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INTERNAL STRUCTURE OF A MAGMA BODY
AT MID—CRUSTAL DEPTHS NEAR
SOCORRO, NEW MEXICO, DEDUCED FROM
MICROEARTHQUAKE REFLECTIONS

by
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ABSTRACT

The characteristics of a strong mid-crustal reflector in the Rio Grande rift southwest of Socorro, New Mexico, were studied using high-quality digital seismograms of microearthquakes. The data are from a tight cluster of hypocenters located almost directly beneath a recording station ($\Delta \approx 2.0$ km). The availability of both well recorded P and S phase reflections at one location provided a unique data set with which to investigate the existence and properties of the mid-crustal reflector. Cross-correlation of the direct and reflected S phases yields coefficients which are positive and large in absolute value (average value of 0.79). Because these two phases leave the focal region 180° out of phase for this swarm, reflection must occur at a discontinuity where a 180° phase shift occurs. Waveforms for the P wave reflections are not identical from event to event within the swarm. The character of the P reflection indicates a magma body with rapid lateral changes in internal structure. The general spectral characteristics of the reflected P phase are suggestive of constructive interference.

The near normal incidence geometry of both the reflected and direct P phase implies an essentially common path through the upper crust and near surface Phanerozoic rocks for these two phases. The earthquakes examined in this study are all quite small ($M_L < 0$), hence, the initial portion of the direct P phase could be used as the response function for the path through the upper crust during modelling of the reflected P phase. Synthetic seismograms, which included attenuation effects and interbed multiples, were calculated and compared to the

observed waveforms. Forward modelling of the P_zP phase supports the hypothesis that the magma body is quite thin yet has resolvable differences in internal structure over the areal extent of the reflection points covered by the swarm.

The results of modelling the stacked spectra of the P_zP phase indicate the most appropriate model of the magma body to be a thin ($\approx 70\text{m}$) layer of non-rigid, low-velocity material underlain by a second, thin ($\approx 60\text{m}$) layer of slightly higher velocity material. The lower layer may be an earlier, partially solidified intrusion or a crystalline mush at the bottom of a single magma chamber. The total thickness of the magma body is estimated to be less than 160m.

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INTRODUCTION

Purpose and Scope

The presence of magma at middle or upper crustal levels has been established in only a few locations. For a variety of reasons, previous studies have been constrained to establishing the existence of magma accumulations rather than providing details on the structure of the bodies. This study has been able to provide details on at least the local structure of an extensive magma body.

The Rio Grande rift within central New Mexico is an area with significant and unique geophysical properties. Many of these properties are related to or influenced by the emplacement of magma bodies within the crust. In particular, an extensive mid-crustal magma body has been identified in the Socorro area using a variety of techniques. Although a great deal of progress has been made to date in establishing the existence of this magma body, specific questions regarding the structure of the body important to the understanding of magmatism in the Rio Grande rift remain unanswered.

This research project involved detailed studies of digitally recorded microearthquakes located within the rift. The primary objective was the development of a model for the internal structure of the mid-crustal magma body. Specifically, we wished to know if the magma occurs in thin layers interbedded with solid rock as suggested by Brocher [1981] or in a simple, single layer as proposed by Rinehart et al. [1979]. Secondary objectives were (1) to develop a procedure by which various factors such as path, source, and reflector effects could

be isolated, (2) to determine the degree to which a unique interpretation could be obtained, and (3) to estimate the seismic quality factor (Q) of the mid-crust in this area.

The primary objective is of significance as it may help resolve fundamental questions regarding the probable origin and mechanism of emplacement of magma at mid-crustal levels. The secondary objectives are of significance in understanding the influence of near surface structure on recorded seismograms, on understanding how "robust" the waveform data actually is, and on isolating estimates of mid-crustal Q from "contamination" by low- Q , near surface materials.

General Outline of Method

In this study, the structure and characteristics of the Socorro mid-crustal magma body were determined from analysis of microearthquake seismograms recorded at very small epicentral distances. Characteristics of various phases on the seismograms were analyzed and a model of the magma body structure proposed. This estimate of structure was refined by forward modelling with synthetic seismograms. The modelling procedure also allowed an estimate of the seismic quality factor, Q , to be obtained for the mid-crust in the Socorro area.

Prior to the computation of the synthetic seismograms, upper-crustal effects were approximated by the direct P-phase and the assumption of source independence justified. In addition to the modelling of individual waveforms, the spectra of several events were stacked. The resulting average spectrum was then compared with synthetic spectra. One, two, and three layer models were compared to

the observed data. The resulting best-fitting model was then compared to models of mid- crustal geology as inferred from other studies. The thermal consequences of the resulting best-fitting model were also briefly considered.

GEOLOGICAL AND GEOPHYSICAL SETTING

The study area is located within the central Rio Grande rift southwest of Socorro, New Mexico. The Rio Grande rift (RGR) is a roughly north-south trending series of en-echelon grabens that extend from near Leadville, Colorado on the north into northern Mexico on the south [Chapin, 1971]. The rift appears to be comprised of three separate segments: (1) the northern rift which ends near the Colorado-New Mexico border, (2) the central rift extending from the state line south to near Socorro, New Mexico, and (3) the southern rift extending from Socorro southward into Mexico. The Socorro portion of the rift is bounded by the Colorado Plateau, Datil-Mogollon volcanic field, and the Basin and Range province (Figure 1). Details on large scale aspects of the rift may be found in Chapin [1971], Chapin and Seager [1975], Sanford et al. [1977], Cordell [1978], and the volume *Rio Grande Rift: Tectonics and Magmatism* edited by Riecker [1979].

Physiographically the Socorro area consists of two basins, the Socorro and La Jencia basins, separated by a horst, the Socorro-Lemitar Mountains. The rift is bounded by the Loma de las Canas uplift on the east and the Magdalena Mountains on the west. Figure 2 shows the major physiographic features of the region as well as the current locations of stations in the local seismic network. Figure 3 is a simplified geologic map of the Socorro area from Carpenter [1984].

The known geological history of the Socorro region dates back to at least 1.7 bya, the age of the oldest rocks exposed in the area [Condie and Budding, 1971]. The general structural grain of these Precambrian rocks trends north-northwest and has probably influenced subsequent structural development [Chapin et al., 1978].

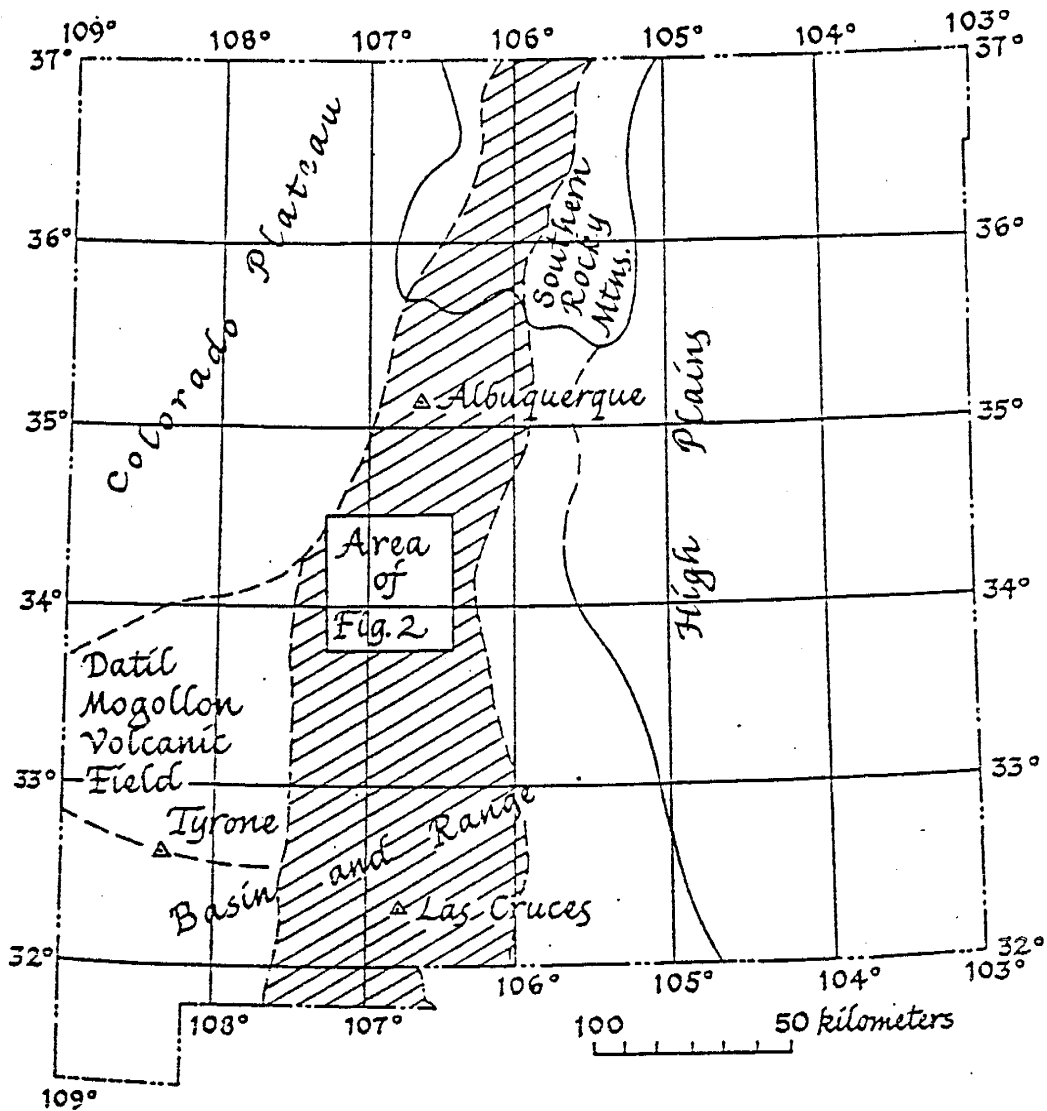


Figure 1. Physiographic provinces and the Rio Grande rift in New Mexico (after Chapin, 1971).

