce a characteristic suite of structural features which are often reflected in the stratigraphy and geomorphology. The direction of tilt of beds and the direction of downthrow across faults may change 180 degrees. Basins may be hinged on the east or on the west on opposite sides of the lineament. Transverse horsts may jut into the basins and one side or the other of the lineament may be persistently up or down relative to the other. Magma injection, volcanism, high heat flow, and geothermal activity are often associated with these areas of transverse tectonics. Recognition of these features may help narrow the search for geothermal exploration targets along the rift. Comparison of geologic and topographic maps with aeromagnetic maps may locate other "quiet" zones where the Curie temperature is at anomalously shallow depths.

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APPENDIX I

A STUDY OF POISSON'S RATIO
IN THE UPPER CRUST IN THE
SOCORRO, NEW MEXICO, AREA

by

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Table of Contents

Acknowledgements	11
Abstract	111
I. Introduction	1
II. Geologic Setting	3
III. Previous Studies of Poisson's Ratio in the Rio Grande Rift	10
IV. Instrumentation	13
V. Data Collection, Reduction, and Analysis	19
VI. Results and Discussion	41
VII. Conclusions	54
References Consulted	55
Appendix A: Wadati diagrams which yielded unreasonably low values for Poisson's ratio	
Appendix B: Discussion of spatial distribution from crustal wedges of low Poisson's ratio and the 167 crustal wedges found from the analysis	
Appendix C: Microearthquake data used in this study	

Abstract

Arrays of 5 to 9 portable, high-gain, short-period seismographs were used to record microearthquakes from April, 1975, through January, 1978, in the vicinity of Socorro, New Mexico. Body wave arrivals from 294 microearthquakes were used to calculate an average value of Poisson's ratio for the upper crust (<15 km). Hypocenter locations were obtained employing a numerical inversion computer program, and the results used to obtain a spatial distribution of Poisson's ratio for the Socorro region. A composite S-P time versus travel-time plot was constructed to determine the correlation among the data.

A weighted least-squares linear regression was performed for 277 Wadati diagrams (plot of S-P time interval versus arrival-time), and the results used to obtain a value of 0.251 ± 0.052 for the average Poisson's ratio in the Socorro area.

The spatial distribution found for the Socorro region revealed 4 areas of anomalously high Poisson's ratio. They are: (1) the southern margin of the La Jencia basin, (2) east-central La Jencia basin, (3) the northern tip of the Chupadera Mountains, and (4) east-central Socorro basin. These regions were of particular interest because they may indicate the presence of shallow partial melts which could serve as a source of geothermal energy.

The results of the composite plot, involving over one-thousand points, showed excellent correlation among the data, yielding a correlation coefficient of 0.91.

I. Introduction

The purpose of this study was to determine an average Poisson's ratio in the upper crust (<15 km) for the local Socorro region using body wave arrivals from 294 located microearthquakes, and to obtain a spatial distribution of Poisson's ratio using information gleaned from the microearthquake locations. The determination and location of areas with anomalously high Poisson's ratio was of particular interest in this study as these segments of the crust may contain partially molten material.

Poisson's ratio (ν) is a dimensionless quantity representing the ratio of the fractional change in length to the fractional change in diameter in a cylindrical volume of material subjected to a tensional or compressional force, such as the transmission of seismic energy through rock. Since core samples of rock for the entire area of investigation were unavailable, an alternate means for obtaining Poisson's ratio was necessary. The method employed in this study used the arrival times of the P (compressional wave) and S (shear wave) phases and the S-P time intervals (see Figure 1). Note that the plot of S-P intervals versus P arrival times gives a straight line with a slope of $(\frac{\alpha}{\beta}-1)$ which appears in the relation:

$$S-P = (P-O)(\frac{\alpha}{\beta}-1), \quad (1)$$

where

- S = S-wave arrival time,
- P = P-wave arrival time,
- O = Origin time of the microearthquake,
- α = P-wave velocity, and
- β = S-wave velocity.

Equation 1 may be derived from the definition of the S-P interval, given by:

$$S-P = \frac{D}{\beta} - \frac{D}{\alpha}, \quad (1a)$$

where D = distance from the earthquake hypocenter to a receiving seismograph station, and may be written as: $D = (P-O)\alpha$.

Poisson's ratio may then be calculated using the relation:

$$\nu = \frac{\left(\frac{\alpha^2}{\beta^2} - 2\right)}{2\left(\frac{\alpha^2}{\beta^2} - 1\right)}, \quad (2)$$

given by Bullen (1963, page 213). The value of Poisson's ratio may vary between 0 and $\frac{1}{2}$ for different materials but usually is in the neighborhood of $\frac{1}{4}$ (Nettleton, 1940). The cases where ν is 0 and $\frac{1}{2}$ correspond to a perfectly rigid solid and a perfect liquid, respectively. Poisson's ratio thus provides a measure of the general characteristics of a material, which may be utilized to detect the shallow (<10 km) magma bodies in the Socorro area proposed by Sanford and others (1977), Shuleski (1977), Caravella (1976), Johnston (1978), and Wallace (1978).

II. Geologic Setting

The area of study lies within the Rio Grande rift, in central New Mexico, approximately 120 km south-southwest of Albuquerque. The Rio Grande rift is a major extensional structure formed by east-west tension beginning about 25 to 29 m.y. ago and continuing to the present (Chapin and Seager, 1975). The rift is composed of a series of en echelon structural depressions having raised margins and a general northern trend. The rift extends from southern New Mexico into central Colorado, and penetrates the southern Rocky Mountains of northern New Mexico and southern Colorado. The rift lies between the Colorado Plateau and Great Plains in central New Mexico, and merges to the south in a complicated manner with the Basin and Range Province (see Figure 2). The area of investigation encompasses a portion of the Rio Grande rift which has been further modified by intergraben horsts believed to be 9 to 10 m.y. old, suggesting relatively recent crustal structure activity (Chapin and Seager, 1975). Figure 3 shows the general physiographic features of the area of study, where intergraben horsts appear as the Socorro-Lemitar and Chupadera mountains separating the La Jencia and Socorro basins.

A large amount of evidence indicates that volcanic activity accompanied formation of the rift. Several overlapping calderas have been mapped in the Socorro area and are believed

Figure 1. Example of microearthquake used in this study. This event was recorded at seismic station CC on February 19, 1976, and shows the P and S wave arrivals as well as two reflection phases believed to be associated with a deep magma body whose upper surface lies between 18 and 20 km below the area of study. The S-P interval is obtained by subtracting the P arrival time from the S arrival time.

to be the source of voluminous ashflow eruptions which deposited the widespread tuff sheets that cap the Datil-Mogollon volcanic pile. K-Ar dates on these volcanic rocks indicate periods of magmatism at 27-20 m.y., 12-7 m.y. and 4 m.y. ago (Chapin et al., 1978). Early volcanism had basaltic andesites

Figure 2. Physiographic provinces in New Mexico. The Rio Grande Rift is shaded (after Topozada, 1974).

Figure 3. Physiographic features of the Socorro area, seismic station locations, and outline of the deep (18 km) magma body (from Rinehart, 1978).

Figure 4. Map showing the epicentral distribution of the microearthquakes used in this study. The 'x' marks represent individual events.

associated with it, while true basalts dominated the flows of very recent age in the Socorro area.

Evidence supporting current activity in and around the rift includes a ribbon of high heat flow (>2.5 HFU) which is roughly coincident with thermal springs occurring along the western margin of the rift (Reiter et al., 1975), a high rate of surface uplift, with a maximum average rate of 6.1 mm per year, approximately 23 km north of Socorro (Reilinger and Oliver, 1976), high microearthquake activity and microearthquakes occurring in swarms. (The epicentral distribution of microearthquakes used in this study is shown in Figure 4.) The hottest and largest thermal springs are located in the southern end of the Socorro-Lemitar mountains where high temperature gradients (maximum, $241^{\circ}\text{C}/\text{km}$) and heat flows (maximum, 11.7 HFU) have been obtained from boreholes within the Socorro Mountain block (Reiter and Smith, 1977).

An extensive magma body whose upper surface is located between 18 and 20 km in depth has been defined using reflection phases appearing in many microearthquake codas (Rinehart, 1976). The outline of the deep magma body is shown in Figure 3, where dotted lines indicate boundaries where more data are required to map them accurately. The body appears to be thin (approximately 1 to 2 km in vertical thickness) and elongate, and has a general north-south trend (Sanford et al., 1977). The upper surface of the magma body is relatively flat with a slight northward dip of two to three degrees (Rinehart, 1978). The

possibility that this deep magma body serves as a source of the proposed shallower magma bodies is at present unresolved.

III. Previous Studies of Poisson's Ratio in the Rio Grande Rift

A study of Poisson's ratio (ν) was conducted in the Rio Grande rift, between Albuquerque and Socorro, New Mexico by Kiet Sakdejayont (1974). The ratio of P-wave velocity to S-wave velocity was examined using the data obtained from 32 well-recorded microearthquakes within 45 km of the Socorro seismograph station, SNM. The values of α/β and ν were found to be 1.664 and 0.217 respectively. These comparatively low values were considered to be normal values for the Rio Grande rift by Sakdejayont.

A more recent paper involving the calculation and spatial distribution of Poisson's ratio in the Socorro, New Mexico area was that of Caravella (1976). In this study, data from 50 microearthquakes located in and around the southern margins of the Socorro and La Jencia basins were used. Caravella found an average ν of 0.262 with a standard deviation of 0.034 using a method similar to that employed in this paper. This average Poisson's ratio is about 21% higher than the value obtained by Sakdejayont for crustal rock further north in the rift. One notes however that two standard deviations applied to Caravella's value of ν would include the result obtained by Sakdejayont, and one standard deviation applied to both would probably produce a non-null union (Sakdejayont did not state a standard deviation for his value of ν .). Caravella attributed the possible difference in Poisson's

Figure 5. The three areas of anomalous Poisson's ratio proposed by Caravella (1976), obtained using crustal wedge intersections. The areas are: (1) the southern La Jencia basin ($\nu = 0.292$), (2) Socorro Mountain ($\nu = 0.289$), and (3) central La Jencia basin ($\nu = 0.284$).

to a difference in the S-wave velocity for the two areas, 3.30 km/sec. versus 3.49 km/sec. Caravella attempted to determine spatial variations in ν , but found that more data were necessary to resolve such variations in a significant fraction of his study area. His data were sufficient to define three anomalous areas (see Figure 5) having average ν 's of 0.292, 0.289 and 0.284. They were: (1) southern La Jencia basin, (2) Socorro Mountain, and (3) central La Jencia basin respectively.

IV. Instrumentation

Microearthquake data used in this study were obtained using two types of portable seismic systems. Throughout the data collection period (April, 1975 to January, 1978) an array of 4 to 6 Sprengnether Instrument Company MEQ-800 analog recording units was deployed. Each MEQ-800 unit consisted of either a Mark Products L4C or Willmore vertical seismometer having natural frequencies of 1.0 and 1.5 Hz respectively, a gain-stable amplifier, a quartz-crystal-controlled timing system, and a smoked-paper helical recorder which operated at a recording speed of 120 mm/min. This basic array was supplemented, starting in April, 1977, with two Sprengnether DR-100 digital recording units equipped with Mark Products L4C vertical seismometers. In addition to these units, a Kinematics PS-1 analog unit, with a recording speed of 150 mm/min., was used to record station LAD which was telemetered via telephone line directly to the campus of New Mexico Institute of Mining and Technology. The MEQ-800 seismographs served as the fundamental source for data, while the PS-1 unit and DR-100 systems were primarily used to improve hypocenter locations by expanding the existing arrays.

During the data collection period, many different arrays were used, and in all, 26 different station locations were occupied (see Figure 3). Table 1 lists these seismograph stations along with their locations and elevations. Elimin-

Figure 6. Magnification response curves for the MEQ-800 seismographs. The high-frequency filter is set in the 30 Hz position (adapted from Rinehart, 1976).

Figure 7. Magnification response curves for the MEQ-800 seismographs. The high-frequency filter is set in the Out position (from Shuleski, 1976).

Table 1. Location of Seismic Stations

<u>Station</u>	<u>Latitude</u>	<u>Longitude</u>	<u>Elevation (m)</u>
BB	34.4090	106.6818	1615
BG	34.2068	106.8205	1516
CC	34.1442	106.9812	1649
CK	34.2725	106.7702	1578
CM	33.9501	106.9576	1640
CU	34.1573	106.7785	1585
DM	34.1075	106.8079	1536
FC	34.8950	107.0504	1850
FM	34.0829	106.8047	1537
FR	33.8745	106.7270	1558
GM	34.1454	107.2345	1945
HC	34.0658	107.2361	2240
IC	33.9870	106.9967	1730
LAD	34.4583	107.0375	1768
LPN	34.3076	106.6336	1737
MG	34.1305	107.2425	2024
MY	34.1667	106.7459	1645
NG	33.9648	106.9933	1730
NJ	33.9924	106.6253	1644
RI	34.4234	107.2075	1530
SC	34.0100	107.0894	2073
SL	34.2234	106.9910	1615
TA	34.0498	106.7751	1558

Table 1. (continued)

<u>Station</u>	<u>Latitude</u>	<u>Longitude</u>	<u>Elevation (m)</u>
TD	34.2339	106.5778	1850
WM	34.0120	106.9929	1673
WT	34.0722	106.9459	1555

ation of atmospheric and man-made noise were two prerequisites for station site selection. All seismograph stations selected rested upon bedrock. In most cases, caves and abandoned mines were used and proved quite satisfactory, but when high background noise levels were encountered, lower gain settings and/or filtering were employed to increase signal to noise ratios.

The MEQ-800 and PS-1 units have gain settings ranging from 60 to 120 db in discrete 6 db increments. Some filtering of the seismic signal was possible for both low frequencies (below 5 or 10 Hz) and high frequencies (above 5, 10, or 30 Hz). For this study, the 30 Hz setting on the high frequency filter was used over 50% of the time and nearly 100% of the time for the noisiest stations (BG, CU, FC, HC, SL, TA, MY). The filtering was necessary to reduce noise generated by wind and small animals. The microearthquake signals were not impaired greatly by this filtering since the MEQ-800 unit still showed good response at that setting for frequencies around

30 Hz, the dominant frequency of the P arrival (Sanford and Holmes, 1962). The DR-100 systems have gain settings ranging from 78 to 96 db in discrete 6 db intervals with both high and low frequency filtering available. In general, the highest possible gain setting with least filtering was used. For the MEQ-800 systems, this resulted in most of the gain settings falling between 78 and 90 db for the data used in this study. Magnification versus frequency plots are shown in Figures 6 and 7 for the MEQ-800 seismographs, for both types of geophones used, at typical gain and filter settings.

V. Data Collection, Reduction, and Analysis

Section 1

Obtaining accurate times for both the first arrival and S phase was of fundamental importance to this study. Before each recording period (4-6 days in duration), the clocks on all seismographs were either synchronized to WWV-UST (broadcast from Fort Collins, Colorado) or the time difference was noted. Similarly, at the end of the recording period, the clocks were checked for any drift which may have occurred. Any clock drift was assumed to have occurred linearly and timing corrections were made accordingly. Several tests of the linearity of clock drift assumption were made, both in the field and inside a single building or mine where the temperature was essentially constant. Two to three drift measurements were made on each seismograph unit at approximately 8 hour intervals so that any deviation of the clock drift from linearity might be better defined. The tests showed that the maximum divergence from linearity was less than 5%, which would result in a time correction error of less than 0.03 seconds.