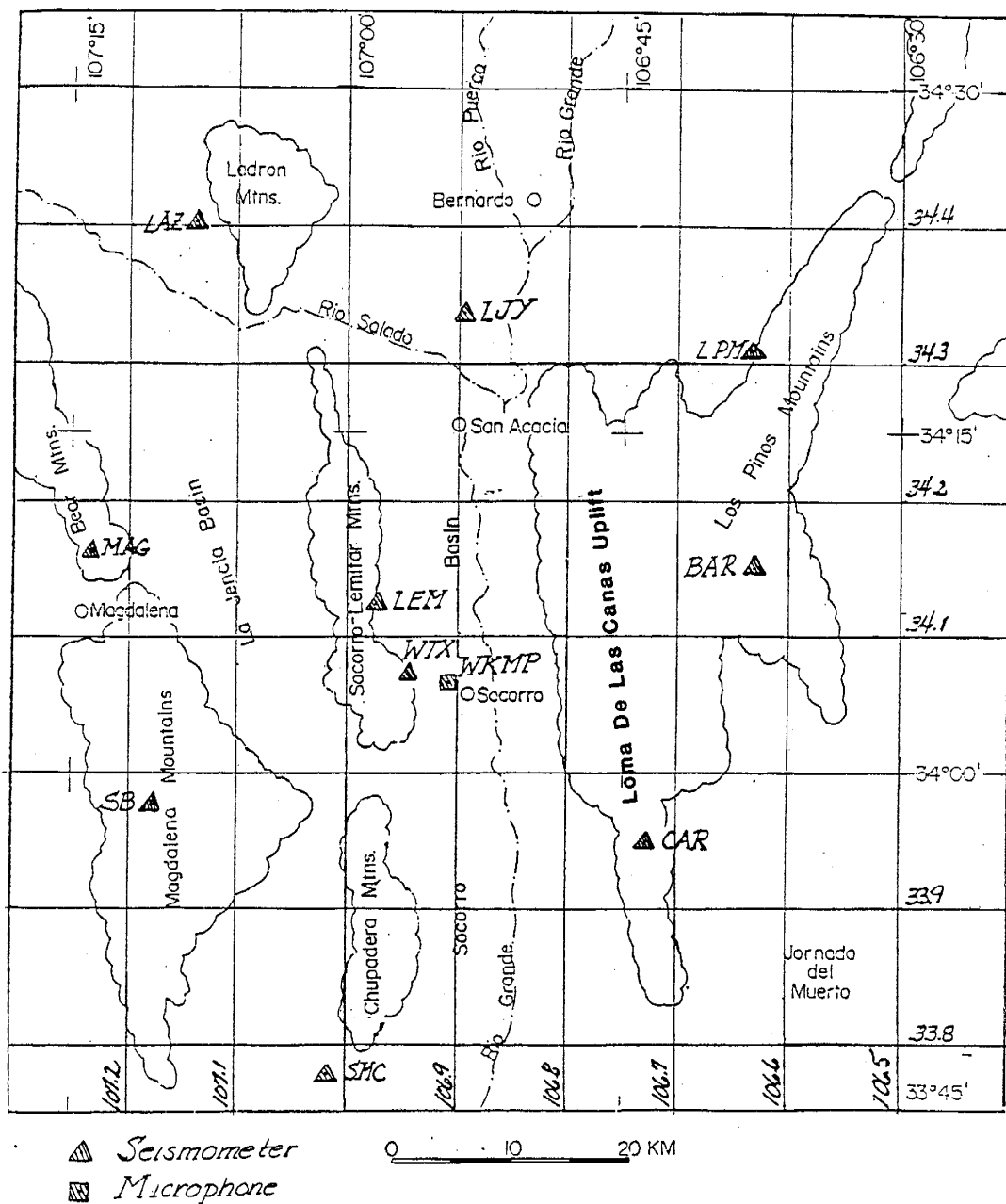


Figure 2. Major physiographic features near Socorro. Also shown are the locations of the NMT-USGS seismograph stations as of mid-1984.

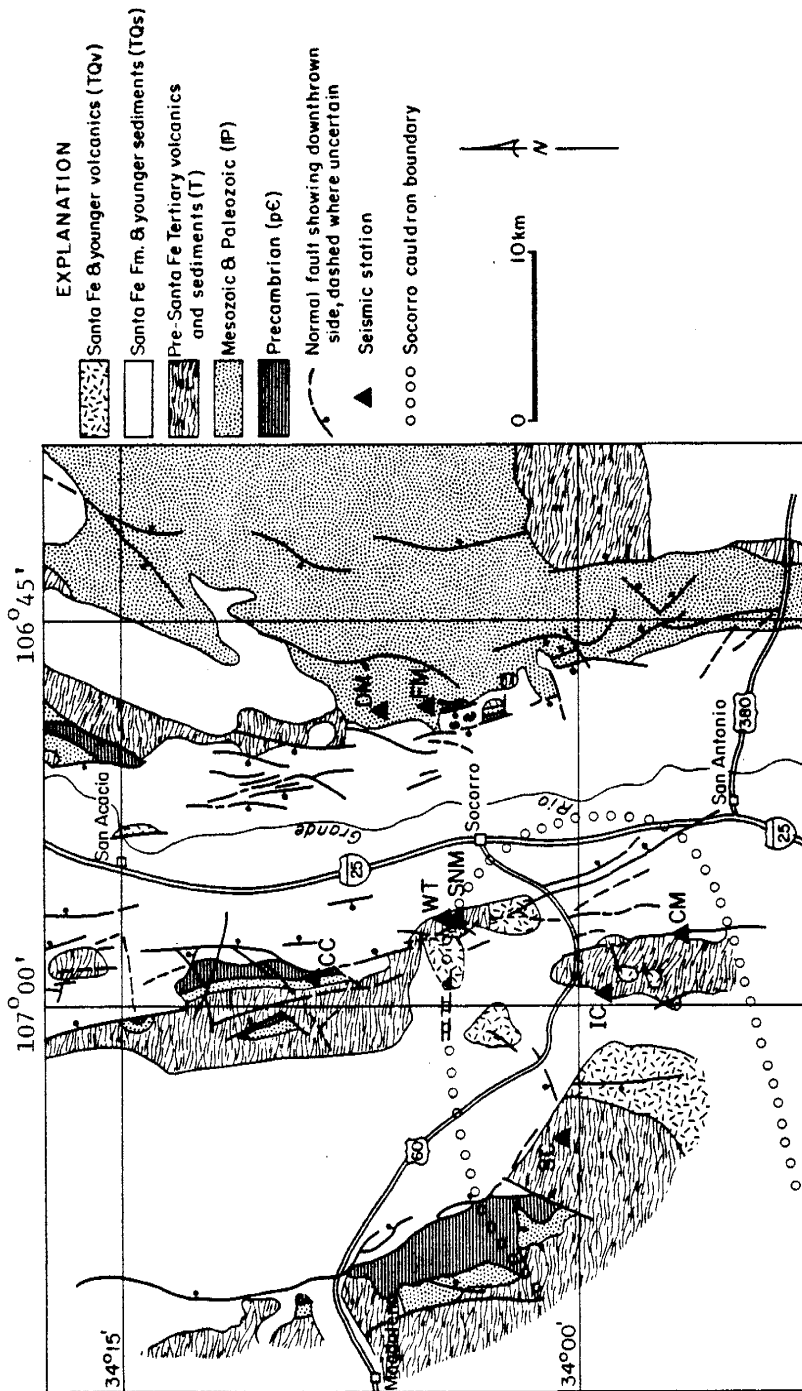


Figure 3. Simplified geological map of the study area. Stations used for digital recording are marked with triangles (after Sanford et al. (1972), from Carpenter (1984)).

Sedimentary rocks were deposited in the area during Paleozoic time and have been preserved east of the Rio Grande, in the Socorro-Lemitar Mountains, and in the Magdalena Mountains. Broad deformation occurred during the Laramide orogeny and has contributed to the subsequent structural complexity in the area. A significant area of positive relief existed in the area during Eocene time, consequently erosion occurred which removed much of the Mesozoic section [Cather, 1983]. As a result, in many places Tertiary rocks were deposited directly upon late Paleozoic or Precambrian rocks.

Activity in the Datil-Mogollon volcanic field during Oligocene and Miocene time covered the region with calc-alkaline ash flow sheets [Chapin and Seager, 1975]. The Socorro cauldron developed about 33 mya as the northeastern extension of the Datil-Mogollon field [Osburn and Chapin, 1983].

Rifting began about 32 mya in the Socorro area [Chapin, 1983]; this was contemporaneous with the later stages of cauldron development. Active rifting continued through mid-Miocene time. Between 7 and 12 mya, rhyolitic dome eruptions occurred along the northern moat of the Socorro cauldron [Chapin et al., 1978]. Subsequently the Socorro-Lemitar and Magdalena Mountain blocks were uplifted and rotated [Chamberlin, 1983]. Tectonic activity continues currently and has been accompanied by the intrusion and eruption of flood basalts during the last 4 my [Kelley and Kudo, 1978].

Important geological and geophysical characteristics of the RGR in the Socorro area include: high heat flow [Reiter and Smith, 1977], deep alluvial basins [Sanford, 1968], roughly north-south trending Quaternary boundary faults [Sanford et al., 1972; Machette and

McGimsey, 1982], the intersection of a transverse shear zone with the rift [Chapin et al., 1978], the presence of an extensive mid-crustal magma body [Sanford et al., 1972; Rinehart et al., 1979], an area of positive surface uplift north of Socorro [Rielinger et al., 1978; Larsen et al., 1986] and a high level of seismicity, frequently in swarms [Sanford et al., 1983].

The region around Socorro is historically and presently the most seismically active segment of the RGR [Sanford and Holmes, 1962; Sanford et al., 1983]. The seismicity of the Socorro region for the period 1982-1985 is shown in Figure 4. For this recording period the most seismically active regions were southwest of Socorro and north of Socorro (near San Acacia). Previous studies [Shuleski, 1976; Johnston 1978; Ward 1980; Sanford and Schlue, 1980; Carpenter and Sanford, 1986] have interpreted some of the activity southwest of Socorro to be related to the upward intrusion of small quantities of magma.

Based upon the analysis of reflected S-phases observed on microearthquake seismograms, Sanford and coworkers identified and mapped the upper surface of a broad magma body at a depth of about 19 km [Sanford et al., 1972; Rinehart et al., 1979]. This mid-crustal magma body was subsequently identified on the reflection seismic profiles acquired by the Consortium for Continental Reflection Profiling (COCORP) north of Socorro [Brown et al., 1979]. Figure 5 shows the position of the COCORP profiles relative to the outline of the mid-crustal magma body as mapped by Sanford and coworkers.

The digital data analyzed in this study were acquired about 14 km southwest of Socorro at station South Canyon (SC). The location of the station is shown in Figure 3. The site lies in the foothills of the

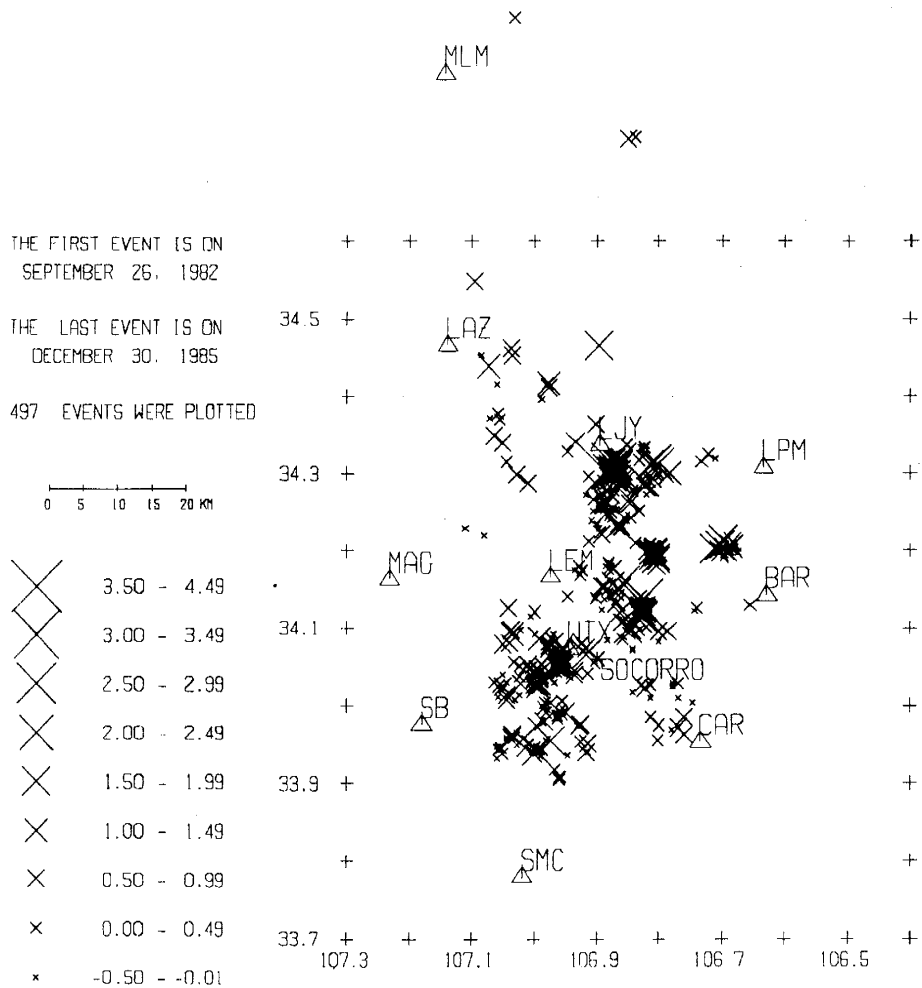


Figure 4. Seismicity map of the Socorro area for the time period September, 1982 through December, 1985. The seismograph stations operational during this period are indicated by triangles.

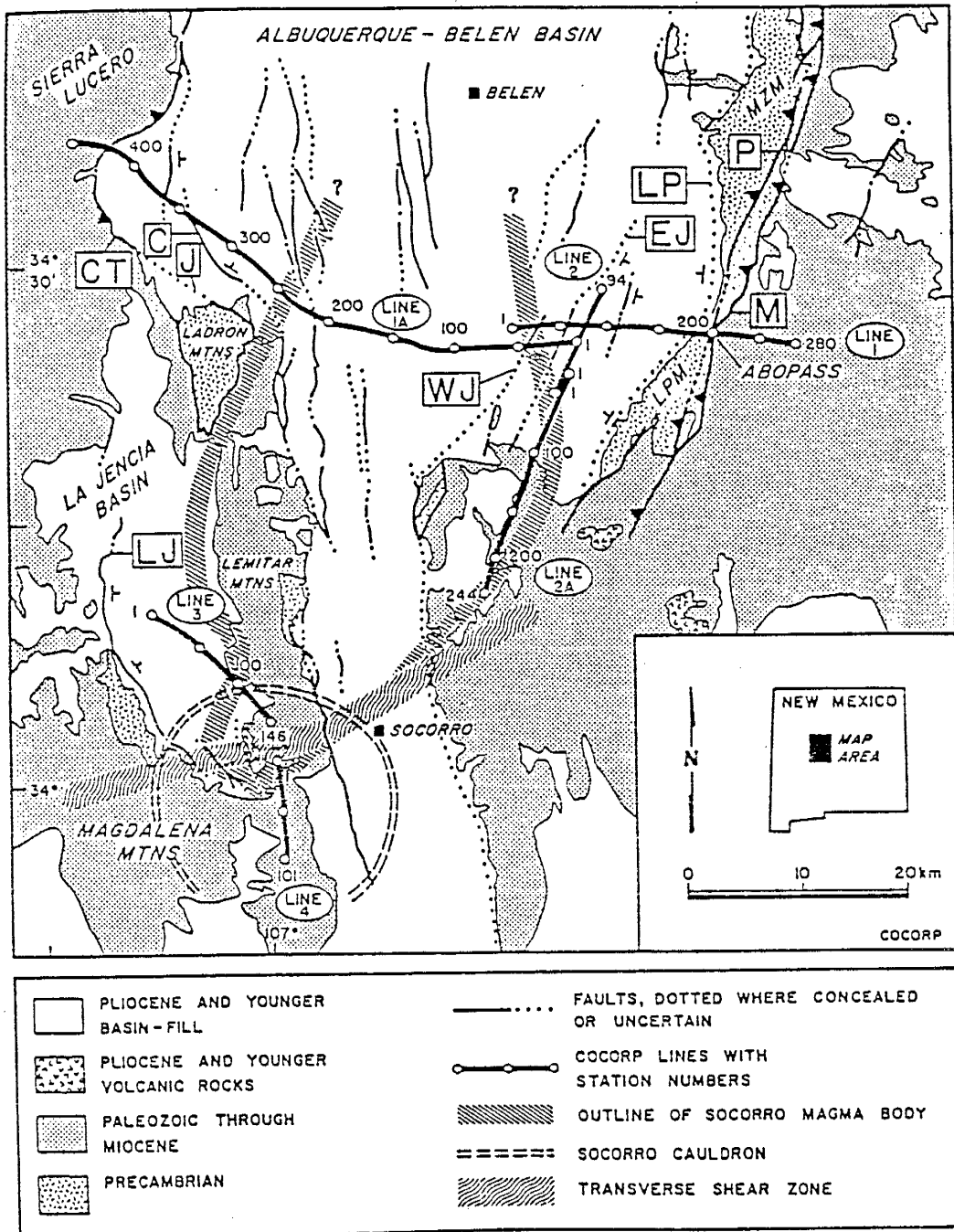


Figure 5. Location of COCORP New Mexico surveys; geology sketched from Kelley (1977) and Machette (1978); approximate extent of Socorro mid-crustal magma body drawn from Sanford et al. (1977) and Rinehart et al. (1979). Faults labeled CT, C, J, WJ, EJ, LP, P, M, and LJ are the Commanche, Coyote, Jeter, West Joyita, East Joyita, Los Pinos, Paloma, Montosa, and La Jencia, respectively (named after Kelley, 1977); LPM and MZM are the Los Pinos and Manzano mountains, respectively (from DeVogd (1986)).

Magdalena Mountains on a thick sequence of Tertiary volcanic rocks. Figure 6 is an simplified cross-section near the station (from Carpenter [1984]).

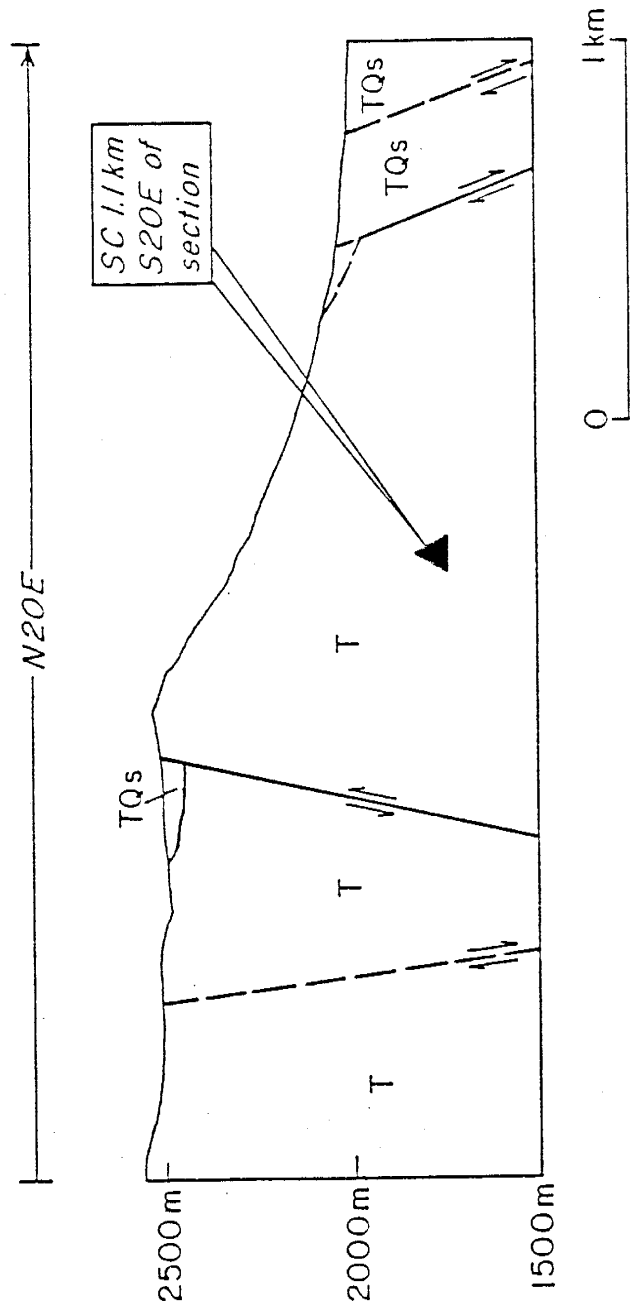


Figure 6. Simplified geologic cross section for station SC (after Osburn, 1978).

PREVIOUS INVESTIGATIONS

An important feature of rifts in general is the intrusion and extrusion of magma, large quantities along oceanic rifts and lesser amounts along continental rifts. Evidence in the geological record points to the existence of many different shapes and sizes of magma chambers. The shapes and sizes, as well as the methods of generation of magma bodies are important in understanding major tectonic processes as well as individual volcanic systems. Various geophysical techniques have been used in a few areas to define the presence of magma in the earth's crust.

The presence of magma will give rise to anomalous physical properties which are detectable by a number of geophysical techniques (Table 1). Seismic techniques fall into two general categories, active, which utilizes an artificial source, and passive, which utilizes earthquakes as the seismic source. This study is an example of a passive experiment.

Lateral variations in the density of crustal rocks due to zones of full melt may be detectable using gravity techniques. Normal crustal rocks have a Curie temperature considerably below the inferred temperature of most magmas. This fact makes magnetic methods potentially useful for detecting the presence of magma in the crust. The resistivity of rocks is proportional to the amount of melt present, hence electrical methods may also be used to detect magma. A summary of relevant physical properties at high temperatures is contained in Table 1. In general however, gravity, magnetic, electrical, and electromagnetic methods are useful for indicating the presence of partial or full melts but they do not have adequate resolving power to be

Table 1. Summary of Temperature Effects on Geophysical Properties of Crustal Rocks.

Rock Type	<u>Compressional Wave Velocity</u>	
	Temp ($^{\circ}\text{C}$)	Velocity (km/sec)
Andesite	20	6.1
	500	5.7
	1000	4.3
Basalt	20	6.6
	500	6.4
	1000	3.0
Granite	20	6.3
	500	5.7
	900	4.2

Rock Type	<u>Electrical Conductivity</u>	
	Temp ($^{\circ}\text{C}$)	Conductivity (mhos/m)
Andesite	400	5×10^{-11}
	750	3×10^{-9}
	1000	7×10^{-8}
Basalt	400	8×10^{-10}
	750	8×10^{-8}
	1000	2×10^{-6}
Granite	400	7×10^{-11}
	750	6×10^{-9}
	1000	2×10^{-7}

Rock Type	<u>Thermal Conductivity</u>	
	Temp ($^{\circ}\text{C}$)	Conductivity (cal/cm sec $^{\circ}\text{C}$)
Andesite	20	4.2
	500	3.1
	900	2.1
Basalt	20	5.3
	500	4.3
	1000	2.9

Rock Type	Temp ($^{\circ}\text{C}$)	<u>Density</u>
		Density (kg/m^3)
Andesite	20	2590
	1000	2450
	1200	2420
Basalt	20	2860
	1000	2800
	1200	2690

Data compiled from Murase and McBirney (1973), Christensen (1979) and Keller (1982).

used for the delineation of magma body structure. In the discussion that follows the focus will be on examples where seismic techniques have indicated the existence of magma and potential field methods are only discussed as supporting evidence.

Both active and passive seismic techniques have been utilized to locate magma bodies beneath the Kamchatka peninsula [Gorshkov, 1956; Utnasin et al., 1976]. Techniques used by the Russian workers include measuring P-wave delays, P-wave attenuation, weakening of S-waves and shadow effects caused by diffraction. Of particular interest is the study of Balesta and Farberov [1968] which used longitudinal and fan shooting profiles to identify regions of strong P-wave attenuation and delays beneath the newly formed Piip crater. These regions were modelled as very shallow pockets of magma with horizontal dimensions of 100 m by 40 m. This survey is one of the highest resolution studies reported to date.