The PS-1 and MEQ-800 seismograms were read with a Gaertner traveling microscope, which has a measuring accuracy of approximately 0.005 mm (~0.003 seconds). Tests show that different individuals can read the same record with an accuracy of about 0.04 mm (~0.02 seconds) (Rinehart, personal

communication, 1978). The DR-100 cassette recordings were usually played back on a strip-chart recorder at a rate of ~60 mm per second, allowing a very accurate measurement to be made with the use of a millimeter scale.

The data used in this study consist of three basic measurements: (1) the P arrival times, (2) the S-P intervals, and (3) the distance between minute marks on either side of the onset of an event, all measured in millimeters (see Appendix C). The adjacent distance between minute marks was used when the minute in which the microearthquake arrived was split when taken off the seismograph drum. Times were then calculated in seconds by dividing measurements 1 and 2 by measurement 3 and multiplying by 60. A similar procedure was used to obtain times from the DR-100 records. The P arrival times were then corrected using the clock drift corrections described above. Uncertainties in these times were estimated by repeating the measurements at least three times for each event used.

Section 2

Microearthquake locations were obtained for 294 microearthquakes using a numerical inversion computer program. The crustal model assumed in this iterative inversion program was an isotropic, homogeneous half-space with a P-wave vel-

ocity of 5.80 km/sec. A trial hypocenter was determined using: (1) chord intersections from a circle plot of the S-P intervals for the stations in the array, (2) a Wadati diagram (plot of the S-P intervals versus the P arrival times), and (3) an assumed depth of 7.0 km. The circle plot yielded trial values for the longitude and latitude for the hypocenter, while the Wadati diagram gave an initial estimate of the origin time for the event. The initial value of 7.0 km for the depth was used as it is the average depth of focus for earthquakes occurring in the Socorro area (Sanford, personal communication, 1978). The average depth found for the 294 microearthquakes located in this study was 6.7 km. The computer printouts of the final microearthquake hypocenters are included in an appendix separate from this paper, whereas the measurements used to obtain the locations may be found in Appendix C of this report.

In addition to the crustal parameters, P arrival times and trial hypocenter, the iterative inversion program utilized station corrections which corrected for local changes in velocity that exist in the rock column directly beneath the seismograph stations. The corrections, which varied from 0 to 0.48 seconds, were subtracted from the P arrival times at the onset of the location procedure. A list of the station corrections used in this study may be found in Table 2. An explanation of the two methods used to obtain these

Table 2. Station Corrections.

<u>Station</u>	<u>Correction</u>	<u>Method of Determination</u>
WT	0.00	<p>explosion</p> <p>iterative inversion</p>
CC	-0.11	
SL	-0.11	
GM	-0.11	
IC	-0.38	
RM	-0.44	
NG	-0.35	
CM	-0.42	
SC	-0.48	
DM	-0.11	
FM	-0.22	
BG	-0.22	
CU	-0.17	
WM	-0.32	
RI	-0.38	
MY	-0.12	
HC	-0.33	
FC	-0.47	
TS	-0.47	
TA	-0.15	
BB	-0.18	
CK	-0.22	

corrections is given by Johnston(1978).

The microearthquake location program utilized a weighted least-squares approach given by Aki (1976). The solutions obtained were rated using the following formula:

$$R = \left[\frac{1}{N} \sum_{i=1}^N \left(\frac{r_i}{w_i} \right)^2 \right]^{\frac{1}{2}}, \quad (3)$$

where N = number of stations used in location,
 r_i = travel-time residuals (difference between theoretical and actual travel times), and
 w_i = a priori uncertainties in the travel-time measurements (0.05 to 0.50 seconds in this study).

The value of R indicates how well the travel-time residuals are minimized within the weighting framework and, when R is close to 1.0, indicates that a solution which is both as good as possible and as good as is necessary has been obtained. (i.e. solutions having an R value close to 1.0 are optimal for the weights assigned the P arrival times for those solutions) Locations with $R \ll 1$ will result in uncertainties in the hypocentral parameters which are unnecessarily large, while solutions with $R \gg 1$ will result in solutions whose hypocentral parameters have unjustifiably small uncertainties. The weights assigned to the P arrival times were dependent upon the background noise levels and the character of P phases observed (impulsive or emergent) on the records. The weight generally assigned to the arrival times was 0.05 seconds, which was obtained by summing the reading and clock

drift correction errors mentioned previously. Of the 294 microearthquakes located, 251 had an R value lying between 0.50 and 1.50, while only 7 earthquakes had values of R greater than 2.50.

Since four unknowns are determined by the location program, at least four P arrival-times (i.e. four stations recording an event) are necessary to obtain a solution. This study includes 75 four-station locations, 121 five-station locations, 77 six-station locations, 17 seven-station locations, and 4 eight-station locations. Having more than four stations usually reduced the uncertainties in the hypocentral parameters. This was of particular importance to this study because the uncertainty in the iterative origin time was used in formulating the weights for the analysis of Poisson's ratio using travel-times.

Section 3

The determination of uncertainties for the measurements used in this study was of particular concern to the author. In order to handle the data collected in a statistical manner, a system for reading the S-P intervals was devised. The procedure consisted of three independent measurements of the S-P time, which was assigned a grade of E, G, or P prior to the first measurement. The criteria for assigning these three grades were as follows:

- E: Very sharp P and S phases. No noise obscuring either arrival.
- G: Some noise affecting the P and/or S phases, i.e. less well-defined arrivals but P and S picks are obvious.
- P: Any combination of noise, clipping and "white-out" (found only on the strongest microearthquakes), or emergent character of P and/or S phases, making S-P time subjective but not unreasonable.

Any S-P interval which was split when the record was taken of the seismograph drum was assigned a grade of P. Examples of each of the above types of S-P intervals encountered in this study are shown in Figures 8a, 8b, and 8c. The two basic properties of amplitude and frequency change in the seismogram trace were used predominantly to pick the S phase, the S-wave generally having the largest amplitude and lower frequency content. The mean and standard deviation were calculated for each trio of S-P measurements, which were then grouped according to their assigned grade. Forty-three percent of the S-P intervals measured were graded E, and were found to have an average standard deviation of 0.007 seconds. S-P intervals receiving a grade of G comprised about 36% of the readings and had an average standard deviation of 0.009 seconds. The remaining S-P measurements received a grade of P and had an average standard deviation of 0.012 seconds. These small average standard deviations appear to represent the

P S

Figure 8a. Example of microearthquake having an S-P interval that received a grade of E. Both the P and S phase exhibit sharp, impulsive onsets. Event was recorded at station WM on February 19, 1976 at 00:08.

inaccuracy introduced by the traveling microscope, which has been described previously. Although a small increase in the standard deviation was found to exist as the quality of the S-P interval decreased, the overall error associated with any S-P measurement is small when compared with the uncertainties in the P arrival time. This result indicates consistency in the picking of the P and S arrivals, but not necessarily the error associated with the S-P determination. (i.e. The ability to pick the same breaks on the seismogram

for the arrivals of the P and S waves does not reflect upon the correctness of the resulting S-P interval, as one might consistently pick the P and/or S arrival incorrectly.) For this reason, the uncertainties assigned to the S-P values were the sum of the P arrival-time uncertainty and the standard deviation found for the S-P time. These uncertainties were then used in the Wadati diagram analysis as the weights in the weighted least-squares linear regression.

Figure 8b. Microearthquake recorded at station SC on February 19, 1976 at 00:54. The S-P interval for this event received a grade of G. The P pick is obvious while the S arrival is somewhat subjective.

P 52

Figure 8c. Microearthquake recorded at station IC on March 23, 1976 at 12:53. The S phase pick is highly subjective for this event, and although the P phase may be easily found, the S-P interval may be subject to a large error if the incorrect S arrival is chosen. For this reason, the grade assigned to the S-P for this event was P.

The weights employed in the travel-time analysis were the sum of the uncertainties in the iterative origin time, S-P interval and the P arrival-time. (The sum of these uncertainties was used, as the measurement and determination of the P arrival, S-P interval, and iterative origin time are not independent.)

Section 4

The procedures used to calculate Poisson's ratio involved the use of the P arrival-times, S-P intervals, distances from seismic stations to earthquake epicenters, azimuths from stations to earthquake locations, and earthquake origin times. The P arrival-times and S-P intervals were obtained directly from the seismograph records, as outlined in Sections 1 and 2, while the remaining information was calculated by the location program discussed in Section 3. The origin-times and P arrival-times were used to determine the P travel-times ($P \text{ travel-time} = P \text{ arrival-time} - \text{iterative origin-time}$), which completed the data needed for this study.

The analysis used to obtain an average value of ν for the upper crust in the area of study utilized only the P arrival-times, S-P intervals and the uncertainties associated with their determination. Of the 294 microearthquakes located, 277 were deemed suitable for this analysis. The reasons for eliminating a microearthquake solution from the calculations were : (1) poor hypocentral determination caused by an inadequate seismic station distribution about the event and (2) an insufficient number (<4) of readable S-P intervals due to saturation and/or clipping of seismic signals and/or apparent absence of the S phase on a number of the seismograph records used to locate the event. A weighted least-squares linear regression was used for each of the

277 Wadati plots constructed. The resulting slopes found for the lines determined by this procedure were then used to calculate Poisson's ratio, using equations 1 and 2.

The method of weighted least-squares was employed because this approach allowed the incorporation of the previously determined uncertainties found for the measurements. The quantity that one desires to minimize in the Wadati weighted least-squares regression is :

$$\sum_{i=1}^N (a + bx_i - y_i)^2 w_i, \quad (4)$$

where a = the y-intercept of the fitted line,
 b = the slope of the line ($= \frac{\alpha}{\beta} - 1$, where α and β are the velocities of the P and S wave respectively),
 x_i = the P arrival-times,
 y_i = the S-P intervals,
 w_i = the associated uncertainties for the S-P intervals, and
 N = the number of points (stations) in the plot.

Expression 4 represents the sum of the squares of the deviations of the data from the weighted least-squares line. To minimize this expression, derivatives with respect to a and b are found and set equal to zero. The resulting two equations are given by :

$$\sum_{i=1}^N (a + bx_i - y_i) w_i = 0 \quad (5)$$

and

$$\sum_{i=1}^N (x_i)(a + bx_i - y_i)w_i = 0. \quad (5a)$$

Equations 5 and 5a may be solved simultaneously for a and b to obtain:

$$a = \frac{\left(\sum_{i=1}^N y_i w_i\right)\left(\sum_{i=1}^N x_i^2 w_i\right) - \left(\sum_{i=1}^N x_i y_i w_i\right)\left(\sum_{i=1}^N x_i w_i\right)}{\left(\sum_{i=1}^N w_i\right)\left(\sum_{i=1}^N x_i^2 w_i\right) - \left(\sum_{i=1}^N x_i w_i\right)^2} \quad (6)$$

and

$$b = \frac{\left(\sum_{i=1}^N x_i w_i\right)\left(\sum_{i=1}^N y_i w_i\right) - \left(\sum_{i=1}^N x_i y_i w_i\right)\left(\sum_{i=1}^N w_i\right)}{\left(\sum_{i=1}^N x_i w_i\right)^2 - \left(\sum_{i=1}^N x_i^2 w_i\right)\left(\sum_{i=1}^N w_i\right)} \quad (6a)$$

Equation 2 may now be rewritten in terms of b, the slope, to get :

$$V = \frac{b^2 - 2b - 1}{2b^2 + 4b} \quad (2')$$

The degree to which the data fitted the deduced line was ex-

pressed by a correlation coefficient given by :

$$R = \left(\frac{\text{total sum of squares} - \text{residual sum of squares}}{\text{total sum of squares}} \right)^{1/2} \quad (7)$$

which for the Wadati diagrams was :

$$R = \left(\frac{\sum_{i=1}^N (y_i - \bar{y})^2 - \sum_{i=1}^N (y_i - (a + bx_i))^2}{\sum_{i=1}^N (y_i - \bar{y})^2} \right)^{1/2} \quad (7a)$$

where $\bar{y}$ = the mean of the y_i .

In addition to having a correlation coefficient for the plots, a method for determining the standard deviation of the calculated Poisson's ratios was devised. The method adopted the approximation :

$$\text{Var}(\sqrt{b}) \approx \left(\frac{d(\sqrt{b})}{db} \right)^2 \text{Var}(b) , \quad (8)$$

where $\sqrt{b}$ = Poisson's ratio as a function of slope (equation 2'),

$\frac{d(\sqrt{b})}{db}$ = the derivative of equation 2' with respect to b , and

$\text{Var}(b)$ and $\text{Var}(\sqrt{b})$ are the variances of the slope and Poisson's ratio respectively (Gutjahr, personal communication).

The standard deviation of $\sqrt{b}$ was obtained by taking the square-root of the right-hand side of equation 8.

In order to obtain the variance of the slope in equation 8, a rearrangement of the summations in equation 6a is necessary to isolate the y_i terms. Consider the generalized product of summations:

$$\sum_{i=1}^N x_i w_i \sum_{j=1}^N y_j w_j = \sum_{i=1}^N \sum_{j=1}^N x_i w_i y_j w_j = \sum_{j=1}^N \sum_{i=1}^N x_j y_j w_j w_i. \quad (9)$$

Given the above identity, one can rewrite the numerator of the right-hand side of equation 6a to obtain :

$$\begin{aligned} \text{numerator of } b &= \sum_{j=1}^N \left(\sum_{i=1}^N x_i w_i \right) w_j y_j - \sum_{j=1}^N \left(\sum_{i=1}^N w_i \right) x_j y_j w_j \\ &= \sum_{j=1}^N \left(\sum_{i=1}^N x_i w_i - \sum_{i=1}^N w_i x_j \right) w_j y_j, \end{aligned} \quad (10)$$

so that equation 6a becomes :

$$b = \frac{\sum_{j=1}^N \left(\sum_{i=1}^N x_i w_i - \sum_{i=1}^N w_i x_j \right) w_j y_j}{\left(\sum_{i=1}^N x_i w_i \right)^2 - \left(\sum_{i=1}^N x_i^2 w_i \right) \left(\sum_{i=1}^N w_i \right)}. \quad (6a')$$

The variance of the slope may be easily found from equation 6a', and is given by :

$$\text{Var}(b) = \frac{\sum_{j=1}^N \left(\sum_{i=1}^N x_i w_i - \sum_{i=1}^N w_i x_j \right)^2 w_j^2 \text{Var}(y_j)}{\left[\left(\sum_{i=1}^N x_i w_i \right)^2 - \sum_{i=1}^N x_i^2 w_i \sum_{i=1}^N w_i \right]^2}, \quad (11)$$

where $\text{Var}(y_j)$ = the variance of the S-P measurements.

Note that the expression, $w_j^2 \text{Var}(y_j)$, is identically equal to one, since the variance of the y_j 's is just the reciprocal of the square of the standard deviations (the w_j 's).