Because of very recent volcanism and extraordinary hydrothermal activity, the presence of a shallow magma chamber beneath Yellowstone National Park was postulated at an early date. Recently a number of geophysical studies have been conducted which strongly support the existence of silicic magma beneath the Yellowstone caldera [Smith et al., 1977; Iyer and Stewart, 1977; Morgan et al., 1977]. Seismicity patterns mimic those observed in other areas of localized high heat flow; focal depths outside the caldera range up to 25 km while beneath the caldera none have been observed greater than 5 km [Smith et al., 1977]. Iyer et al. [1981] and Daniel and Boore [1984] used teleseismic and shear wave delays to map a large, deep low-velocity body beneath the Yellowstone caldera. The body was

estimated to extend from the upper crust to a depth of about 250km. Gravity and magnetic data both exhibit a pronounced low roughly coincident with the caldera boundaries [Smith et al., 1977]. These results are summarized in Figure 7. The amplitude and frequency characteristics of P and S-waves from local earthquakes indicates the magma is not uniformly distributed in the upper crust.

The Long Valley caldera, east of the Sierra Nevada Mountains in California, is an area where a number of geophysical investigations have indicated the presence of magma in the crust. The caldera is quite young, ($\approx 700,000$ years old) and has a history of very recent volcanic activity. A low-velocity body with a diameter of 10-15 km and a 10-15 percent velocity contrast was mapped beneath the caldera by Steeples and Iyer [1976] using teleseismic delays. Microearthquake studies by Ryall and Ryall [1981] also suggested the existence of magma accumulations in the crust. Results of potential field surveys also indicate the existence of upper crustal magma [Jachens and Roberts, 1985]. Unusually high earthquake activity in portions of the caldera as well as recent measured uplift as great as 25 cm [Savage and Clark, 1982] are consistent with the inflation of an upper crustal magma chamber. No high resolution seismic studies pertaining to the structure of the body have yet been published.

The Krafla region in Iceland has provided one of the best documented episodes of magmatic activity and magma accumulation to date. Interpretation of seismic, gravity, magnetotelluric, and crustal deformation data has led to the development of a model of inflation and eruption for the Krafla volcano [Einarsson, 1978; Bjornsson et al., 1979; Tryggvason, 1980; Thayer et al., 1981]. Magma appears to

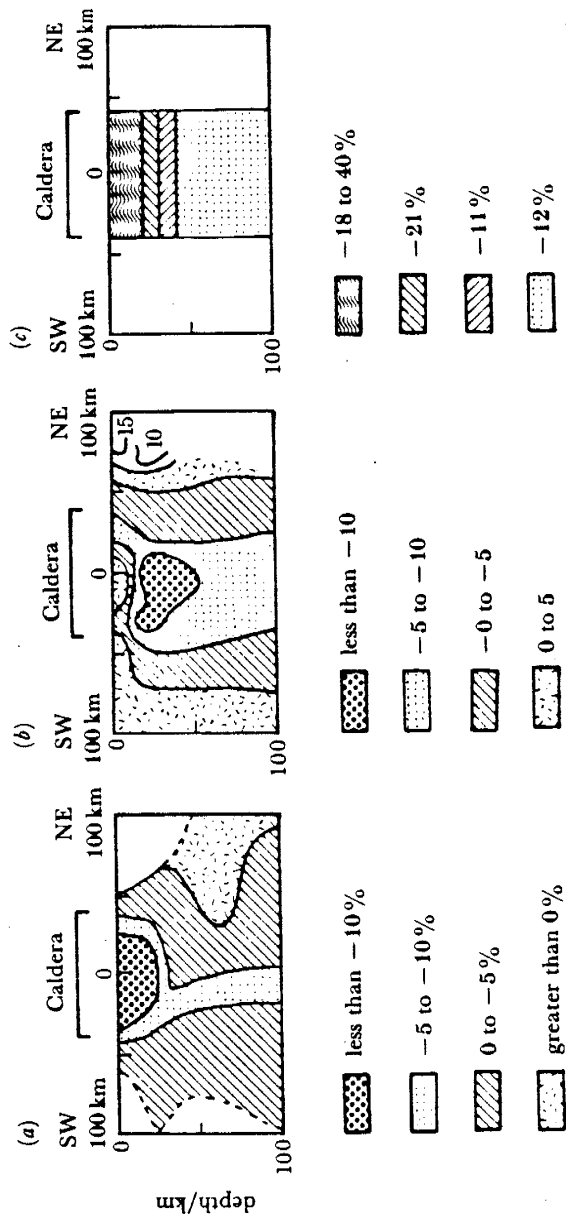


Figure 7. Southwest-northeast cross sections of seismic wave velocity and density anomaly models for Yellowstone region. (a) P-wave velocity, Iyer et al. (1981); contour interval 5%. (b) Density in units of 0.01 g cm Evoy (1978). (c) S-wave velocity, Daniel & Boore (1982) (from Iyer (1983)).

accumulate beneath the volcano at a depth of about 3 km leading to inflation of the volcano. At some critical point, deflation begins to occur and small quantities of magma are erupted into the caldera. Most of the magma appears to be injected laterally away from the volcano.

Einarsson [1978] studied S-wave screening beneath the Krafla volcano. Using ray tracing techniques, two zones of high S-phase attenuation were identified. The attenuative bodies have horizontal dimensions of 1 to 2 km. The depth to the top of the bodies is 3 km or less. While the bottom is not well constrained, it appears to be shallower than 7 km. The shallow focal depths of earthquakes beneath the caldera support the inferred location of the magma body. However, the seismic data give little information on the internal structure of the magma chamber [Sanford and Einarsson, 1982].

Mizoue et al. [1982] mapped an unusual crustal discontinuity beneath central Japan using near vertical S to P (S_xP) and S to S (S_xS) wave reflections from microearthquakes. The discontinuity was estimated to have a depth of about 14 km with some dip and surface irregularities (Figure 8). The theoretical amplitude ratio of the S_xP and S_xS reflections agrees well with the observational results when the S-wave velocity decreases from 3.35 km/sec to 0.00 km/sec across the discontinuity. Mizoue et al. suggest that due to the proximity of the volcanic front in the area, an extensive magma body is a plausible explanation for the reflections. The observed distribution of focal depths in the area agrees with that observed in other areas where the presence of magma in the crust has been postulated (Figure 9). The study of Mizoue et al. was able to map the upper surface of the body

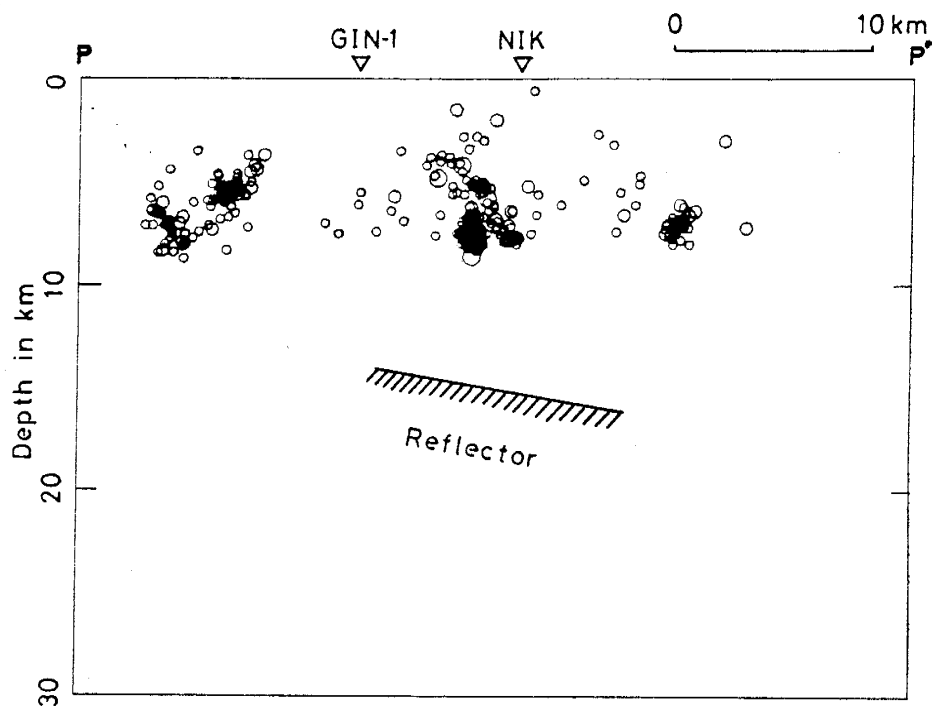


Figure 8. Relationship of microearthquake distribution and mid-crustal discontinuity, Tochigi Prefecture, Japan. Earthquake distribution is projected onto the plane P-P' in Figure 9 (from Mizoue et al. (1982)).

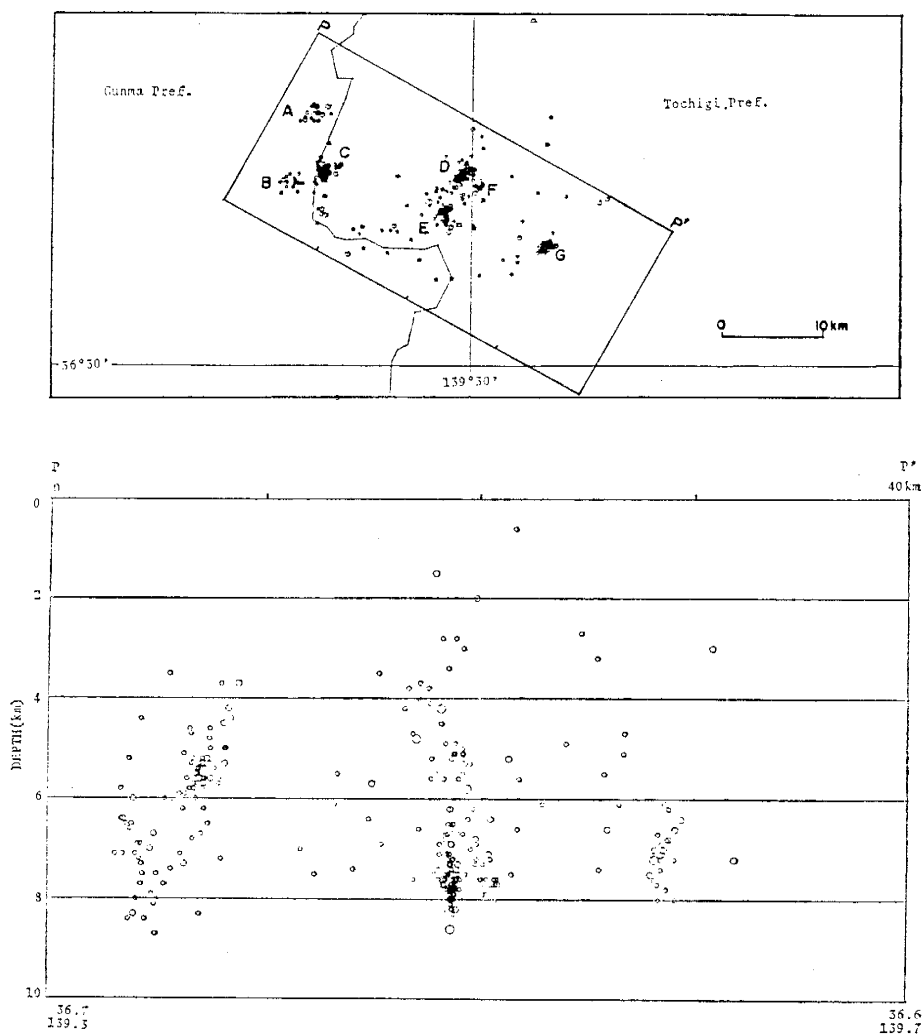


Figure 9. Relationship of epicenters and hypocenters from Tochigi Prefecture, Japan. Epicentral groups A through G (top) are projected onto P-P' and displayed as a depth section below (from Mizoue et al. (1982)).

but was unable to speculate about the internal structure of the body.