The final calculation performed for the Wadati diagram analysis was the determination of weighted least-squares origin-times. These origin times were used to check the reliability of the iterative origin-times generated by the microearthquake location program. Large differences (several seconds) between iterative and weighted least-squares origin-times were occasionally encountered in microearthquake solutions. When a difference of greater than one second was found between origin-times, the data for that plot were excluded from any further analysis. Solutions exhibiting this inconsistency were either solutions for earthquakes falling well outside the array, or were events having all recording stations nearly equidistant from their hypocenters. (A discussion of the latter situation, and its effects upon the inversion problem may be found in the paper by Aki(1976).) The reasons for excluding these earthquakes from the analysis were : (1) the travel-times obtained from the iterative origin-times and P arrival-times for an event were not consistent with any of the S-P measurements for the microearthquake, and (2) when all the recording seismograph stations were equidistant from the earthquake, the corresponding Wadati diagram was very poorly defined, with all the

(a)

(b)

Figure 9. (a) Wadati diagram for an eight station location used in this report. (The weights were all 0.05 sec.)

(b) Travel-time diagram used to obtain v for crustal wedge from station SC.

data plots plotting in essentially the same area. An example of a Wadati diagram used in the above analysis is shown in Figure 9a.

Section 5

The next analysis to be performed was designed to ascertain whether or not plots of S-P intervals versus travel-times could be used to obtain Poisson's ratio (i.e. whether the ratio $\frac{\alpha}{\beta}$ is relatively constant for the study area). The method involved the construction of a single composite plot of all the microearthquake data used in this study. Travel-times were calculated using iterative origin-times from 259 microearthquakes and their associated P arrival-times. The weights employed in the composite plot were obtained by summing the uncertainties in origin-time, P arrival-time and S-P interval for each point in the plot. The results of the analysis were expressed in terms of a correlation coefficient given by equation 12, below :

$$R = \left[\frac{\sum_{i=1}^N (y_i - \bar{y})^2 - \sum_{i=1}^N (y_i - bt_i)^2}{\sum_{i=1}^N (y_i - \bar{y})^2} \right]^{\frac{1}{2}}, \quad (12)$$

where y_i = the S-P intervals,
 $\bar{y}$ = the mean of the y_i 's,
 b = the slope of the fitted line, and
 t_i = the travel-times associated with the S-P intervals.

Section 6

The final segment of the data analysis and reduction involved the determination of spatial variations in Poisson's ratio in the upper crust for the area of study. For a complete description of the technique employed, the reader is directed to the paper by Caravella(1976). Essentially, plots of S-P intervals versus travel-times were constructed for body waves arriving along particular ten degree increments of azimuth about a seismograph station. The principle is represented graphically in Figure 10. A weighted least-squares linear regression was performed for each of the travel-time diagrams which had a minimum of three data points. Poisson's ratio was then calculated, using equations 1 and 2, from the slopes obtained from the linear regressions.

The method of weighted least-squares linear regression, as applied to S-P interval versus travel-time plots, is essentially the same as that employed for the Wadati analysis, although the algebra involved is simplified somewhat. The expression used to calculate the correlation coefficients for the travel-time diagrams is given by equation 12, above. The quantity minimized in this analysis given by:

$$\sum_{i=1}^N (bt_i - y_i)^2 w_i \quad (13)$$

where N is the number of points used in the plot and t_i , y_i , and w_i represent the estimates (measurements or calculations) of the travel-time, S-P interval, and the weight associated with the S-P interval respectively. Thus the travel-time analysis was reduced to the determination of a single variable; the slope of the fitted line through the origin. Values for Poisson's ratio obtained from the slopes were

Figure 10. Graphical representation of the technique employed to spatially distribute Poisson's ratio (from Caravella, 1976). Body wave arrivals along ten degree increments of azimuth about a seismograph station are used to form an S-P interval versus travel-time plot. Crustal wedges are then constructed for each plot made.

assigned to the appropriate ten-degree sector of the seismograph station being examined.

To minimize the expression in 13, the derivative with respect to b is found and set equal to zero to obtain:

$$b \sum_{i=1}^N t_i^2 w_i - \sum_{i=1}^N t_i y_i w_i = 0. \quad (14)$$

the slope is then found to be:

$$b = \frac{\sum_{i=1}^N t_i y_i w_i}{\sum_{i=1}^N t_i^2 w_i} \quad (15)$$

The approximation given in equation 8 was again employed in the calculation of the standard deviation of Poisson's ratio. The expression used to obtain the variance of the slope for the travel-time plots is:

$$\text{Var}(b) = \frac{\sum_{i=1}^N t_i^2 w_i^2 \text{Var}(y_i)}{\left(\sum_{i=1}^N t_i^2 w_i \right)^2}, \quad (16)$$

where again the $w_i^2 \text{Var}(y_i)$ term is identically equal to one, as $\text{Var}(y_i) = 1/w_i^2$. An example of a travel-time diagram used in this study is shown in Figure 9b. The information obtained by this segment of the study was used to construct crustal wedges having angular widths of ten degrees and lengths corresponding to the most distant epicenter used in the plot. (see Figure 10) Intersections of crustal wedges exhibiting abnormally high (> 0.275) Poisson's ratios were used

Figure 11. Diagram showing the distribution of Poisson's ratios found from the Wadati diagram analysis. Values of Poisson's ratio were grouped by increments of 0.05, where the numbers appearing above a particular block represent the number of Wadati plots yielding a value of ν within that 0.05 interval.

to define anomalous areas in the crust for the region of study. (see Figures 12-16).

VI. Results and Discussion

The average value for Poisson's ratio, obtained from the Wadati analysis, was 0.2514 with a standard deviation of 0.0516. The Poisson's ratios calculated from individual Wadati diagrams were found to be uniformly distributed about their mean (see Figure 11). This latter result assures statistical significance of the average Poisson's ratio obtained, at a one standard deviation confidence level. The average ν found by Sakdejayont (1974) and Caravella (1976) are both within one standard deviation of the result obtained in this paper.

Several of the Wadati diagrams, which could not be eliminated from the analysis by any of the criteria previously mentioned, yielded unreasonably low values for Poisson's ratio (0.039 to 0.111). The five Wadati plots exhibiting this property were felt to be of importance, as unlikely values of ν are still obtained when a two standard deviation confidence interval about these values of ν is applied. The Wadati diagrams for these microearthquakes are included in Appendix A. The author believes that these anomalous Wadati plots were the result of an insufficient amount of data to effectively determine a line. No correlation between the hypocenter locations or station distributions for these anomalous diagrams was found.

The results of the composite S-P interval versus travel-time plot, discussed in Section 5 of this report, showed substantial correlation among the data. A correlation coefficient of 0.9065 was obtained for the plot, which consisted of more than 1,000 data points. This rela-

tively high correlation indicates that the travel-times were consistent with their associated S-P intervals. This provides some justification for using the travel-times to obtain values for Poisson's ratio in the spatial distribution analysis.

The final segment of this report involved the determination of spatial variations in Poisson's ratio. The raypath technique devised by Caravella (1976) and discussed in Section 6 of this paper was used to determine values of ν for crustal wedges originating from various seismograph stations. An azimuthal increment of ten degrees was chosen as the angular width of the wedges because it was felt that smaller azimuthal increments were not justified (due to hypocenter uncertainties in the microearthquake locations), and that sufficient accuracy could be attained using a ten degree width.

A total of 167 wedges were constructed from the data, using at least three points (three S-P intervals and their associated travel-times) for any ten degree sector. Of the 167 wedges, nearly 30% had Poisson's ratios greater than 0.275. In order to map areas with anomalously high Poisson's ratio, it was decided to use only those wedges whose values of ν were greater than 0.275 at the 95% confidence level, i.e. wedges whose ν 's were less than 0.275 after having two standard deviations subtracted from them were not considered anomalous. This technique reduced the total number of anomalous wedges to 40. Intersections among the 40 remaining wedges were examined in order to map areas of anomalously high Poisson's ratio.

The mapping of wedges having low (<0.170), normal (0.170 to 0.274) and high (>0.275) values of Poisson's ratio was used to assure

Figure 12. Anomalous area A, defined by 7 intersecting crustal wedges having Poisson's ratios of 0.275 and above at the 95% confidence level.

Figure 13. Anomalous area B, determined from the intersections of 6 crustal wedges having Poisson's ratio values of 0.280 and above at the 95% confidence level.

Figure 14. Anomalous area C, determined by 6 crustal wedges having values of Poisson's ratio of greater than 0.279 at the 95% confidence level.

Figure 15. Anomalous area D and anomalous area that cannot be supported by this report. Area D was determined from 6 wedges having Poisson's ratios of 0.275 or greater at the 95% confidence level.

consistency between the data and the locations of anomalous areas of Poisson's ratio, and to provide control for determining the size and shape of those areas. The rule followed in the mapping procedure was to explain the maximum number of anomalous crustal wedges with the least number of anomalous areas. The results of this analysis are shown in Figures 12-16.

A total of 5 areas of anomalous Poisson's ratio was determined using intersections of four or more crustal wedges having Poisson's ratios exceeding 0.275 at the 95% confidence level. When crustal wedges of both normal and low values of ν were added to the analysis, only 4 of the 5 areas of anomalous Poisson's ratio could be supported. These four areas, labeled A, B, C, and D in Figures 12-16, can be used to explain all but three of the anomalous crustal wedges obtained in this study.

Anomalous area A (Figure 12) was located using seven intersecting crustal wedges having Poisson's ratios greater than 0.275 at the 95% confidence level. The seismograph stations from which these wedges originated were CC, CK, DM, SL and WT, with two adjacent wedges from DM. The size and shape of area A was established by using the above seven wedges and crustal wedges of normal and low Poisson's ratio from stations BG, CC, CK, MY, LAD and WT. Additional crustal wedges exhibiting anomalously high, normal, and low values for ν could have been used in locating area A, however their addition would not have altered any of the boundaries already determined, and would merely duplicate the existing information. Such extraneous crustal wedges will henceforth not be mentioned.

The second region of anomalously high Poisson's ratio determined by this study is located along the southern margin of the La Jencia basin, approximately 3 km west of station IC. The intersections of six crustal wedges having Poisson's ratios in excess of 0.280 at the 95% confidence level were used to determine this area (denoted B, in Figure 13). Wedges having normal to low values of Poisson's ratio from stations CM, HC, NG, SC, WM and WT were used to establish the boundaries of area B while stations CM, FC, NG, and SC were the sources of the six anomalous wedges. The region shown in Figure 13 could consist of several smaller areas having anomalously high Poisson's ratios, as proposed by Johnston (1978) and Wallace (1978) (see Figures 17 and 18), however the crustal wedge approach adopted in this paper is not able to resolve such fine detail. Area B is the largest anomalous region found in this study, with an areal extent of nearly 20 km².

A third anomalous area, determined by six intersecting crustal wedges having Poisson's ratios greater than 0.279 at the 95% confidence level, is located along the northern tip of the Chupadera Mountain Block, approximately 1 km east of seismograph station IC. (Area C, Figure 14). This region was determined by anomalous wedges originating from stations CM, DM, NG, SC and TS, and was bounded by crustal wedges having normal to low values of Poisson's ratio from stations CM, DM, NG, SC and WT. Area C might have been better resolved if data had been available to construct crustal wedges crossing the anomaly from seismic stations IC and WM.

The final area of anomalously high Poisson's ratio that can be sup-

ported by this study is located approximately 4 km north-west of station RM (Region D; Figure 15). Six wedges having Poisson's ratios of 0.275 or above at the 95% confidence level were drawn from seismograph stations CM, GM, HC, NG and SC to obtain region D. The boundaries of area D were primarily determined by crustal wedges having normal values of Poisson's ratio from stations GM, IC and WT. An absence of earthquake epicenters (see Figure 4) is associated with this region, which might indicate a relatively shallow anomaly (Sanford, personal communication, 1978).

A fifth area defined by intersections of five crustal wedges having Poisson's ratios of 0.275 or greater at the 95% confidence level is shown in Figure 15. This area is located approximately 6 km north-east of station SL. The author felt that the anomalous wedges from stations DM and CM, used to locate the area, have already been explained by regions A and C, and thus should not be used to resolve an additional anomalous area. The two remaining wedges from CC and BB, which have Poisson's ratios of 0.286 and 0.281 at a 95% confidence level respectively, can not be explained either by reading error or a lack of sufficient data. The intersecting wedges from stations BG, CK, SL, LAD and LPN all have normal to low values of Poisson's ratio. The small amount of corroborating evidence combined with the disagreement among the data was felt to be sufficient grounds for not defining an anomalous region in this area.

As was stated previously, all but three of the anomalous crustal wedges found in this study could be explained by the four areas discussed above. The two anomalous wedges from stations BB and CC have

Figure 16: Map showing the four regions of anomalously high (> 0.275 at the 95% confidence level) Poisson's ratio determined in this study.

Figure 17. The positions of the eight proposed attenuating volumes of crustal rock relative to the physiographic features of the Socorro area (from Johnston (1978)).

Figure 18. The positions of the eight proposed crustal volumes relative to the microearthquake swarms located by Wallace (1978).

already been mentioned. The final unexplained crustal wedge, which has a Poisson's ratio of 0.281 at the 95% confidence level, is the 50-60 degree sector from station CM. There are no other crustal wedges (anomalous or normal) that intersect this wedge, and no errors in the measurements or analysis were found. The shallow magma body proposed by Johnston (1978), located to the northeast of station CM, could explain the high value of Poisson's ratio obtained for this crustal wedge, but this proposed anomaly is not able to be supported by this study.

Figure 16 displays the four anomalous areas, determined by this report, on a single map so that the results of the studies of Caravella (1976), Johnston (1978) and Wallace (1978) may be more easily compared.

VII. Conclusions

The average value for Poisson's ratio for the upper crust in the Socorro, New Mexico area is 0.251 ± 0.052 . Localized areas of anomalously high Poisson's ratio (0.275 at the 95% confidence level) exist, and may be spatially mapped using body wave arrivals from microearthquakes.

The method of determining the spatial distribution of Poisson's ratio using crustal wedge intersections provides an excellent indirect procedure for mapping anomalous areas of Poisson's ratio. Smaller azimuthal increments and larger data sets could be employed in future studies to obtain a more detailed spatial distribution.

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APPENDIX A :

Wadati diagrams which yielded unreasonably
low values for Poisson's ratio

S-P
Interval
(sec.)

APPENDIX B:

Discussion of spatial distribution of crustal wedges of low
Poisson's ratio and the 167 crustal wedges found
from the analysis

Although this report was primarily concerned with mapping areas of anomalously high Poisson's ratio, regions having anomalously low and normal values of ν should also be considered in order to make the spatial distribution of ν in the area of study complete. For this reason, the 167 crustal wedges determined in this study are plotted together on the following pages. The mean values of ν , rather than values obtained for a 95% confidence level, were employed in these plots. The color coding and letter designations for these diagrams is as follows:

A	red	$\nu > 0.30$
B	orange	$0.29 < \nu \leq 0.30$
C	yellow	$0.28 < \nu \leq 0.29$
D	purple	$0.27 < \nu \leq 0.28$
E	light green	$0.26 < \nu \leq 0.27$
F	blue	$0.25 < \nu \leq 0.26$
G	brown	$0.24 < \nu \leq 0.25$
H	black	$0.23 < \nu \leq 0.24$
I	dark green	$\nu < 0.23$

A high degree of correlation among crustal wedges from stations BG, CK and LAD indicates that an area of anomalously low Poisson's ratio (< 0.23) exists in the northern Socorro basin, running nearly parallel with the Rio Salado. Wedges of low to normal ν correlate quite well for the northern two thirds of the Socorro-Lemitar Mountain Block, the southern extension of the Los Pinos Mountains, the

eastern portion of the Magdalena Mountains, and the northern one third of the La Jencia Basin. Lack of a sufficient amount of data prevents making further generalizations as to the values of Poisson's ratio in the area of investigation.

0 10 20 KM

S-P INTERVAL (secs)

$V_p = 5.8 \text{ km/sec}$

$V_b = 3.35 \text{ km/sec}$

0 10 20 KM

S-P INTERVAL (secs)

$V_p = 5.8 \text{ km/sec}$ $V_b = 3.35 \text{ km/sec}$

0 10 20 KM

S-P INTERVAL (secs)

$V_p = 5.8 \text{ km/sec}$ $V_b = 3.35 \text{ km/sec}$

0 10 20 KM

S-P INTERVAL (secs)

$V_p = 5.8 \text{ km/sec}$

$V_s = 3.35 \text{ km/sec}$

0 10 20 KM

S-P INTERVAL (secs)

$V_p = 5.8 \text{ km/sec}$

$V_b = 3.35 \text{ km/sec}$

0 10 20 KM

107.2 107.1 107.0 106.9 106.8 106.7 106.6 106.5

S-P INTERVAL (secs)

$V_p = 5.8 \text{ km/sec}$

$V_s = 3.35 \text{ km/sec}$

APPENDIX II

MICROEARTHQUAKE FREQUENCY ATTENUATION
OF S PHASES IN THE RIO GRANDE RIFT
NEAR SOCORRO, NEW MEXICO

by

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fulfillment

of

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ABSTRACT

Arrays of 5 to 9 portable, high-gain ($\sim 10^6$ magnification at 40 Hz) seismographs were used to record microearthquakes from April, 1975, through January, 1978, in the vicinity of Socorro, New Mexico. Eight volumes in the crust have been proposed to explain anomalous absorption of the higher frequencies of S waves in the recorded microearthquakes. Crustal volumes with anomalous attenuation were defined by intersecting raypaths, along which unusual high frequency S wave absorption was observed. The degree of absorption was determined by measuring the dominant frequency of the S wave using a zero-crossing counting technique and relating this quantity to Q, the quality factor. Approximately 5 percent of the raypaths examined showed anomalous S wave frequency attenuation.

Five possible mechanisms for the anomalous absorption were considered (1) variation of source parameters, (2) spectra variation with change in azimuth around the focus, (3) normal absorption for raypaths traversing large distances, (4) station effects, and (5) raypath travel through an anomalous crustal volume. The effects of the first four mechanisms are shown to be minimal. Partial melts, heated rock, or fluid-filled microfractures could explain the fifth mechanism. Heated rock appears to be the most likely explanation for the observations.

Three crustal volumes are defined which are most likely to be associated with magmatic intrusion (1) the region ~ 3.5 km ESE of station WT, in the Socorro Basin (depth: ~ 3.2 km), (2) the region ~ 3.0 km NW of station WM in the southeast of the La Jencia Basin (depth: ~ 4.6 km), and (3) the region ~ 2.5 km NE of station SC in the southwest of the La Jencia Basin (depth: ~ 4.1 km). The depths given are the shallowest

points of the anomalous crustal volumes.

The relation $Q_\alpha/Q_\beta = v_\alpha/v_\beta$ is proposed (v = dominant frequency, α and β refer to P and S waves, respectively). The normal, unattenuated value of Q_α/Q_β for the Socorro area is observed to be 1.60. The mean values for the three anomalous volumes described above, as calculated from the Q_α/Q_β ratios associated with the raypaths defining the volumes, are (1) 2.76, (2) 2.71, and (3) 2.29. The higher the ratio the greater the absorption. The higher ratios are all attributable to an increase in the absorption of S wave energy.

INTRODUCTION

Microearthquakes have been recorded regularly in the Socorro area since June 1960. In the past few years these events have been used to investigate the presence of magma at intermediate depths beneath the Rio Grande rift. Several geophysical observations have led researchers to believe that an extensive ($\sim 1700 \text{ km}^2$) layer of magma exists at depths of 18 to 20 km (1) diffuse geographical distribution of seismic activity, (2) unusually strong S to P (S_xP) and S to S (S_xS) reflections, (3) high heat flow, and (4) historical surface uplift (Sanford et al., 1976). In addition to these investigations research is presently being conducted to attempt to identify the spatial extent of shallower ($< 13 \text{ km}$) magma bodies whose source could be the deeper magma layer. Such research is the goal of the study described in this report.

The purpose of this investigation was to map the volumes of crust defined by intersecting raypaths along which frequency attenuation in the S wave of microearthquakes was observed. Assuming such attenuation to be due to heated rock the crustal volumes so defined should approximately locate volumes affected by partial melts at greater depths. Previous

related studies include (1) Shuleski (1976), (2) Sanford (1977), (3) Matumoto (1971), and (4) Kubota and Berg (1968). All these studies involve the mapping of intersecting raypaths along which screening, or nearly complete screening, of the S wave of microearthquakes is observed. In particular, Shuleski (1976) performed a complete analysis of fault plane solutions and angles of incidence in the Socorro area and determined that only 38 percent of the weak or missing S waves could be explained by these factors. The remaining 62 percent he attributed to screening by magma bodies.

Figures 1 and 2 are presentations of the results of Shuleski (1976) as modified by Sanford (1977). The lines represent raypaths, drawn from hypocenters to the stations at which screening was observed. Ratios of P wave to S wave amplitudes are plotted to indicate the degree of screening. These ratios eliminate problems resulting from differences in strength of microearthquakes and distances to recording stations (Shuleski, 1976). The tentative locations for four magma bodies are outlined; when dashed lines are used the location is not well known. The numbered points, lying along the axes of the outlined regions, are depths calculated to raypaths crossing the axes of the regions.

The basic measurement in this study was the counting of the number of zero-crossings observed in the S wave. This was then converted to dominant frequency. Microearthquake data were obtained from April 1975 to January 1978 using a moveable array of 5 to 9 seismograph stations. Considerable effort was made to obtain accurate hypocenter locations as the accuracy of the final placement of the anomalous volumes depended on correct raypath plotting.

Figure 1. The S wave screening data of Shuleski (1976), as modified by Sanford (1977).

Figure 2. The S wave screening data of Shuleski (1976), as modified by Sanford (1977).

GEOLOGICAL SETTING

The area of investigation is located within the Rio Grande rift approximately 125 km south of Albuquerque. The rift is a major extensional structure formed by east-west tension beginning about 25 to 29 m.y. ago and continuing to the present (Chapin and Seager, 1975). The rift extends from southern New Mexico into central Colorado and consists of a series of interrelated structural depressions with raised margins having a general northern trend. The rift penetrates the Southern Rocky Mountains of northern New Mexico and southern Colorado, lies between the Colorado Plateau and High Plains in central New Mexico, and merges to the south in a complicated manner with the Basin and Range province. Approximately 50 km north to 30 km south of Socorro the rift is further modified by intergraben horsts believed to be only 9 to 10 m.y. old, suggesting a recent crustal structural anomaly (Chapin and Seager, 1975). The map shown in Figure 3 illustrates the general physiographic features and locations of stations utilized in this study. Intergraben horsts shown are the Socorro-Lemitar and Chupadera mountains, separating the La Jencia and Socorro basins.

A large amount of evidence indicates that volcanic activity accompanied formation of the rift. The most active periods appear to have been from 20 to 26 m.y. ago and from 5 m.y. ago to the present (Chapin and Seager, 1975). The first period of volcanism has basaltic andesites associated with it, while true basalts dominate the flows of very recent age in the Socorro area.

A ribbon of high heat flow (>2.5 HFU) is roughly coincident with a number of thermal springs occurring along the western margin of the rift (Reiter et al., 1975), the hottest and largest of which are located in

Figure 3. Physiographic features of the Socorro area, seismic station locations, and outline of the deep (18 km) magma body (from Sanford, and others, 1977).

the southern end of the Socorro-Lemitar mountains. This is also an area of very high heat flow (Sanford et al., 1977).

The outline of the deep, extensive magma body located between 18 and 20 km in depth is shown in Figure 3. Dotted lines indicate boundaries where more data is required to map them accurately. The shape of the body appears to be a thin, elongate sill having a general north-south trend with its axis following beneath the Socorro-Lemitar mountain block. The minimum depth to the body of 18 km is located beneath Socorro Mountain. It deepens to approximately 20 km along the margins (Rinehart, 1976). The body dips northward approximately 6° (Sanford, Alptekin, and Topozada, 1973).

INSTRUMENTATION

Two basic types of portable seismic systems were employed throughout the data collection period. For the interval April 1975 through January 1978, an array of 5 to 6 Sprengnether MEQ-800 analog recording units was deployed. This basic array was supplemented, starting in April, 1977, with two Sprengnether DR-100 digital recording units. In addition to these units a Kinematics PS-1 analog unit was used to record station LAD which is telemetered via telephone line directly to the campus of New Mexico Institute of Mining and Technology. The MEQ-800 seismographs served as the primary source for the data. The DR-100 systems were mainly used to improve hypocenter locations by expanding and better defining the arrays. They were also used to determine the dominant frequencies of unattenuated P and S phases because they produced the best high-resolution records.

From April, 1975 to September, 1977 most of the arrays were deployed in the southern part of the area shown in Figure 3 (roughly south of

34°05'N). From September, 1977 to January, 1978, the arrays were to the north (between 34°08'N and 34°30'N). However most of the data collected was from the south and thus this study is primarily concerned with that area.

The MEQ-800 seismic system records helically on a drum covered with smoked paper at a speed of 120 ± 0.6 mm/min. The gain stable amplifier has settings ranging from 60 to 140 db in discrete 6 db intervals. The most common settings used in the field were 84 and 90 db. Both low and high frequency filters were available. The low frequency filter was always set at Out and the high frequency filter was set at either Out or 30 Hz. The high frequency 30 Hz setting attenuated frequencies higher than 30 Hz, but did not seriously affect the response of the system to the dominant frequencies observed in microearthquakes (P wave < 30 Hz). Magnification response curves for the system at both settings of the high frequency filter, Out and 30 Hz, are shown in Figure 4 for both the Willmore vertical component (1.5 Hz natural frequency) and Mark-Products L4-C vertical component (1.0 Hz natural frequency) seismometers.

Each MEQ-800 has a quartz crystal chronometer which produces time marks on the record at one minute intervals. The chronometers were synchronized at the beginning of each recording week with a time signal from WWV and were checked for drift at the end of the week in the same manner. The recording week generally started Monday morning and the units were retrieved Friday afternoon. During this time period the chronometer drift was assumed to be linear. This has been verified by trial runs of the units in which twice-daily checks were made under field conditions with the time signal from WWV. The maximum variation from linearity found was approximately 0.03 sec, but most of the time the variation was

Figure 4a. Magnification response curves for the MEQ-800 seismographs. The high-frequency filter is set in the 30 Hz position (adapted from Rinehart, 1976).

Figure 4b. Magnification response curves for the MEQ-800 seismographs. The high-frequency filter is set in the Out position (from Shuleski, 1976).

much less than this. The linear drift observed was always less than 0.10 sec/day.

The two DR-100 units record digitally on magnetic tape cassettes at a sampling rate of 100/sec. The amplifier gains were set at discrete 6 db intervals in the range 78 to 96 db. The filters were set at High-Out and Low-Out the entire time the units were in the field.

The seismometers for the DR-100 units were Mark-Products L4-C vertical component geophones. A Sprengnether DP-100 tape playback unit and Beckman Dynograph strip-chart recorder were used to obtain analog records from the digital magnetic tapes. The strip-chart recorder produced an ink on paper record with a wide variety of playback speeds possible.

Listed in Table 1 are the locations of the seismic stations occupied during this study.

DATA ACQUISITION AND REDUCTION

Almost all of the primary data collected for this study (the counting of zero-crossings in S waves) were taken from the MEQ-800 records. The entire set of records, from April, 1975 to January, 1978, was first scanned by eye to pick out events that appeared to have attenuation of high frequencies in the S wave but not in the P phase. In doing this a 'whiteout' criterion was used to make a rapid determination of this type of attenuation. The recording pen leaves a white trace on the smoked paper as the microearthquake signature is written. At high frequencies, individual, adjacent cycles are indiscernable and the trace leaves a band of white. This is the definition of 'whiteout' and photographic examples are presented in Plate 1. However, at lower dominant frequencies individual cycles can be resolved and the zero-crossings counted. From

Table 1. Location of Seismic Stations

<u>Station</u>	<u>Latitude</u>	<u>Longitude</u>	<u>Elevation (m)</u>
BG	34.2068	106.8205	1516
CC	34.1442	106.9812	1649
CM	33.9501	106.9576	1640
CU	34.1573	106.7785	1585
DM	34.1075	106.8079	1536
FM	34.0829	106.8047	1537
HC	34.0658	107.2361	2240
IC	33.9870	106.9967	1730
LAD	34.4583	107.0375	1768
LPM	34.3076	106.6336	1737
MG	34.1305	107.2425	2024
MY	34.1667	106.7459	1645
NJ	33.9924	106.6253	1644
SC	34.0100	107.0894	2073
SL	34.2234	106.9910	1615
TA	34.0498	106.7751	1558
WM	34.0120	106.9929	1673
WT	34.0722	106.9459	1555
GM	34.1454	107.2345	1945
RI	34.4234	107.2075	1530
CK	34.2725	106.7702	1578
BB	34.4090	106.6818	1615
FC	33.8950	107.0504	1850

0 1 2

.01

AUGUST 12, 1975 MY
7:09 UST

0 1

.01

JUNE 3, 1975 FM
9:38 UST

0 1

JUNE 3, 1975 SC
15:10 UST

0 1 3

.01

MAY 29, 1975 CM
7:14 UST

Plate 1. Examples of events in which the P and S waves are unattenuated.

measurements on records approaching this limit, it was found that the lower frequency range in which whiteout was still observed was approximately 12 to 18 Hz. This frequency range was a function of trace thickness and general background noise. In general, resolution on most records was quite good and frequencies in the range of interest for this study were readily obtained by the zero-crossing counting technique. Thus, when first scanning the records by eye, events were chosen in which the P phase showed 'whiteout' and the S phase exhibited easily counted zero crossings. Examples of events meeting these criteria are shown in the photographs of Plate 2.

After the initial selection and location of the attenuated event, the zero crossings were converted to dominant frequency using the relation,

$$v = \frac{n-1}{2} , \quad (1)$$

where n is the number of zero crossings. The zero reference was defined by extending the line of the undisturbed seismic trace preceeding the event across the signature of the P and S phases through the coda.

The final result of this event selection process was a group of 83 microearthquake events in which the P phase appeared relatively unattenuated but the S wave showed an unusually low dominant frequency. The total number of events examined was approximately 320. Assuming an average of 5 raypaths per event, about 5 percent of the raypaths examined showed an unusually low dominant S wave frequency.

Two types of computer location routines were employed to determine the placement of the hypocenters. The first was an iterative numerical inversion method in which an initial estimate of the location and origin time were input along with the times of the direct P arrivals at each

.01

AUGUST 25, 1975 FM
19:37 UST

.01

JANUARY 23, 1976 DM
2:53 UST

.01

JANUARY 21, 1976 CM
14:18 UST

.01

AUGUST 17, 1977 DM
2:20 UST

Plate 2. Examples of events in which the P wave is unattenuated, but the S wave shows loss of high frequencies.