Seismic reflection profiling using a VIBROSEIS (trademark Continental Oil Co.) source is a very powerful technique for imaging the crust. This technique has been used to aid in the detection of magma bodies in two areas, Death Valley, California and the Rio Grande rift, New Mexico. De Voogd et al. [1987] discovered exceptionally strong reflections at a depth of 15 km on COCORP profiles from Death Valley. Due to the similarity with reflections observed on earlier COCORP profiles in the Rio Grande rift (to be discussed below), these reflections were interpreted as coming from a mid-crustal magma body (Figure 10a). A possible magma conduit connecting the magma body at 15 km to a surface cinder cone about 690,000 yrs old was identified (Figure 10b). No estimate of the structure of the magma body was made from initial interpretations of the Death Valley data.

Rio Grande Rift

The initial discovery of a magma body in the Rio Grande rift near Socorro, New Mexico was made using reflected phases from local earthquakes. Early studies [Sanford and Long, 1965] indicated a significant fraction of the local earthquakes had at least two sharp reflected phases in addition to the usual P and S-phases. These were interpreted to be S to P (S_zP) and S to S (S_zS) phases from a mid-crustal discontinuity (Figure 11). Subsequently, more detailed studies determined the depth of the reflecting horizon to be the top of a thin magma body at a depth of 19.2 ± 0.6 km [Sanford et al., 1972; Sanford et al., 1977; Rinehart et al., 1979].

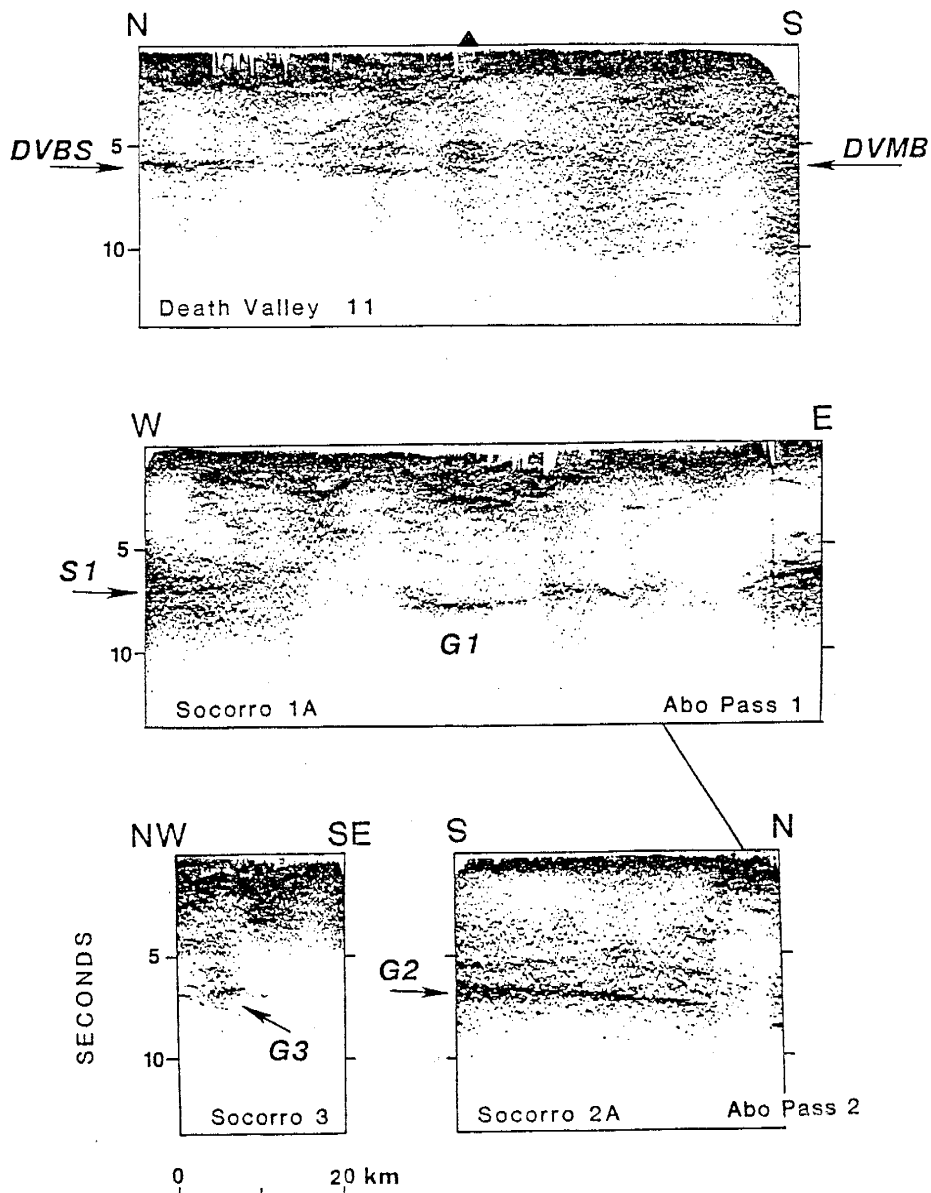


Figure 10a. True amplitude sections from Death Valley line 11 (top) and the Rio Grande rift COCORP data. Unusually prominent reflections DVBS, G1, G2 and G3 are attributed to partially molten rocks (from DeVogd (1986)).

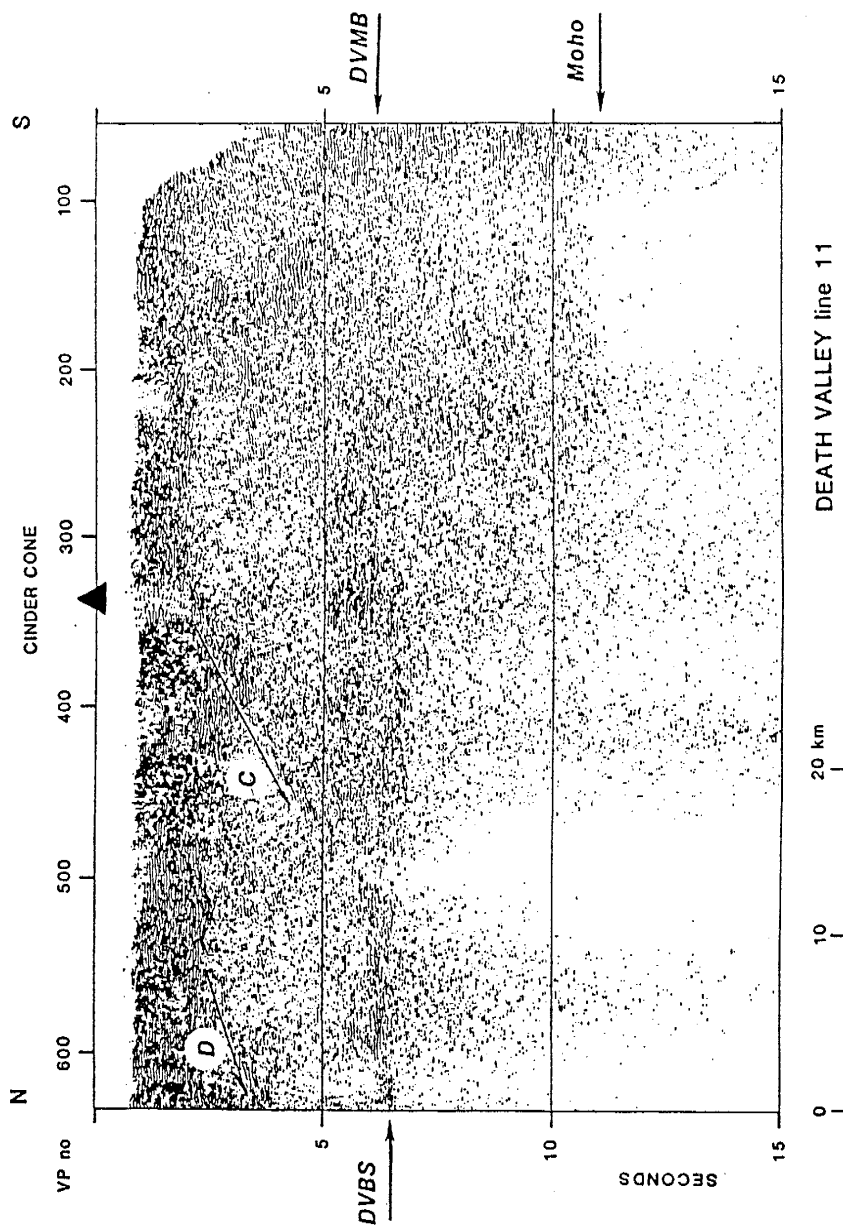


Figure 10b. True amplitude section of COCORP Death Valley line 11. The unusually strong reflector labelled DVBS is interpreted as a magma body. C is inferred to be a magma conduit connecting the magma chamber to the cinder cone at the surface (from DeVoogd (1986)).

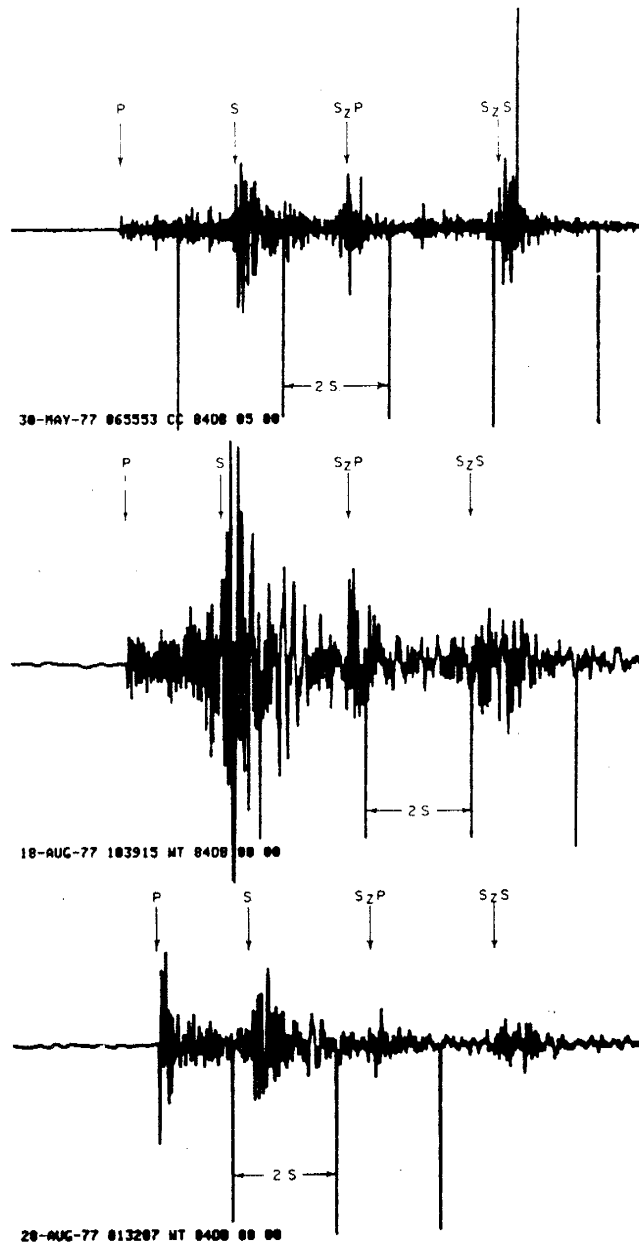


Figure 11. Microearthquake seismograms with well-defined S to P and S to S reflections from the top of the mid-crustal magma layer (from Sanford et al. (1977)).