.01

SEPTEMBER 2, 1977 CC
13:29 UST

.01

JANUARY 23, 1976 CM
7:22 UST

.01

OCTOBER 8, 1976 DM
00:21 UST

.01

FEBRUARY 10, 1977 DM
7:33 UST

Plate 2. (Continued)

of the stations in the array. The initial estimate for the origin time was taken from a Wadati plot and the initial epicenter location was determined from intersecting arcs whose radii were based on S-P times obtained from the records at each station in the array. An initial depth estimate of 7.0 km was usually made. Depth was the least well determined parameter. Rinehart (1976) gives a detailed account of the location procedure and errors involved.

After an acceptable location was obtained from the iterative routine the resulting origin time, along with the measured P arrival times, were used in a computer adaptation of the graphical procedure of intersecting hemispheres described by Richter (1958). This program produced a map of intersecting circles. Chords were drawn between the intersection points of the circles and the middle of the area defined by the intersection of these chords established the epicenter. The tightness of the cluster of intersecting chords indicated the quality of the solution. If the chords intersected at nearly the same point, the solution was considered good and no further modifications were made. If there was a large spread in the intersection of the chords, the origin time was allowed to vary within 0.5 seconds of the Wadati origin time until the tightest cluster of chord intersections was obtained. Usually the solution obtained in this manner was very close to the location produced by the iterative inversion program.

All the microearthquakes used in this study were located once following the above procedure. They were then located a second time using the locations obtained the first time as initial estimates. New station corrections, discussed below, and, in some cases, new direct P arrival times, were used in the second locations. The results from these second

locations were then used in the study. Only the iterative, numerical inversion routine was used for the second locations.

Corrections applied to the P arrival times prior to insertion in the location programs were (1) clock drift, and (2) station corrections. The clock drift for the four day recording week was assumed to be linear and separate drift curves were constructed every week for each unit in the field. Station corrections were estimated by carefully measuring arrival times of explosions whose locations and origin times were accurately known (Sanford, personal communication). Differences between theoretical and observed travel times for these explosions were attributed to effects of the geological column immediately beneath each station. Once accurate corrections were known for five or more stations in the array, other stations without corrections could be added to the array one at a time and run through the iterative inversion routine. Residuals obtained at the added station for a number of microearthquake events were then used to determine a station correction for the added station. Station corrections obtained from explosion data and the iterative technique are listed in Table 2.

Hypocenters were considered acceptable if (1) residuals calculated by the iterative inversion program were approximately equal to the timing error (0.05 sec), (2) error bars calculated by the program were of the order of the weighting used for each parameter (~1 km for longitude and latitude, ~3 km for depth and ~0.3 sec for the origin time), and (3) standard deviations, as calculated by the program, were less than ~0.06. Hypocenter locations meeting these criteria are listed in Table 3 along with the best estimate of the error bars associated with each parameter. The events in Table 3 constitute the data set used in this study. Of

Table 2. Station Corrections.

<u>Station</u>	<u>Correction</u>	<u>Method of Determination</u>
WT	0.00	<p>explosion</p> <p>iterative inversion</p>
CC	-0.11	
SL	-0.11	
GM	-0.11	
IC	-0.38	
RM	-0.44	
NG	-0.35	
CM	-0.42	
SC	-0.48	
DM	-0.11	
FM	-0.22	
BG	-0.22	
CU	-0.17	
WM	-0.32	
RI	-0.38	
MY	-0.12	
HC	-0.33	
FC	-0.47	
TS	-0.47	
TA	-0.15	
BB	-0.18	
CK	-0.22	

Table 3. Acceptable Hypocenter Locations.

Date	Origin Time	Error (sec)	Latitude	Error (km)	Longitude	Error (km)	Depth	Error (km)
*5-22-75 (2)	11:36:28.59	0.18	33.9135	1.15	106.9572	0.34	3.6	0.67
7-30-75 (34)	21:44:41.40	0.17	34.0760	0.38	106.9236	0.30	12.7	1.19
8-8-75 (31)	10:53:57.55	0.12	34.0482	0.49	106.9194	0.28	9.6	0.99
8-19-75 (26)	8:11:46.27	0.15	34.0440	0.28	106.9698	0.30	11.5	1.18
*8-19-75 (3)	10:00: 5.85	0.46	33.9551	1.23	107.0249	0.76	15.2	2.67
8-20-75 (33)	12:20:51.25	0.19	34.0749	0.40	106.9123	0.32	13.6	1.31
8-25-75 (32)	19:37:40.23	0.15	34.0682	0.35	106.9194	0.29	11.8	1.13
9-19-75 (13)	8:42:57.19	0.08	34.0032	0.40	106.8620	0.20	7.7	0.95
11-5-75 (48)	14.35: 4.24	0.38	34.0090	0.64	107.0956	1.45	13.4	2.10
11-7-75 (22)	13:52:34.77	0.18	34.0251	0.37	107.0930	1.76	6.3	1.09
1-21-76 (4)	14:18:27.96	0.20	33.9587	0.86	106.9595	0.32	8.5	1.31
1-23-76 (16)	2:53:32.71	0.29	34.0076	0.50	107.0418	0.76	10.8	1.97
1-23-76 (28)	7:22:14.37	0.19	34.0499	0.45	107.0470	0.65	6.1	1.65
1-27-76 (38)	8:37:43.36	0.53	34.1542	1.45	106.7963	1.33	15.7	2.92
1-29-76 (7)	18:24:27.27	0.16	33.9676	0.65	106.9899	0.31	8.5	1.16
4-13-76 (6)	11:58:34.35	0.12	33.9760	0.41	106.9713	0.35	6.5	0.79
4-16-76 (29)	9:33:42.54	0.15	34.0565	0.23	107.0212	0.34	7.8	1.20
4-20-76 (27)	2:52:19.50	0.36	34.0456	1.26	107.0705	1.47	9.0	1.14
*4-20-76 (36)	8:32:19.18	0.17	34.1069	0.57	106.8428	0.46	3.4	1.96
4-23-76 (8)	5:57:59.16	0.42	34.0327	1.60	107.0787	1.77	9.7	1.27
8-10-76 (19)	4:38:25.70	0.08	34.0109	0.45	107.0629	0.22	5.6	0.74
*10-6-76 (35)	15:12:40.83	0.33	34.0778	0.67	106.8017	1.32	13.4	1.76
10-7-76 (17)	23:21: 9.70	0.16	34.0076	0.86	107.0367	0.23	9.8	1.12
2-9-77 (20)	10:59:58.59	0.26	34.0173	0.56	107.0012	0.48	6.3	2.69
2-10-77 (37)	7:33:28.25	0.25	34.1230	0.95	106.9262	0.25	4.0	3.19
4-15-77 (24)	6:35:36.68	0.12	34.0371	0.57	107.0637	0.35	8.9	1.15
4-27-77 (18)	12:15:56.23	0.10	34.0011	0.48	107.0643	0.27	8.4	0.73
4-27-77 (21)	12:23:27.18	0.08	34.0178	0.33	107.0575	0.23	7.5	0.75
6-2-77 (9)	6:50:24.20	0.09	33.9993	0.46	107.0665	0.27	7.4	0.69
6-3-77 (14)	3:49: 1.35	0.09	34.0065	0.45	107.0619	0.26	7.8	0.72
6-3-77 (43)	20:45: 2.77	0.16	34.2284	0.47	106.8973	0.47	10.9	1.95

Table 3. Continued

Date	Origin Time	Error (sec)	Latitude	Error (km)	Longitude	Error (km)	Depth	Error (km)
*6-3-77 (5)	23:01:18.86	0.15	33.9717	0.65	107.0086	0.29	8.5	1.19
7-14-77 (39)	10:00:32.48	0.05	34.1602	0.19	106.8688	0.19	7.3	0.58
7-14-77 (23)	20:24:16.50	0.09	34.0347	0.31	107.0523	0.22	9.1	0.79
7-19-77 (42)	6:16:54.87	0.10	34.1581	0.37	106.8744	0.29	4.1	1.54
8-17-77 (44)	2:20:20.18	0.13	34.2877	0.23	107.0506	0.27	11.0	1.86
8-18-77 (40)	9:30:13.35	0.10	34.1609	0.30	106.8760	0.26	4.8	1.28
8-19-77 (15)	9:28:22.57	0.12	34.0072	0.55	107.0675	0.31	9.6	0.92
8-24-77 (10)	11:22:35.40	0.11	34.0009	0.43	107.0539	0.28	9.8	0.87
9-1-77 (30)	18:20: 2.29	0.14	34.0582	0.32	106.7625	0.69	8.3	1.19
9-1-77 (11)	21:58:48.35	0.12	34.0025	0.45	107.0415	0.27	9.9	0.98
9-2-77 (45)	13:29:30.83	0.38	34.3206	1.46	106.7799	1.34	15.3	1.38
9-20-77 (25)	1:20: 8.62	0.10	34.0376	0.55	107.0509	0.23	10.4	0.85
9-21-77 (41)	6:19: 8.46	0.05	34.1658	0.25	106.8784	0.17	3.9	0.91
9-22-77 (47)	5:20:27.38	0.11	34.3506	0.37	106.8960	0.24	7.4	2.39
9-22-77 (46)	19:19:16.39	0.09	34.3329	0.23	106.8897	0.21	11.5	1.13

the 83 events observed to have significant S wave frequency attenuation, 42 were unable to be used because of either (1) a poor hypocenter location or, (2) insufficient stations in the array to produce a reliable location (two stations or less). Also included in Table 3 are the locations of five microearthquakes in which no attenuation of either the P or S waves occurs at any of the recording stations (dates marked by an asterisk).

THEORETICAL DISCUSSION

Suggested causes for attenuation of the S wave in microearthquakes are (1) variation of source parameters, (2) spectra variation with change in azimuth around the focus, (3) normal loss of the higher frequency components of the spectra because of the distance traversed, (4) station effects, and (5) travel through an anomalous volume of the crust.

It is well known that earthquake spectra are a function of certain source parameters. Among the most important of these are fault size (dislocation area) and velocity and acceleration of rupture. In addition, magnitude and stress drop can be calculated from an analysis of earthquake spectra. For a complete discussion of these parameters see Brune (1970), Knopoff and Mouton (1975), and Randall (1973). For the same source model, however, the spectra for the P and S waves should be identical at the time rupture occurs. Thus if an abnormally low dominant frequency is observed in the S wave it should also be present in the P wave if source parameters are the only cause. Events in which P and S have similar frequency content are shown in the photographs of Plate 3. The effect of source parameters was eliminated in the data collection process by choosing only those events in which the frequency content of the S wave was substantially lower than the P wave.

SEPTEMBER 3, 1976 SC
13:25 UST

AUGUST 24, 1977 CM
11:22 UST

APRIL 15, 1976 SC
18:28 UST

APRIL 13, 1976 IC
9:45 UST

Plate 3. Examples of events in which the P and S waves have similar frequency content.

Variation of the spectra with change in azimuth around the focus is the second possibility. Due to the unequal distribution of energy propagated by the P and S radiation lobes (double-couple point source model) there could conceivably be some variation in the S wave spectra which would not necessarily be duplicated by the frequency content of the P wave. Douglas, Ryall and Williams (1971) performed a study on the variation of spectra with azimuth for microearthquakes in central Nevada. They averaged groups of spectra to eliminate peculiarities in individual records and considered events at approximately the same hypocentral distance. For the lengths and amplitudes of the seismic signals analyzed they concluded that the S wave contributed most significantly to the Fourier amplitude spectrum. They found, for six separate event groups, representing an azimuth range of approximately 140° , that there was no systematic variation of the spectra.

The third consideration is the normal loss of high frequency observed for raypaths traversing long distances through geologically complex volumes of crust. It is well documented in the literature (Sanford and Holmes, 1962, Douglas, Ryall, and Williams, 1971) that such losses occur and that the S wave, due to its transverse mode of propagation, is more susceptible to loss of the higher frequency spectral components than the P wave. However, from observations of the wave forms of microearthquakes arriving at stations from great (>30 km) distances it is quite apparent that the S wave suffers little of the unusual frequency attenuation of the type present in the data used in this study. A list of events of this type is presented in Table 4. In addition, an average dominant frequency was obtained for unattenuated microearthquakes, as they appear at stations FM, CC, WT and SC, from a zero-crossing counting

Table 4. Events with Unattenuated Arrivals at Distant (>30 km) Stations.

Date	Event	Origin Time	Approximate Hypocentral Distance (km)	Recording Station
9-19-75		8:42:57.19	38.7	MG
			36.1	HC
11-5-75		14:35: 4.24	41.2	MY
6-3-77		20:45: 2.77	30.6	LAD
			33.2	CM
7-14-77		10:00:32.48	34.5	GM
			37.3	LAD
7-19-77		6:16:54.87	33.5	GM
8-18-77		9:30:13.35	33.5	GM
9-1-77		18:20: 2.29	45.4	GM
9-20-77		1:20: 8.62	46.4	RI
			49.9	TD
9-21-77		6:19: 8.46	41.8	RI
9-22-77		5:20:27.38	31.3	RI
			32.2	TD
9-22-77		19:19:16.39	39.7	GM
			33.0	RI

analysis of a random selection of events recorded on the digital DR-100 units. This data is presented in Table 5. These results were considered typical of microearthquake activity in the Socorro area. The events listed in Table 5 were considered 'normal' for the area in that they possess the frequency requirement for whiteout on an MEQ-800 record. The dominant frequency found for the P wave was 19.7 Hz (s.d. 1.5 Hz), the S wave was 12.3 Hz (s.d. 0.3 Hz). Considering the wide range of hypocentral distances and amplitudes included in this analysis, the standard deviations for the dominant frequencies are quite small. This also indicates that frequency attenuation in the S wave is not strongly dependent on the distance, alone, traversed in the crust of the Socorro area.

The fourth possibility is that anomalous S wave frequency attenuation occurs in the near surface geological column directly beneath the stations. Whenever possible, the stations were located on solid rock off the deep alluvial fill present in much of the La Jencia and Socorro basins, so that this effect should have been minimized. However, the station corrections which appear in Table 2 indicate that there may be substantial differences in geology beneath the stations. To determine if this was a factor influencing the spectra of the S wave, seismograms showing unattenuated S waves, for a varied azimuth range, were found for each station showing anomalous S wave attenuation, except BG. The special case for the station BG will be discussed under conclusions. These unattenuated raypaths often travelled through the same station material the attenuated raypaths traversed. Stations for which this occurred are CM, SC, FM, DM, CC, GM, and NG. A list of the events which were sources for these raypaths, and the stations at which they were observed, unattenuated, is

Table 5. Data for Determination of Unattenuated
Dominant Frequencies.