A remarkable feature of the reflected phases is their amplitude relative to the direct phases. Sanford and coworkers demonstrated with theoretical calculations that ordinary mid-crustal discontinuities could not account for the amplitude of the reflections even if radiation from the source was 10 times as great in the downgoing, reflected pulse as in the direct phase. However, if the discontinuity is assumed to be singular (i.e. a rigid above a non-rigid material) then the absolute amplitude of the reflected phases as well as the amplitude ratio of the reflected phases can be explained. The general structure of the body was assumed to be a single, sill-like body with a thickness less than 1 km [Sanford and Einarsson, 1982].

Deep reflection profiles acquired by COCORP in 1975 and 1976 showed distinctive reflections coincident in depth and lateral extent with the positions of the reflections from the microearthquake studies. Based upon the strength and position of the reflections they were interpreted as being due to the presence of magma at these depths [Brown et al., 1979]. In a detailed study of the COCORP data at a single location, Brocher [1981] inferred characteristics of the physical nature of the magma body by computing reflectivities. Reflectivity is the ratio of the spectra of the incident and reflected waves. Brocher compared observed reflectivities to those calculated from models of thin (35-38 m), plane parallel-layered intrusions. The best-fitting models were those with several layered intrusions (Figure 12).

To date the majority of studies describing magma bodies have utilized methods that do not have adequate resolving power to decipher anything about the structure of the bodies. only infer their existence. For example, the method of teleseismic delays uses

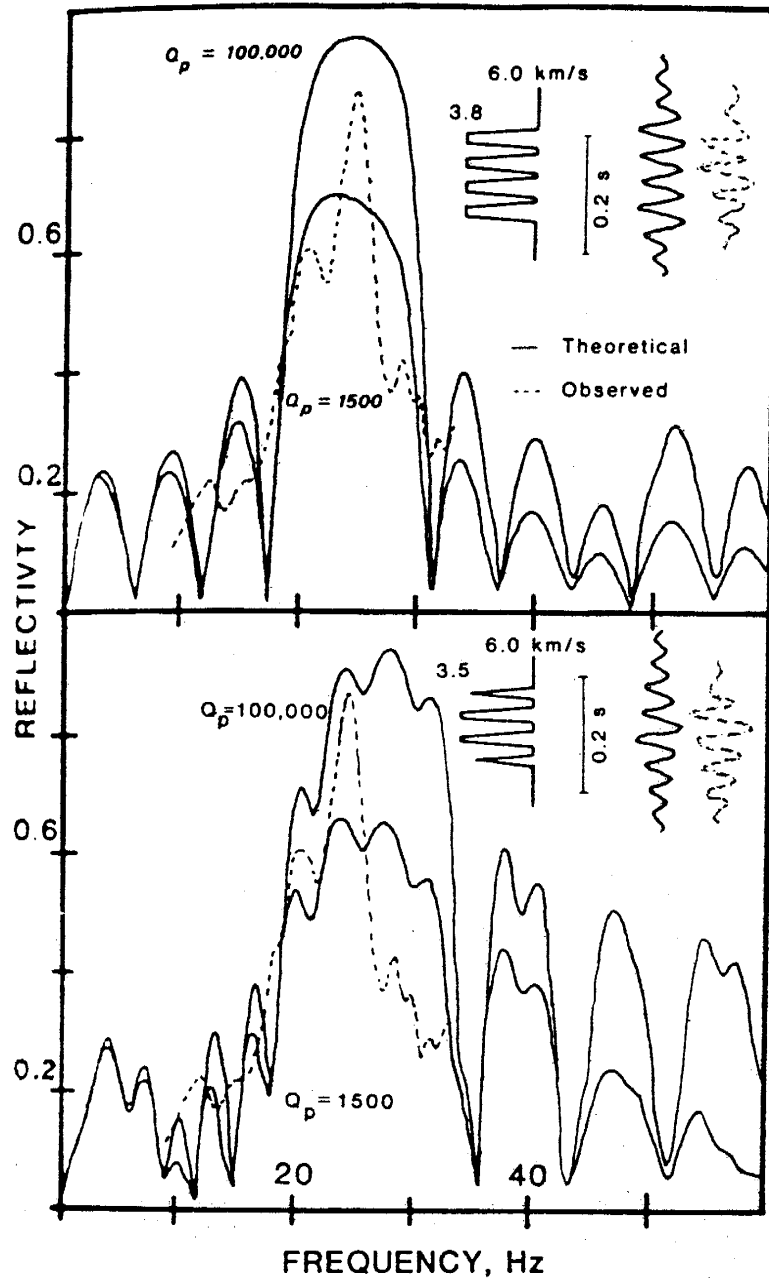


Figure 12. Theoretical reflectivity as a function of frequency calculated assuming vertical incidence and no density contrast for the impedance models shown above right in units of km/s and plotted as a function of a two-way travel time. Observed reflectivity at VP15 of Socorro line 1 by 0.010 s multiplied by layer velocity. Time domain response of model shown at top far right as a solid line, observed seismic waveform shown as a dotted line. The amplitude scales of the observed and calculated seismograms differ (from Brocher (1981)).

wavelengths many km long, hence resolving any details of small or thin bodies becomes impossible. The Kamchatka study of Balesta and Farberov and the analysis of COCORP data by Brocher are the only studies other than the present one to try to infer anything about the actual size and configuration of magma bodies. The Kamchatka bodies were quite small and shallow, structure was not resolved per se, only limits placed upon the size by the position of anomalous ray-paths.

While the COCORP survey was an active seismic experiment, the present study is the first to utilize microearthquakes to actually image magma bodies in the crust. Studying the upper and middle crust with signals from microearthquakes provides a method that complements other techniques for studying the geometry of magma bodies. Because the signals from microearthquakes originate at depth, they may intrinsically be better sources of energy for high resolution studies of a magma body at mid-crustal levels than signals generated at the surface in conventional surveys. This is the result of two factors, the very broad band nature of the microearthquake source (as opposed to the band-limited vibrator sweeps) and the geometry of the surface based surveys which requires the waveform to pass through the highly attenuative surface material twice (as opposed to only once for the microearthquake source).

DATA

All earlier microearthquake reflection studies of the Socorro mid-crustal magma body were based upon amplitude and arrival time data from analog recordings made at recording speeds of 1 to 4 mm/sec. In this study, digital recordings were used in the analyses. The physical properties of the magma body were deduced from analysis of the phase, duration and general waveform characteristics of the S and P-wave reflections. This is the first study to present microearthquake P-wave reflection data that can easily be compared to the COCORP observations.

The data utilized in this study are from a moderate (for this region) microearthquake swarm located ≈ 16 km southwest of Socorro. The swarm occurred over a six day interval in 1977 (≈ 0600 May 30 to ≈ 0600 June 4) with the strongest event reaching a local magnitude of 2.2. About 75 shocks were recorded during the swarm sequence. Figure 13 shows the seismicity of the Socorro area from mid-1975 to late-1978 and the location of the swarm. Many of the details in this section are from a very detailed analysis of the swarm contained in Sanford et al. [1984].

The digital data with which the bulk of the analysis was done was acquired with a Sprengnether DR-100 digital event recorder deployed at station SC. The DR-100 has a 12-bit analog to digital converter and was operated at 100 samples per second (sps) throughout the swarm. The 100 sps sampling rate gives a 50 hz Nyquist frequency. The DR-100 unit continuously monitors the seismic background and compares a short-term average amplitude (STA) to a long-term average amplitude (LTA). If the STA/LTA ratio exceeds a

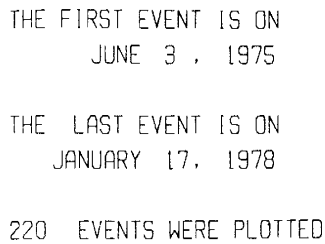


Figure 13. Seismicity of the Socorro area for the 1975-1978 time period. Seismic stations indicated are those in operation during 1975-1978. Location of COCORP profiles, outline of mid-crustal magma body and swarm location are indicated.

prescribed value, then the unit triggers and records the event. A low-cut 5 hz filter was used at station SC during the swarm to reduce triggering by teleseisms. All data were acquired with a vertical-component, 1 second natural period Mark Products L4C seismometer. Details on the calibration of the instrumentation are provided in Appendix I and in Carpenter [1984]. A general discussion on the playback and processing of the digital data is contained in Carpenter [1984]. The data utilized in this study have only the initial processing applied with one exception. To suppress side-lobe effects when comparing spectra of individual phases, a 10% cosine-tapered window was applied prior to transforming the data to the frequency domain.

Although ≈ 75 events were recorded while the DR-100 was deployed at station SC, only 25 of the events recorded were read onto computer tapes, 19 from May 30 and 6 from June 2, before the field tapes were erased. This unfortunate circumstance has limited the amount of data available for this study. Two additional factors contributed to the further exclusion of some data from the final data set. First, for reasons that will be explained in a later section, criteria based upon event strength were set that resulted in only events below some threshold magnitude being included in the final data set for analysis. Secondly, for reasons still not fully understood, the larger events in the swarm did not contain a well-recorded P_zP phase. In general the quality of the data recorded at station SC was exceptional. The signal to noise ratio of even the smallest events was quite good. In addition to the P and S phases, the reflected phases P to P and S to S were also well recorded (Figure 14).