Event		Recording Station	Dominant P Wave Frequency	Dominant S Wave Frequency
Date	Approximate Origin Time			
5-30-77	6:55	SC	19.56	12.11
5-30-77	7:01	CC	17.77	12.42
5-30-77	7:25	SC	23.10	12.11
5-31-77	5:48	SC	18.73	11.83
6-2-77	6:50	SC	18.39	11.50
6-3-77	3:49	SC	19.60	12.49
8-9-77	9:50	WT	17.46	10.97
8-11-77	9:54	CC	22.32	12.70
8-17-77	12:08	CC	17.42	12.29
8-17-77	14:14	CC	19.53	11.68
8-18-77	8:11	CC	19.11	12.16
9-1-77	21:59	WT	18.06	12.47
9-22-77	5:20	WT	19.55	11.95
10-18-77	8:16	SC	20.71	13.48
11-15-77	19:02	FM	23.99	14.13

given in Table 6. Assuming, more generally, that the geological section beneath each station is approximately uniform azimuthally near (~ 0.50 km) the station, then the highest dominant frequency S wave observed passing through that section should be close to the value observed for every incoming travel path. This assumes the ray has not already passed through an anomalously attenuating volume. If this is a valid approximation, then the observation of just a few unattenuated raypaths at a station is sufficient to justify minimizing station effects as the cause of the anomalous S wave attenuation. Furthermore, the criterion that the P wave be high frequency when attenuation is observed in the S wave is also beneficial in eliminating station effects, as one would expect some attenuation of the P wave to occur for such great attenuation in the S wave. This would be particularly true for attenuation due to the static frictional dissipation encountered in granular media. The manner in which the data was collected precludes this possibility. Thus, station effects alone do not seem to be able to account entirely for the frequency attenuation observed in the S wave.

The fifth possibility is an anomalous region in the crust through which the raypath traverses, which causes the observed frequency attenuation in the S phase. Walsh (1969) has theoretically analyzed a model of partially melted rock. The model consists of the melt occurring as thin films along grain boundaries. The melt thus makes up a very small fraction of the total volume of the rock. Walsh derived expressions for the complex bulk modulus and rigidity. He found that rigidity and attenuation in pure shear depend on the number of sites at which melting has occurred rather than the total concentration of the melt. Response in pure dilatation, though, is a function of concentration alone. The

Table 6. Unattenuated Raypaths for Determination of Station Effects.

Event		
Date	Approximate Origin Time	Recording Station(s)
5-22-75	11:36	CC, CM
7-30-75	21:44	CM, CC
8-8-75	10:53	FM
8-19-75	8:11	SC, CC
8-19-75	10:00	CM, SC
8-20-75	12:20	CM, SC
8-25-75	19:37	CC
9-19-75	8:42	GM
11-7-75	13:52	CC
1-23-76	2:53	SC
4-20-76	8:32	DM
4-23-76	5:57	DM
8-10-76	4:38	NG
10-6-76	15:12	DM
2-9-77	10:59	NG
2-10-77	7:33	CC
4-15-77	6:35	GM
4-27-77	12:23	CC
6-2-77	6:50	GM
6-3-77	20:45	CM
6-3-77	23:01	CM
7-14-77	10:00	GM
7-19-77	6:16	GM
8-17-77	2:20	GM
8-24-77	11:22	NG
9-1-77	18:20	GM
9-1-77	21:58	NG

result is that attenuation in pure dilatation is low and bulk modulus is approximately equal to the value obtained for unmelted material. Thus for a partial melt of very low total concentration one would theoretically expect a great effect on the S wave spectra but little or no change in the P wave spectra.

Due to the difficulty of reproducing the conditions of a partial melt in the laboratory direct experimental verification of the results outlined above is lacking. However, Spetzler and Anderson (1968) were able to study the effects of another type of partial melt on attenuation in longitudinal and shear waves. They investigated the attenuating properties of a simple binary system of $\text{NaCl} \cdot \text{H}_2\text{O}$. Research was conducted in the range from a complete solid at low temperature to a 17 percent melt at higher temperatures. As a solid the system was a mixture of H_2O (ice) and $\text{NaCl} \cdot 2\text{H}_2\text{O}$. At higher temperatures the system was a mixture of ice and NaCl brine. In the completely solid regime, seismic velocities and Q (quality factor) changed slowly with increasing temperature. At the onset of melting, though, there was a marked decrease in both seismic velocities and Q. The authors concluded that a small fraction of liquid has a large effect on seismic velocities and Q, and that the shear wave was considerably more affected than the longitudinal wave. In addition, the most dramatic and pronounced effects occurred at the onset of melting rather than in melts with higher percentages of fluid. These conclusions, obtained experimentally, are remarkably similar to the theoretical results obtained by Walsh (1969).

A study more closely related to the situation believed to exist in the Socorro area is the work done by Matumoto (1971) in the vicinity of Mount Katmai, Alaska. He observed the absence, or almost complete

absence, of the S wave for certain microearthquakes recorded in that region. Shuleski (1976) observed a similar effect in the Socorro area. The S wave screening observed by both Matumoto and Shuleski was explained by the passage of the raypath in which screening was observed through a magma body or chamber. Matumoto also observed some attenuation of the higher frequencies in the P wave spectra for the cases in which the S wave was completely absent. This also appears to be the case, to a lesser degree, for some of the events used in this study. In some cases, parts of the P wave would not always completely fulfill the whiteout criterion discussed above. This indicates that, for the more severe cases of S wave attenuation, the P wave is also affected to a lesser extent. The effect was never so great, however, that an accurate measurement of the dominant frequency of the P wave could be measured by the zero-crossing counting technique on the MEQ-800 records.

From the discussion above it is apparent that a partial melt of very low liquid concentration has an abrupt and significant effect on shear waves. It is possible that the magnitude of the attenuation observed in the shear waves of microearthquakes in the Socorro area is not severe enough to warrant postulation of a partial melt. Instead, the anomalous regions may be volumes of very hot rock at a temperature well below the liquidus. This could result in a less severe absorption of the seismic energy of the shear wave than would be the case if the S wave traversed a volume of partial melt. Evidence for this possibility is indicated in a paper by Solomon (1972). The relevant results are presented in Figure 5. The increase of Q , first seen in the samples for moderate temperatures, is probably the result of a loss of moisture. However, at higher temperatures (approximately 450°C to 500°C) the trend abruptly reverses

Figure 5. Q as a function of temperature, at surface pressure, for various rock types (from Solomon, 1972).

and a significant decrease in Q occurs. The effect is observed up to the last temperature plotted (600°C) for all samples. The experiment was conducted at surface pressure. Extrapolation of these results to higher temperatures could lead to a value for Q which explains the absorption of seismic shear wave energy in the Socorro crust.

From the results obtained in the studies discussed above it appears that a low value of Q for shear waves propagating in the vicinity of a partial, or near partial, melt leads to a selective loss of the higher frequencies in the amplitude spectra. This yields a decrease in the dominant frequencies observed on the seismograms as measured by the zero-crossing counting technique. From the magnitude of the decrease observed in the S wave, a crustal anomaly that could be postulated is a volume of heated rock whose temperature is somewhere below the liquidus.

INTERPRETATION

Primarily, the interpretation of the data consisted of mapping the raypaths along which attenuation of high frequencies occurred in the S phase. The volumes of crust defined by the intersection of these raypaths were then identified as possible regions of heated rock. The additional stipulation that these regions be free of hypocenters was also considered in their final placement. The high temperatures in a volume of heated rock would preclude the buildup of the elastic strain energy required for microearthquakes to occur. The high temperatures are thus assumed to promote stable sliding rather than stick-slip movements in the volumes of heated rock (Brace and Byerlee, 1966).

Also considered in the analysis were those raypaths along which no frequency attenuation in either the P or S waves was observed. Theoretically, these raypaths should not cross the regions of heated rock and

were thus used to eliminate anomalous areas which appeared at more than one location along a raypath.

A three-dimensional scale (0.392 in. = 1 km) model was constructed to aid in the spatial placement of the volumes. The general layout of the model may be seen in the photographs of Plate 4. The model consists of a styrofoam base (2 in. x 2 ft. x 2 ft.) with four plexiglass rods mounted perpendicular to the base at each of its four corners. Supported horizontally on these rods is a plexiglass sheet (0.25 in. x 2 ft. x 2 ft.) with its top surface parallel to the styrofoam base, 7.84 in. (20 km) above the base. The top surface of the plexiglass sheet is a map representing an area which extends approximately 33.7 km W, 27.8 km E, 16.5 km S, and 44.8 km N of Socorro (total area: 3748 km²). Marked on this map are the major towns, seismic stations used (two and three letter designations), and the epicenters (numbered). Below each epicenter is a black polystyrene bead (average diameter: ~0.25 in.) mounted on a 2 mm diameter glass rod supported in the styrofoam. The distance from the top of each bead to the top surface of the plexiglass sheet is the model depth for the respective hypocenter. The size of the beads gives a rough idea of the error bars associated with a typical hypocenter location. See Table 3 for the actual error bars associated with a particular hypocenter. Attached to the tops of the beads are threads representing seismic raypaths. The threads are of three different colors, representing three separate groups into which the data was subdivided (1) blue, raypaths where both P and S waves are unattenuated, (2) yellow, raypaths where the P wave is unattenuated but the S wave is ($8 \text{ Hz} \leq v_s < 12 \text{ Hz}$), and (3) red, same as (2) but the range for dominant S wave frequency is $v_s < 8 \text{ Hz}$. The threads are attached to the stations at which raypaths

Plate 4. (Continued) Side views of the three-dimensional model.

exhibiting criteria (1), (2), or (3) were observed. The data was subdivided into groups (2) and (3) to attempt to define portions of the anomalous volumes which were of higher temperature than others. It was thought this would also indicate the complexity of the shapes of the attenuating volumes.

The primary advantages of using the three-dimensional model rather than a simple two-dimensional surface projection for interpretation are (1) regions can be defined which appear to overlap an area of epicenters at the surface, (2) minimum depth to the region is easily defined, and (3) interpretation of raypaths of type (1), which pass close to raypaths of type (2) and (3), is easier.

The main objective of interpretation using the model was to explain as many of the attenuated raypaths as possible with a minimum number of volumes. Pieces of aluminum foil were used to experiment with placement of the volumes as they were readily able to be moved about and introduced into the spaces between the thread. It became immediately obvious during placement that the attenuating volumes are not of simple shape but highly irregular. Raypaths of type (2) and (3) crisscross the zones with no obvious systematic spatial variation. Often, type (1) raypaths cross near a clustering of type (2) and (3) raypaths, further indicating that not only are the volumes complex in shape, but the boundary between the attenuating volume and country rock is probably quite sharp. Some of these features may be seen in the photographs of Plate 4. The final approximate locations of the attenuating volumes are represented by the placement of the eight pieces of aluminum foil. The long stems attached to the upper, broader portions of foil have no physical meaning, but only serve to support the upper part of the foil in the correct position.

Eight separate volumes have been proposed to account for all but three of the attenuated raypaths. Two of these raypaths cross areas in which there is otherwise very little data. They are common to a region approximately 7 km southeast of the station SL, at a depth near 6 km, but there is not enough data to postulate an anomalous volume. The third raypath (from hypocenter (37) to DM) also crosses a region lacking in data. No attempt has been made to explain its presence. It is interesting to note, though, that the error bar for the depth in the location of event (37) is unusually large (3.19 km), indicating that whatever is causing the absorption may exist at a greater depth than appears to be the case from the depth listed in Table 3 for the hypocenter.

The locations (longitude, latitude, and depth) of the eight proposed volumes, as well as the number, type, and hypocenter origin of the raypaths used in the interpretation of each volume, are listed in Table 7. The epicenters of the events listed in Table 7 are plotted in Figure 6. Unattenuated raypaths used in the interpretation are listed as well. Figures 7 and 8 show maps of the locations of the attenuating volumes relative to the epicenters of the study, the surface projections of the attenuated raypaths, and the general physiographic features of the Socorro area. The latitudes, longitudes, and depths given in Table 7 roughly locate the shallowest points of the tops of the anomalous zones. In Figure 7, type (2) raypaths indicate potentially lower temperature rock than type (3) raypaths.

CONCLUSIONS

An alternate interpretation to the heated rock theory to explain the absorption discussed above is suggested by the work of Wallace (1978). He relates microearthquake swarm activity to magmatic intrusion in the

Table 7. Locations of Proposed Attenuating Crustal Volumes and the Defining Raypaths.

Volume*	Longitude	Latitude	Depth (km)	Raypaths Used for Volume Location		
				Hypocenter Origin**	Type	Recording Station
1	106.8591	34.1734	2.9	47	2	FM
				46	2	FM
				41	2	BG
				40	2	BG
				42	3	BG
				39	2	BG
				43	3	DM
2	106.9194	34.9573	3.2	34	2	FM
				32	3	FM
				33	2	FM
				20	3	DM
3	106.9096	33.9988	3.4	31	3	CM
				38	3	CM
				13	3	CM
				30	2	NG
				13	1	HC
				13	1	MG
4	106.9610	33.9686	3.6	7	3	CM
				4	3	CM
				6	3	CM
				34	1	CM

Table 7. Continued

Volume*	Longitude	Latitude	Depth (km)	Raypaths Used for Volume Location		
				Hypocenter Origin**	Type	Recording Station
4	106.9610	33.9686	3.6	33	1	CM
				26	2	CM
				5	1	CM
				43	1	CM
5	107.0132	34.0310	4.6	9	2	DM
				14	2	DM
				28	3	CM
				29	3	CM
				16	3	DM
				17	3	DM
				16	1	WT
				17	1	WT
				20	1	SC
6	107.0483	34.0314	7.1	13	1	HC
				21	1	CC
				21	1	WT
				18	2	CC
				10	2	CC
				23	2	CM

Table 7. Continued

Volume*	Longitude	Latitude	Depth (km)	Raypaths Used for Volume Location		
				Hypocenter Origin**	Type	Recording Station
7	107.0651	34.0496	8.7	25	2	GM
				27	2	CC
				8	3	BG
				27	1	WT
8	107.0752	34.0156	4.1	22	3	CM
				11	2	SC
				3	1	SC
				15	3	SC
				21	2	SC
				9	1	GM
				48	3	CM
				24	2	SC
				19	1	HC
				19	2	SC
				16	1	SC

* See Figure 7 for crustal volume coding

** See Figure 6 for hypocenter coding

Figure 6. Hypocenter coding used in Tables 7, 8, and in the text.

Figure 7. The positions of the eight proposed attenuating volumes of crustal rock relative to the attenuated raypaths, and the epicenters.

Figure 8. The positions of the eight proposed attenuating volumes of crustal rock relative to the physiographic features of the Socorro area.

Socorro area. He concludes that the model most likely to explain seismic swarm activity and an associated high b-value (1.38) starts with microfracturing of the crust due to magma injection, possibly to dilatancy, then strain release occurring at some later time due to the circulation of hydrothermal solutions migrating to the zone of microfracturing. If the volume of microfractures makes up a significant percentage of the total volume of the rock, and the fractures are fluid-filled, then it is possible that this could cause the type of absorption observed in the S waves in this study. These volumes would still approximately define the possible position for partial melts, as the microfracture zones would probably occur above, and adjacent to, a partial melt. They would then appear, as would the volumes in the heated rock theory, at a depth shallower than the partial melt.

Wallace further concludes that microearthquake swarms, for the most part, will be confined to crustal regions adjacent to shallow magma bodies and that the high b-value indicated above reflects structural heterogeneity. Figure 9 is a map locating the swarms used by Wallace relative to the surface projections of the eight proposed attenuating volumes. In many instances, there is good correlation between the swarms and attenuating volumes in that the swarm activity often appears adjacent to the anomalous volumes. The depth of the swarms is generally deeper than the anomalous volumes. The diffuse nature of the swarm activity in the southern La Jencia basin indicates a great structural heterogeneity which may accompany the presence of the large number of attenuating volumes proposed in that area. Only one swarm appears to be associated with the volume appearing southwest of the station BG. This could be the result of a data set which is less inclusive than the one used in this study.

Figure 9. The positions of the eight proposed crustal volumes relative to the microearthquake swarms located by Wallace (1978).

Wallace used data from June, 1975 to June, 1977. The seismic arrays were moved farther north soon afterward, enhancing the collection of data in the area around the station BG.

Of the three theories presented to explain the anomalous absorption occurring in the S wave, (1) partial melting, (2) heated rock, and (3) fluid-filled microfractures, the second appears to be the most probable. The observed absorption may not be great enough for a partial melt to be the cause, and the percentage volume of fluid-filled microfractures may not be great enough for this to be a cause. As is pointed out in the Appendix, absorption of energy in ordinary rock matrix pore space, filled with fluid, is insignificant for seismic frequencies (<100 Hz). If the amount of fluid present does not make up a significant portion of the rock volume, it will have little affect on absorption of seismic shear wave energy.

One of the primary assumptions made in the interpretation above is that dominant frequency is directly proportional to the quality factor, Q . Thus a decrease in dominant frequency has been associated with a lower Q value and increased absorption. A relationship between Q and dominant frequency, ν , has been proposed and justified in the Appendix. The relevant relation is,

$$Q_{\alpha}/Q_{\beta} = \nu_{\alpha}/\nu_{\beta} \quad , \quad (2)$$

where the subscripts α and β refer to P and S waves, respectively. The value of this ratio, using the unattenuated dominant frequencies determined from the data in Table 5, is 1.60. An increase in this ratio should indicate that Q_{β} is decreasing relative to Q_{α} and thus the absorption of the shear wave is increasing. Table 8 lists all attenuated ray-paths with their associated Q_{α}/Q_{β} ratios. In the calculation of the ratio,

Table 8. Attenuated Raypaths and Associated Q_α/Q_β Ratios

Raypath		Q_α/Q_β
Hypocenter Origin*	Recording Station	
34	FM	1.71
33	FM	2.05
31	CM	3.85
26	CM	2.11
32	FM	3.24
13	CM	2.55
38	CM	3.22
6	CM	2.96
20	DM	4.02
37	DM	3.96
21	SC	2.37
14	DM	1.89
43	DM	2.66
39	BG	2.43
48	CM	3.17
4	CM	2.55
28	CM	4.22
7	CM	3.08
29	CM	2.59
27	CC	1.80
8	BG	2.91
19	SC	1.70
24	SC	2.00
18	CC	1.82
9	DM	1.94
42	BG	2.51
15	SC	1.75
46	FM	2.25
41	BG	2.27
25	GM	2.33

Table 8. Continued

Raypath		Q_α/Q_β
Hypocenter Origin*	Recording Station	
22	CM	3.09
45	CC	2.22
40	BG	2.40
44	DM	3.36
23	CM	2.40
47	FM	2.45
10	CC	1.90
17	DM	2.89
16	DM	2.71
11	SC	1.97
30	NG	2.31

*See Figure 6 for hypocenter coding

the dominant frequency of the P wave was assumed constant, within one standard deviation, of the value 19.7 Hz. This was considered a good approximation for P waves fulfilling the whiteout criterion. The larger the ratio, the greater is the absorption of S wave relative to P wave energy and, presumably, the hotter is the material the associated ray-path has traversed. This does not alter the results obtained above but serves to relate those results to a more appropriate measure of absorption, Q , appearing in the literature reviewed in the theoretical discussion section.

By averaging the raypath values of Q_α/Q_β which define a particular volume, it is possible to assign a mean relative absorption factor to that volume. Table 9 lists the volumes determined in the interpretation section and their respective mean Q_α/Q_β ratios. These values are possibly low as an average for the entire volume, because the raypaths used to locate the attenuating volumes probably define their boundaries. S waves traversing the center, hotter portions of the volumes are likely to be completely screened. This argument primarily applies to the partial melt, or perhaps heated rock theory, to explain the anomalous volumes. If the absorption is due to fluid-filled microfractures, the Q_α/Q_β ratios will be closer to the true average for the entire volume as there should be little change in the absorbing characteristics of the center of the region relative to its boundaries. S wave energy should thus be able to propagate through all parts of the volume. Whatever the cause, the values listed in Table 9 are directly proportional to the amount of absorption occurring within the respective anomalous volume.

It should be emphasized that a good deal of potentially valuable data had to be eliminated due to poor hypocenter locations. It should

Table 9. Proposed Attenuating Volumes and Associated Q_α/Q_β Ratios.

Volume	Mean Q_α/Q_β	Standard Deviation
1	2.42	0.14
2	2.76	1.07
3	2.98	0.69
4	2.68	0.44
5	2.71	0.85
6	2.04	0.31
7	2.35	0.56
8	2.29	0.61

not be assumed, therefore, that all possible anomalous crustal volumes have been located, nor that the ones that have been located are accurate. The volumes which are probably related to shallow magmatic intrusion, and which are most accurately located, are (1) the region to the east of WT (depth: ~3.2 km), (2) the region to the northeast of SC (depth: ~4.1 km), and (3) the region to the northwest of WM (depth: ~4.6 km). The poorest established relation between a proposed anomalous volume and magmatic intrusion is for the region southwest of BG. Most of the raypaths which defined the volume were observed at BG, a station where the absorption effects of the local geology could not be accurately determined. For each of the attenuated raypaths, the P wave conformed to the whiteout criterion, but an S wave was never observed at BG which satisfactorily fulfilled the whiteout criterion. Control of station effects was thus not as complete as for the other stations. The spatial extent of intrusive activity near BG may be such that it is rare to observe unattenuated S waves from the areas of microearthquake activity located to date. It is for this reason, and the observation of additional raypaths traversing the volume which are seen attenuated at stations other than BG, that the volume has been included.

Future studies could reveal that many of the proposed attenuating volumes are related or connected to each other. To obtain a truly accurate and detailed model of the variation of Q within the crust a great deal more data is needed. Preferably, this data would be in digital form so that a detailed spectral analysis of each raypath could be made. This would allow the absolute determination of Q for both P and S waves for each raypath. The crust could be subdivided into cubes and the raypaths plotted within the subdivided volume. Each time a raypath crossed

a cube the cube would have that value of Q associated with it (see Figure 10). All Q values for each cube could be averaged and the standard deviation calculated. Those cubes with Q values possessing an acceptably low standard deviation would be assigned that Q value. Q could then be three dimensionally contoured, or anomalous volumes constructed from cubes with similar Q values. With a sufficient amount of data the entire analysis could be readily performed on a digital computer with a graphics terminal. The final output might be most revealing in the form of a lucid, three-dimensional plot in which all the cubes are shown marked in their centers with their respective mean Q values.

Figure 10. Top view of subdivided volume. The cubes take on the values of the raypaths crossing them.

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APPENDIX

LOSS MECHANISMS AND Q FOR COMPRESSIONAL AND SHEAR WAVES IN SHALLOW CRUSTAL MATERIALS

by

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Abstract

Five loss mechanisms have been reviewed as to their importance in absorption observed in shallow (<20 km) crustal materials for a frequency range of <100 Hz (1) solid friction, (2) viscous, (3) relative motion of pore fluid and rock matrix, (4) thermoelastic, and (5) nonlinear theory. Parameters (3) and (5) are insignificant, (1) dominates by far, while (2) and (4) are less dominant. The quality factor, Q, is found to be essentially independent of frequency ($Q = 213f^{-0.015}$, $10^{-3} < f < 10^7$, Attewell and Ramana, 1966). The relation $Q_\alpha/Q_\beta = v_\alpha/v_\beta$ (α, β refer to P and S waves, respectively; v is dominant frequency) is proposed. Q_α/Q_β is then 1.60 for the Socorro area as calculated from microearthquake data. It is shown how Q_α/Q_β can be used as a valuable constraint in earth modelling.

Introduction

The primary purpose of this paper is to review some of the different

mechanisms used to explain seismic absorption in shallow (<20 km) crustal materials. A realistic earth mechanism will then be justified to explain the absorption observed in the laboratory and field. Some idea of the magnitude of Q for the crust will also be determined. In addition, an effort will be made to formulate, as quantitatively as possible, a relation between the relative amounts of attenuation of compressional and shear waves in the range of frequencies of interest in seismology and seismic prospecting (<100 Hz). Finally, these ideas will be extended to the Socorro area and an indication of the degree of absorption normally occurring, estimated, by noting the frequency content of natural seismic activity (microearthquakes).

Attenuation in seismic waves is due to a variety of sources. The factors involved include geometrical spreading, energy loss at a reflecting boundary, dispersion, scattering, diffraction, and absorption. The attenuation to be discussed in this paper is concerned only with the last of these, absorption. Absorption is a function of composition of the material in which a seismic disturbance is propagating and indicates inelastic effects are operating within the material. Through absorption, the energy of a travelling wave is ultimately transferred completely to the propagation material in the form of heat, resulting in disappearance of the wave motion. If the effect of absorption can be measured accurately, and the additional quantities listed above accounted for, then the value obtained for the absorption could be used to identify the propagation material. Conversely, knowing the material composition establishes the absorption factor and provides a valuable constraint in various seismic investigations. This is the value of obtaining a sound theoretical and experimental understanding of the variation of absorption

with composition of propagation material.

A large number of interrelated quantities have been used to describe the degree of absorption. The most common of these is the quality factor, Q . It expresses the sharpness of resonance of a mechanical system and is defined by,

$$\frac{1}{Q} = \frac{2\Delta f}{f_n} \quad (1)$$

where f_n is the natural frequency of the system (see Figure 1).

Figure 1. Amplitude vs. frequency response for a mechanical system showing the parameters used in the definition of Q . (from White, 1965)

The quality factor may also be expressed,

$$\frac{2\pi}{Q} = \frac{\Delta W}{W} \quad (2)$$

where $\Delta W/W$ is the fractional strain energy loss per stress cycle.

Other expressions for Q include,

$$\frac{\pi}{Q} = \delta = V \frac{\alpha}{f} \quad (3)$$

where V is the velocity of the wave in the propagating medium, α is the absorption coefficient and f is the frequency of the wave. The quantity δ is the logarithmic decrement and may be defined by,

$$\delta = \ln \frac{A_2}{A_1} \quad (4)$$

where A_1 is wave amplitude and A_2 is the amplitude one cycle later. Another common expression is the absorption coefficient, given by,

$$A_1 = A e^{-\alpha x}, \quad (5)$$

where A and A_1 are the values of wave amplitude separated a distance, x . The absorption coefficient is α . In addition to these definitions other measures of attenuation are given by a and θ . The quantity, a , is simply referred to as attenuation and is given by,

$$\frac{a}{\pi f} = \frac{1}{Q}, \quad (6)$$

or,

$$\frac{a}{f} = \delta. \quad (7)$$

The factor θ is the phase angle between stress and strain and is,

$$\theta = \frac{1}{Q}. \quad (8)$$

Useful combinations of the above equations include,

$$\delta = \pi \theta, \quad (9)$$

$$\theta = \frac{a}{\pi f}, \quad (10)$$

and,

$$\frac{\Delta W}{W} = 2\pi \theta. \quad (11)$$

These are the quantities most commonly used in the literature to describe absorption. The parameter Q has tended to be utilized most frequently, however the logarithmic decrement, δ , facilitates comparison between different experiments and wave types (body and surface) better as it expresses a general loss observable in a wide range of phenomena. That is, amplitude loss per wavelength, expressed by δ , is readily

applied to any travelling wave. Thus, whenever possible, the various degrees of absorption will attempt to be stated in terms of Q or δ .

Loss Mechanisms and Experimental Results

Loss mechanisms may generally be classified into five groups (1) solid friction losses (intergranular sliding, sticking and sliding), (2) viscous losses (primarily associated with fluids, in this paper), (3) losses due to motion of a fluid in matrix pore space, (4) thermo-elastic losses, and (5) non-linear loss theory. To some degree, absorption in shallow crustal materials is probably a function of all these effects, as well as overburden pressure.

Solid friction and viscous losses are the best understood and have been investigated most thoroughly. They probably explain most of the absorption observed in the shallow crust. Figure 2 shows a stress-strain diagram for solid and viscous losses. For viscous losses, the plot is an ellipse whose area represents the energy lost per cycle. This loss is proportional to the cyclic stress frequency. For very low frequencies, the energy loss per cycle approaches zero and the ellipse degenerates to a straight

Figure 2. Stress-strain diagram, a_m is maximum stress, b_m is maximum displacement. (from Born, 1941)

line. If absorption is due to solid friction alone, the area of the

ellipse is independent of frequency and depends only on composition of the propagation material and the maximum strain amplitude. Thus, in the case of solid friction losses alone, δ is independent of frequency, while δ is proportional to the first power of frequency for viscous losses. Born (1941) has investigated the effects of frequency on δ for a dry and partially saturated sandstone. He used a resonant bar experimental procedure on an Amherst sandstone for a frequency range of approximately 500 to 3900 Hz. His results are summarized in Figure 3.

Figure 3. Decrement vs. frequency for dry and saturated Amherst Sandstone samples. (from Born, 1941)

The logarithmic decrement, δ , was found to be essentially independent of frequency for the dry specimen but varied linearly with frequency for various water saturations. These results indicate that the observed decrement is a function of both solid friction and viscous losses, primarily solid friction at low frequencies.

Knopoff and MacDonald (1958) reviewed experimental data on the dissipation of energy in rock-forming materials over the frequency range 10^{-2} to 10^6 Hz. and found that $1/Q$ was practically constant. This

observation also implies a solid friction loss mechanism for these materials at surface temperature and pressure. Knopoff and MacDonald (1960) also proposed a macroscopic model for the attenuation of small amplitude stress waves in solids in which solid friction varies as the gradient of local stress. They illustrated the model with a mass-spring system sliding on a rough surface in which the roughness of the surface increases in the direction the mass is moving. They found Q for the model to be independent of frequency. In addition, they cite experimental evidence which indicates a Q dependence on grain surface area in silicate aggregates. Both these observations suggest validity of the solid friction loss mechanism.

There is a large amount of experimental evidence for shallow crustal materials indicating that solid friction losses are of primary importance as an absorptive mechanism while viscous losses are much less a factor. Bruckshaw and Mahanta (1954) determined the variation of Young's Modulus, E , and fractional energy loss per cycle ($\Delta W/W$), with frequency in the range 40 to 120 Hz. They used rock specimens in the form of beams and set the system into forced vibrations. Their results are summarized in Figures 4 and 5. The plots are relatively constant over much of the frequency range, but deviate from linearity somewhat towards the lower frequencies, indicating a loss mechanism occurring which is not strictly of a linear, or solid friction, type. They found the greatest energy losses in the softer, less compacted and poorly indurated rocks. In particular, the value of $\Delta W/W$ for the sandstone seems high.

Usher (1967) performed a similar series of experiments for small beam-shaped rock samples over a frequency range of 2 to 40 Hz. His results are shown in Figures 6, 7 and 8.

Figure 4. Young's Modulus versus frequency. (from Bruckshaw and Mahanta, 1954)

Figure 5. Fractional energy lost per cycle versus frequency. (from Bruckshaw and Mahanta, 1954)

Figure 6. Young's Modulus versus Frequency. (from Usher, 1967)

Figure 7. Energy Loss versus Frequency. (from Usher, 1967)

Figure 8. Effect of water content on energy loss as a function of frequency. (from Usher, 1967)

These plots, as do Born's (1941), clearly show the dominant effects of solid friction, with viscous losses becoming more important at lower frequencies and higher percentages of water content. When fluid is forced from compressed to dilated regions energy is dissipated by viscous

means such that $\Delta W/W$ increases with frequency. Usher suggests modifying Knopoff and MacDonald's (1960) model of pure solid friction grain boundary slipping to include viscous slipping at lower frequencies, but pure solid friction at higher frequencies. Usher also points out that the trends seen in Figures 6 through 8 may not be as drastic for rocks in situ, due to increased hydrostatic pressure.