In addition to the digital data recorded at station SC, analog

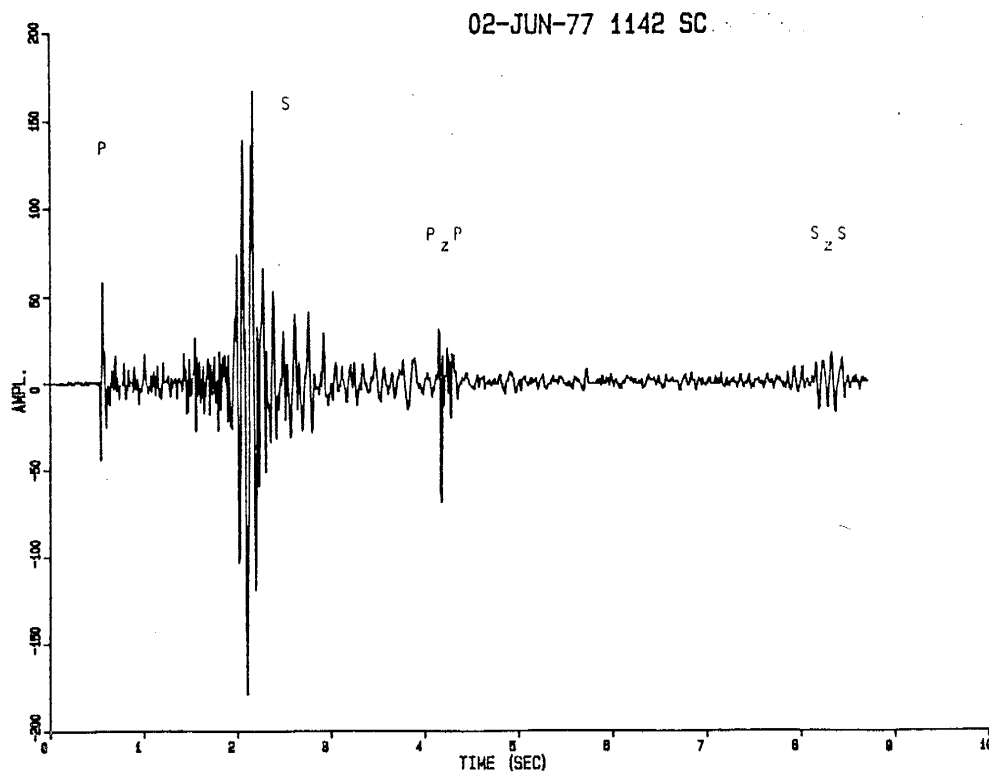


Figure 14. Example waveform from the May-June 1977 South Canyon microearthquake swarm. Event illustrates the well-defined, impulsive direct and reflected P and S phases. Abscissa is in digital units, local magnitude of the event is less than zero.

seismographs were operational at seven locations at various times during the microearthquake swarm. Data from these stations were utilized in the location of the events and in the determination of fault plane solutions. Figure 15 shows the locations of the seismograph stations relative to the epicentral region of the swarm.

The earthquakes were located with the algorithm HYPO71 (Revised) [Lee and Lahr, 1975]. The locations were obtained using a half-space P-wave velocity of 5.85 km/sec, a Poisson's ratio of 0.25 and station corrections. The magnitude and sign of these corrections essentially accounts for the thickness and type of material beneath each station. The initial station corrections were based upon the work of Ward [1980]. These corrections were then modified for this particular swarm location [Sanford et al., 1984]. This "fine-tuning" of station corrections has been done in the analysis of several other microearthquake swarms in the Socorro area [Jarpe, 1984; Ake, 1984]. Final average errors in epicenter (ERH) and depth (ERZ) were 0.31 km and 0.38 km respectively [Sanford et al., 1984]. Sanford et al. [1984] determined the average focal depth to be 9.7 km and the epicentral distance from station SC to be about 2.5 km for the swarm events (Figure 16). The vertical migration of hypocenters was minimal (≈ 125 m) during the course of the swarm. Changes in arrival time at station CC but not at station SC indicates some lateral migration of hypocenters however [Sanford et al., 1984].

Directions of P-wave first motions and amplitude ratios (direct P to direct S and direct S to reflected S) for the strongest event were used to determine the nature of faulting during the swarm by Sanford et al. [1984]. Figure 17 shows the focal mechanism determined with

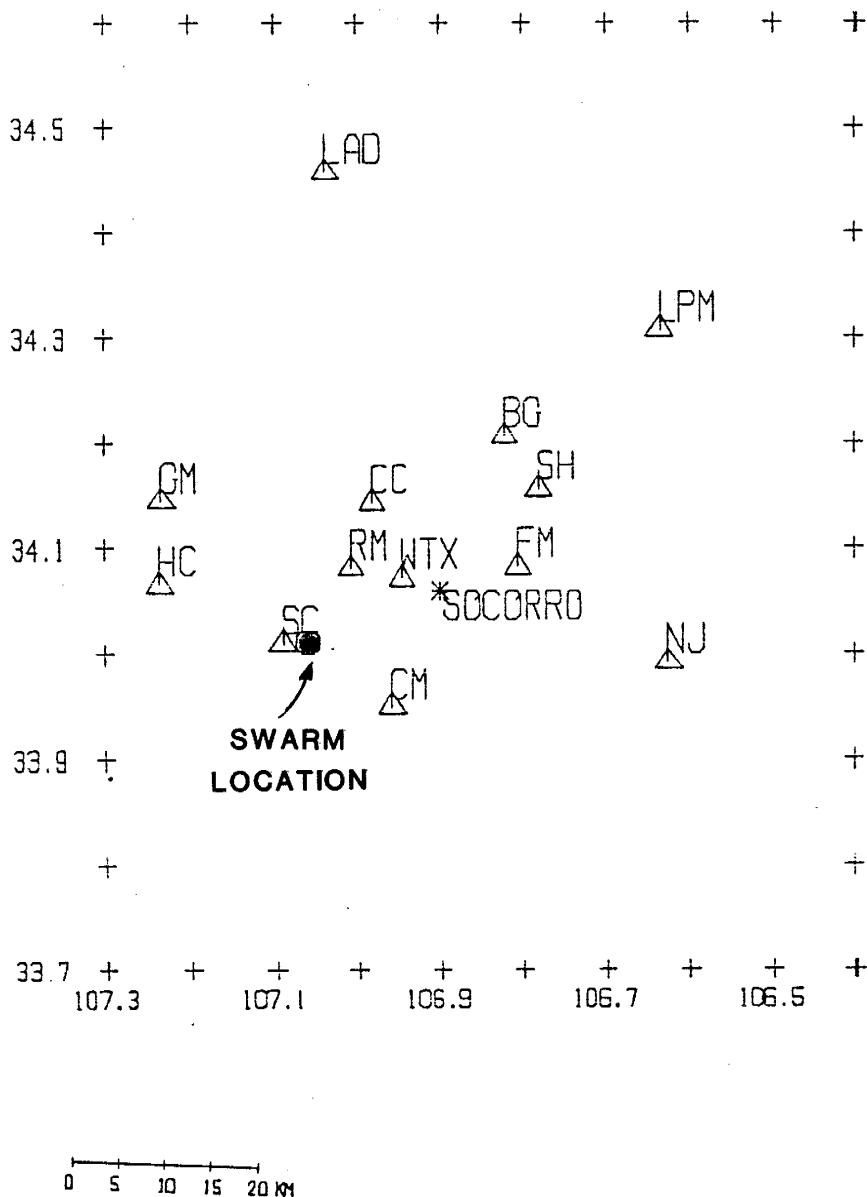


Figure 15. Relationship of the epicentral region of the May-June microearthquake swarm to the stations operational at that time.

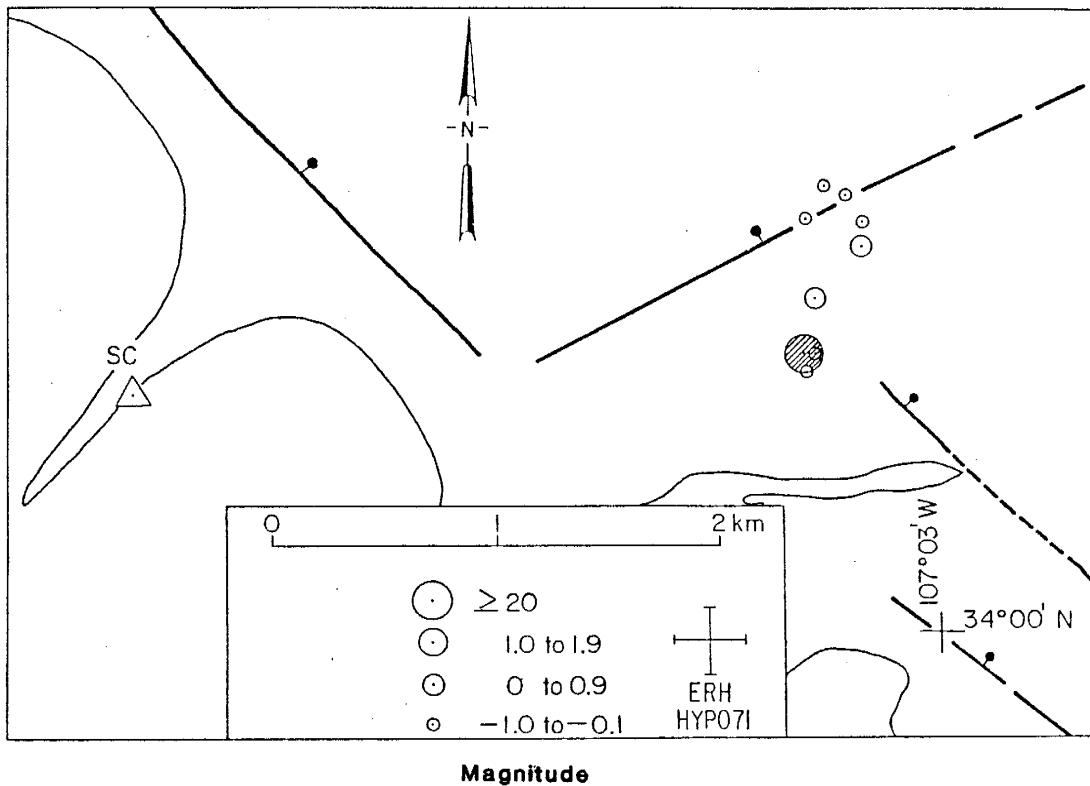


Figure 16. Relationship of epicenters to mapped faults for located May-June 1977 events. Station SC and topographic contours are indicated (from Sanford et al. (1984)).

the first motions and the amplitude ratios. The mechanism is a normal fault striking N13°W and dipping 50° to the west. Sanford et al. concluded that rupture must have been dominantly unilaterally downward due to the observed S to S_zS amplitude ratio.

Figure 18 shows the position of the earthquake swarm relative to the digital recording station (SC) and the top of the mid-crustal reflecting surface. Also shown are the radiation lobes from the focal region for a composite fault mechanism based upon P-wave first motions and the amplitude ratios as determined by Sanford et al. [1984]. The focal mechanism appears to be well constrained. A limited amount of rotation of the fault planes is permissible, however it is not enough to affect the relative polarity of the direct and reflected phases as recorded at station SC.

Two features of Figure 18 are important in understanding some of the characteristics of the reflected waveforms. First is the phase of the P and S signals as they leave the focus. For the direct and reflected P, the phase is the same, a dilatation. The direct and reflected S leave the focal region with the opposite sense of shear however. The second feature to note is that the raypaths at the recording station for direct and reflected phases have angles of incidence that differ by only 7° (assuming a half space velocity model).