Horton (1959) formulated a loss mechanism for the Pierre shale in terms of a modified Kelvin solid. The model he used is shown in Figure 9. He then compared α (attenuation coefficient) versus frequency curves, calculated using the model, with experimental data obtained by McDonal, et al. (1958) for the Pierre shale. The results for shear waves are presented in Figure 10 and those for compressional waves in Figure 11. The theoretical and experimental data agree well for most parts of the curves and indicate that the model is a good approximation to the actual loss mechanism in the Pierre Shale. McDonal et al. (1958) obtained a value of $\alpha = 0.12 f$ (db/1000 ft.) from Figure 11 and $\alpha = 1.05 f$ from Figure 10. Figure 11 applies to vertically-travelling compressional waves over the frequency range 50 to 450 Hz., while Figure 10 includes data from horizontally-travelling shear waves with vertical motion in the frequency range 20 to 120 Hz.

Figure 9. Modified Kelvin solid. (from Horton, 1959)

Figure 10. Attenuation of shear waves versus frequency. The solid curve is calculated from theory. (from Horton, 1959)

Figure 11. Attenuation of compressional waves versus frequency. The solid curve is calculated from theory. (from Horton, 1959)

Peselnick and Outerbridge (1961) performed a torsion pendulum experiment at room temperature on the Solenhofen Limestone over a frequency range from 4 Hz to 10 MHz to determine internal friction in shear, and shear modulus. They found the shear modulus to be constant within ± 2 percent, at 2.64×10^{11} dynes/cm², over the entire frequency range. The results obtained for the logarithmic decrement, δ , are presented in Table 1.

Table 1. Solenhofen Limestone Data
(from Peselnick and Outerbridge, 1961)

Sample	Density, g/cm ³	Frequency, cps	Rigidity Modulus, dynes/cm ²	Shear Velocity, km/s	Logarithmic Decrement, δ	Remarks
B	2.67	3.89	2.58×10^{11}	3.1	$3.4-4.1 \times 10^{-3}$	(a)
A	2.67	6.52	2.59	3.1	4×10^{-3}	(b)
C	2.67	8,200	2.66	3.15	5	(c)
C	2.67	16,400	2.66	3.15	5	(c)
B	2.67	9,500	2.68	3.17	5	(c)
B	2.67	18,600	2.61	3.12	5	(c)
B	2.67	28,500	2.60	3.18	...	(c)
S-25	2.66	10×10^6	2.6	3.1	...	(d)
S-1	2.59	9×10^6	2.2	2.9	17	(d)

(a) 7.2 kg/cm² axial tension; maximum shear strain $\sim 7 \times 10^{-6}$ (precision in $\delta = \pm 5$ per cent).

(b) 4.3 kg/cm² axial compression; maximum shear strain $\sim 5 \times 10^{-6}$ (precision in $\delta = \pm 5$ per cent).

(c) Bar resonance technique (precision in $\delta = \pm 10$ per cent).

(d) Pulse-echo technique (precision in $\delta = \pm 15$ per cent).

These results also indicate a solid-hysteresis behavior acting predominantly as a loss mechanism. A possible cause was discussed above in the review of Knopoff and MacDonald's (1960) work.

An interesting compilation of data from a variety of authors and experimental methods has been made by Attewell and Ramana (1966). Their results are summarized in Figure 12.

Figure 12. Attenuation coefficient versus frequency for a wide variety of authors and experiments. (from Attewell and Ramana, 1966)

From 26 authors, representing 168 points, they determined the following value for the absorption coefficient,

$$\alpha = 1.99 \times 10^{-6} f^{1.0386} \text{ db/cm} ,$$

for a frequency range of 10^{-3} to 10^8 Hz. This implies a Q which is practically independent of frequency, supporting the general arguments concerning solid friction losses discussed above.

Figure 13 shows a less inclusive but similar set of data compiled by Hamilton (1972). He worked with saturated marine sediments but concluded that intergrain friction appears to be the dominant cause of wave energy damping. As can be seen from Figure 13, α appears to depend on the first power of frequency and thus Q is independent of frequency. The line marked A in Figure 13 is the result of data collected for sedimentary land strata at relatively low frequencies.

Attenuation versus frequency in natural, saturated sediments and sedimentary strata; data from Tables 1-4b. Symbols: circles-sands (all grades); squares-clayey silt, silty clay; triangles-mixed sizes (e.g., silty sand, sandy silt, sand-silt-clay); sand data at 500 and 1000 kHz from Busby and Richardson (1957). Low-frequency data: line A-Zemstov, 1969 (land, sedimentary strata); B-Tullos and Reid, 1969 (Gulf of Mexico coastal clay-sand); C-Bennett, 1967 (sea floor, reflection technique); Z-Zhadin (in Vasil'ev and Gurevich, 1962); E and B-Epinatyeva et al, and Berzon (in Zemstov, 1969). Line labelled f^1 indicates slope of any line having a dependence of attenuation on the first power of frequency.

Figure 13. Compiled attenuation versus frequency. (from Hamilton, 1972)

The third type of loss mechanism is that due to the motion of fluid in pore space relative to the rock matrix. White (1965) has reviewed this mechanism which has been treated in detail by Biot. White utilized

Biot's results by substituting typical values for rock and fluid into Biot's equations for a_p and a_s , attenuation of compressional and shear waves, respectively. This resulted in values of,

$$a_p = 9 \times 10^{-12} f^2 \text{ sec}^2/\text{cm},$$

and,

$$a_s = 65 \times 10^{-12} f^2 \text{ sec}^2/\text{cm}.$$

These values should be valid to frequencies of approximately 20,000 Hz (White, 1965). However, for frequencies of interest in seismology and seismic prospecting (< 100 Hz), the values obtained for a_p and a_s are so much lower than typically measured values, that absorption due to this mechanism is probably insignificant (White, 1965).

Hamilton (1972), in his study of attenuation of compressional waves in saturated marine sediments, also concluded that viscous losses, due to the relative movement of pore water and mineral structure, are negligible. Thus, for the frequencies of interest in this paper, the absorptive losses due to this type of mechanism will not be considered important.

The fourth type of loss mechanism to consider is that due to thermoelastic effects. Since all real materials have a finite thermal conductivity, stress waves propagating through such material will lose energy due to heat conduction. Treitel (1959) has derived equations of motion and temperature for an elastic solid with a finite thermal conductivity using the irreversible form of the second law of thermodynamics. He obtained the following expression for the absorption coefficient,

$$\alpha = \frac{1}{8} \frac{Kw^2}{V_p^3}, \quad (12)$$

where K is thermal diffusivity, ω is angular frequency ($\omega = 2\pi f$) and V_p is compressional wave velocity. This expression implies a Q which varies as the first power of frequency. In this respect, this mechanism produces the same type of frequency dependent Q as the viscous loss mechanism. Figure 14 is a plot of the results of equation (12) using the following values,

copper: $\omega \ll 1.8 \times 10^{11}$ rad/sec

$K = 1.14$ cm²/sec

$V_p = 4.6 \times 10^5$ cm/sec

rock: $\omega \ll 2.5 \times 10^{13}$ rad/sec

$K = 0.01$ cm²/sec

$V_p = 5 \times 10^5$ cm/sec

Figure 14. α versus circular frequency for copper and rock with the properties described above. (from Treitel, 1959)

The last loss mechanism to be considered is non-linear processes. A nonlinear mechanism for absorption was initially proposed primarily to account for the observation that the Q spectrum for many solids

appeared to be flat (α depends precisely on the first power of frequency). Collins and Lee (1956), though, showed that by assuming a variation of α with f^{1+n} , where n is a very small number, a strictly linear theory could be applied. There is a large amount of evidence, some of which has been reviewed above, that the Q spectrum is not completely flat. Thus Collins and Lee's assumption appears to be experimentally valid. Several linear mechanisms have been proposed and discussed above which adequately describe the Q spectrum within the limits of the calculated Q responses, using the f^{1+n} assumption. It has also been pointed out by Savage and Hasegawa (1967) that a nonlinear mechanism should produce a cusping, or sharpening, of seismic pulses. Yet it is observed that seismic pulses are broadened, and cusps smoothed during transmission through shallow, crustal materials. Thus, it is not necessary to introduce the added complications of nonlinear loss theory.

Discussion

The primary result from the above observations is that absorptive loss in shallow crustal materials is mainly attributable to solid friction, in which Q is approximately independent of frequency. The slight observed deviations from this relation may be explained by losses due to either a viscous or thermoelastic mechanism. Losses of these latter two types produces a Q which varies approximately as the first power of frequency.

Various values of δ , α and Q , as determined experimentally above, can be used to give a rough indication of the average value of Q likely to be observed in near-surface crustal materials. A summary of various experimenters' results, which includes many of the works cited above, has been presented in Figure 12. Attewell and Ramana (1966) have calculated

an average Q for this data of,

$$\bar{Q} = 213f^{-0.015}, \quad 10^{-3} < f < 10^7.$$

The size of the exponent indicates that Q is practically independent of frequency. This value is probably too low for crustal material in situ. The pressure of overburden will substantially raise Q . Bath (1973) indicates a Q for the crust of approximately 470, while Press (1964) determined Q , from underground nuclear explosions in Nevada, to be 260 ± 40 . Thus a value around 300 may be more typical of crustal material in situ.

Assuming the effect of overburden pressure has an equal effect on Q_α and Q_β (Q for compressional and shear waves, respectively), a calculation of the ratio Q_α/Q_β can be made for absorption data taken at relatively shallow depths in the field. A study, discussed above, particularly useful in this regard is the determination of absorption for compressional and shear waves in the Pierre Shale, performed by McDonal et al. (1958). A summary of their results is presented by Bradley and Fort (Clark, 1966) and is reproduced in Table 2, after conversion to Q from percent energy lost per cycle. In calculating average values for Q_α and Q_β , and forming their ratio, the effects of the frequency at which the measurements were taken was not considered.

Table 2. Q_α and Q_β for the Pierre Shale

Q_α	Q_β
31.4	11.5
37.8	11.4
37.4	12.4
38.6	12.0
36.1	9.1
36.9	9.4
33.4	10.4
34.3	11.1
36.4	

The justification for this has already been established in the discussion above. Average values for the compressional and shear quality factors are,

$$\bar{Q}_\alpha = 35.8 \text{ and } \bar{Q}_\beta = 10.9 .$$

The ratio $Q_\alpha/Q_\beta = 3.28$. The measurements leading to these results were performed on the Pierre Shale, a formation known for its' homogeneous and isotropic nature, at depths ranging from 260 to 750 feet. Should this Q_α/Q_β for the Pierre Shale be considered a reasonable result applicable to different rock types at different depths? Bremaecker et al. (1966) measured similar data on an aa lava flow, unconsolidated cinders, and compact limestone, at the surface. By assuming a complex shear modulus but a real λ (Lamé constant) they determined a relation for Q_α/Q_β

$$Q_\alpha/Q_\beta = 0.5 \alpha^2/\beta^2 , \quad (13)$$

where α is compressional velocity and β is shear velocity. The justifications for the assumptions leading to equation (13) first start with the statement that imperfect elasticity can be expressed by making both μ (shear modulus) and λ complex. However, for sedimentary rocks, λ' (complex Lamé constant) approaches zero, resulting in equation (13) (De Bremaecker et al., 1966).

Now consider the values for α and β as determined by McDonal et al. (1958) for the Pierre Shale: $\alpha = 7100$ ft/sec, $\beta = 2630$ ft/sec. Substitution into equation (13) yields:

$$Q_\alpha/Q_\beta = 3.64 .$$

This should be compared with the ratio calculated from Table 2 of $Q_\alpha/Q_\beta = 3.28$. Considering the great range of environments, experimental conditions, and the assumption of equal overburden pressure effect on Q_α and Q_β , the

values are remarkably close. This suggests that equation (13) could have rather general applications to a wide range of rock types and depths of burial.

An attempt can now be made to obtain the ratio Q_α/Q_β for shallow (<20 km) crustal materials in the Socorro area by using equation (13). The values for compressional and shear wave velocities commonly used in this area are: $\alpha = 5.80$ km/sec and $\beta = 3.35$ km/sec. Substitution into equation (13) yields $Q_\alpha/Q_\beta = 1.50$. This value is considerably lower than the one obtained above for the Pierre Shale. The difference is probably the result of using body waves which have traversed a volume of deeper rock and thus greater overburden pressure. In near surface, unconsolidated rock, one might expect the shear wave to be affected to a greater extent than the compressional wave. This difference would tend to diminish as the waves reached better indurated rock at higher confining pressure. This is reflected in the lowering of the ratio Q_α/Q_β .

This value for Q_α/Q_β of 1.50 for the Socorro area can now be used to help indicate whether calculating the ratio, by using the dominant frequencies of P and S waves, is valid. The proposed relation is,

$$Q_\alpha/Q_\beta = v_\alpha/v_\beta, \quad (14)$$

where v_α and v_β are the dominant frequencies of the P and S waves, respectively. These values have been determined to be $v_\alpha = 19.7 \pm 1.5$ Hz and $v_\beta = 12.3 \pm 0.3$ Hz using a zero-cross counting technique for a group of 15 randomly selected microearthquake events. Thus $v_\alpha/v_\beta = 1.60$. This value, considering the standard deviations for v_α and v_β , is within the range of the ratio $Q_\alpha/Q_\beta = 1.50$ obtained with equation (13), indicating that equation (14) may be approximately valid.

The basic assumptions used in these determinations of Q have been (1) Q is independent of frequency, (2) the loss mechanism is primarily of the solid friction type, and (3) observed absorption is an approximately linear process. These assumptions have been justified in the preceding discussion.

Conclusions

It is well known that shear waves are less efficiently transmitted in bulk, in situ, crustal materials, than P waves. The value of $Q_\alpha/Q_\beta = 1.60$ obtained above reinforces this belief. If values calculated in this manner can be believed, and if the ratio varies with rock type, Q_α/Q_β could be used to determine the average composition of material through which a seismic disturbance propagates. By performing field experiments in which travel paths from seismic sources are plotted and have individual, calculated Q_α/Q_β ratios associated with them, volumes of crust might be defined which exhibit a particular value of the ratio. Ratios for different volumes could then be compared and relative material properties assigned to each volume. If relatively unique ratios can be found, or determined experimentally, for particular material compositions, then the rock types in these regions could be approximately identified. The effects of different depths of burial would have to be accounted for, but the correction could become rather small below a certain critical depth and above 20 km, such that the total variation of Q_α/Q_β with overburden pressure could be insignificant in this range (critical depth to 20 km). In addition, by using equation (13), relative velocities could be assigned to these crustal volumes and thus aid in determining a velocity model. This could be particularly valuable in an area which is geologically complex and does not readily conform to a model consisting of homogeneous

layers.

The approximate spatial variation of the ratio Q_α/Q_β , as determined by travel path plotting, can thus be a valuable constraint in the formulation of earth models of the shallow crust.

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