Presented in Figures 19 through 23 are panels of microearthquake seismogram data obtained at station SC. Figure 19 contains complete seismograms with direct and reflected P and S phases. The figure demonstrates the overall similarity of the waveforms. The magnitudes of the events considered in the study were all less than zero. Figures 20 through 23 contain windows of the various phases from

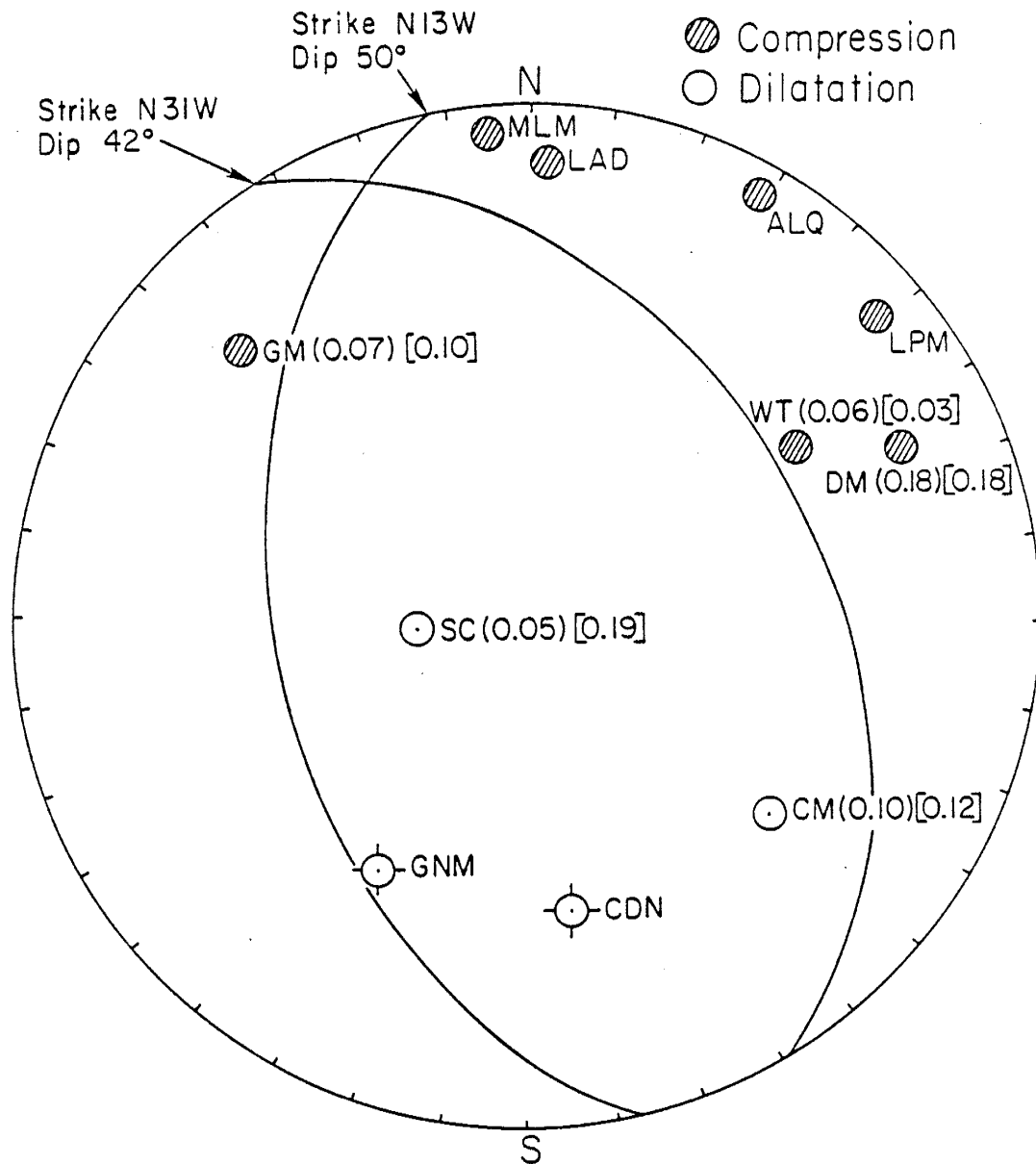


Figure 17. Fault mechanism for swarm earthquakes. First motions are from the strongest event in the swarm (June 2 at 06:48). The numbers in parenthesis are observed direct P to direct S, corrected to the focal sphere. Numbers in square brackets are theoretical ratios, direct P to direct S, at the focal sphere (from Sanford et al. (1984)).

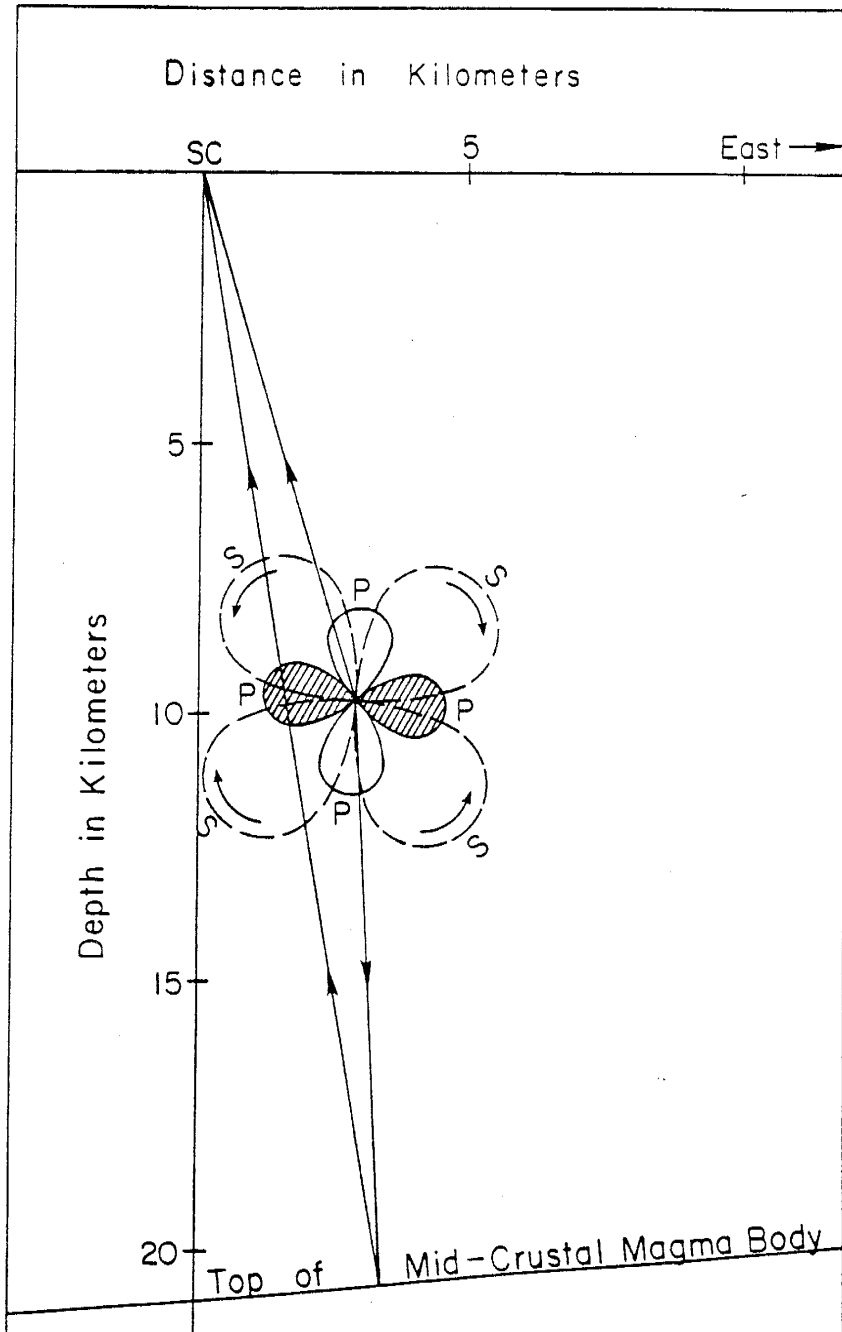


Figure 18. Positions of microearthquake hypocenters relative to the recording station (SC) and the reflection interface. The radiation lobes are illustrative of position but not magnitude. The raypaths for the direct and reflected phase are indicated. The focal mechanism appears to be well constrained. A very limited amount of rotation of the nodal planes is permissible, however, it is not enough to change the relative polarity of the direct and reflected phases.

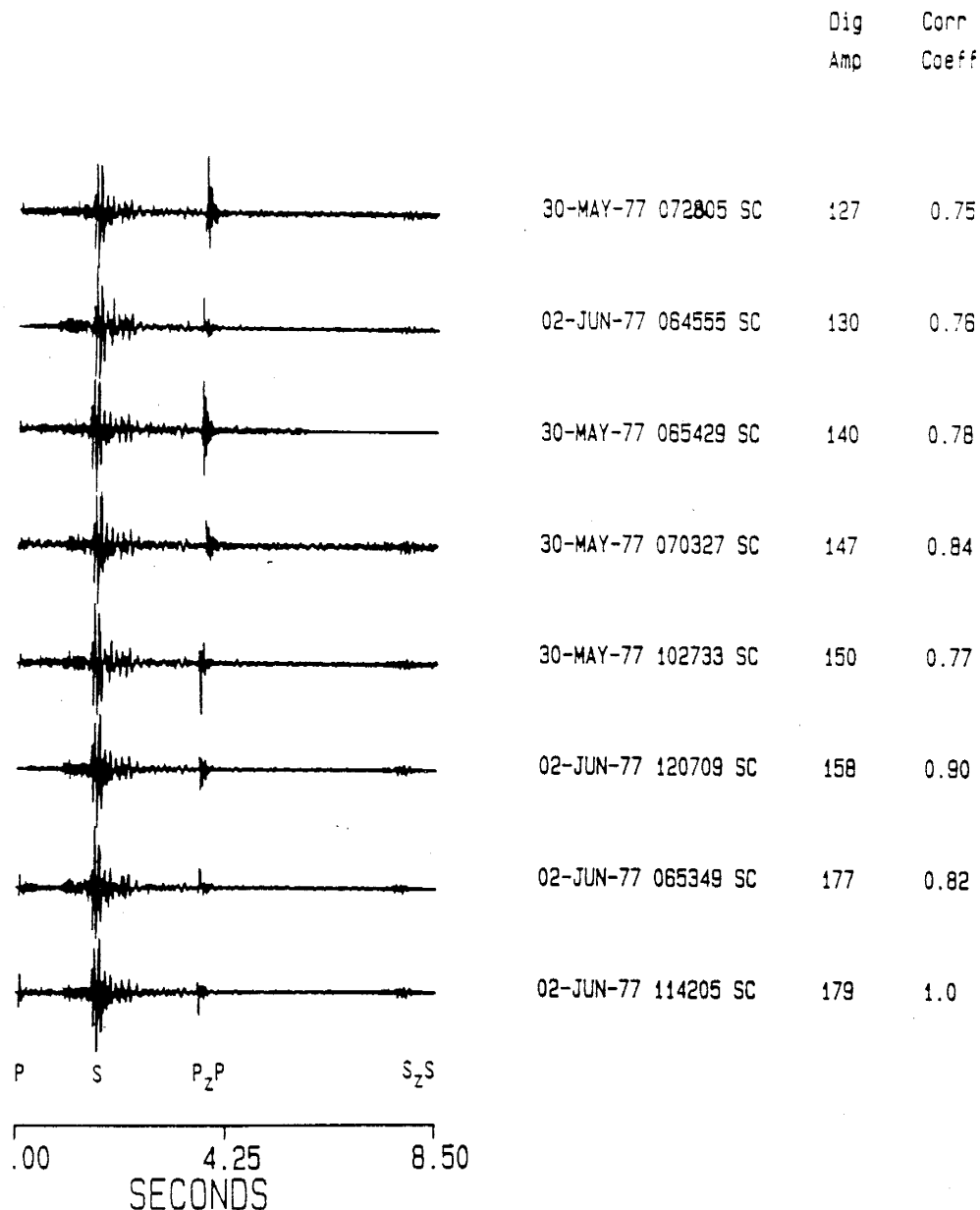


Figure 19. Seismograms showing direct P and S phases and reflected P₂P and S₂S phases. Local magnitudes of the events are all less² than zero. Amplitudes have been normalized for plotting. Maximum digital amplitudes and cross correlation coefficients relative to the last earthquake are listed for each event. For some events either the beginning or ending of the event was not recorded and appears as a straight line. The sampling interval is 0.01 second. Recordings are of ground velocity.

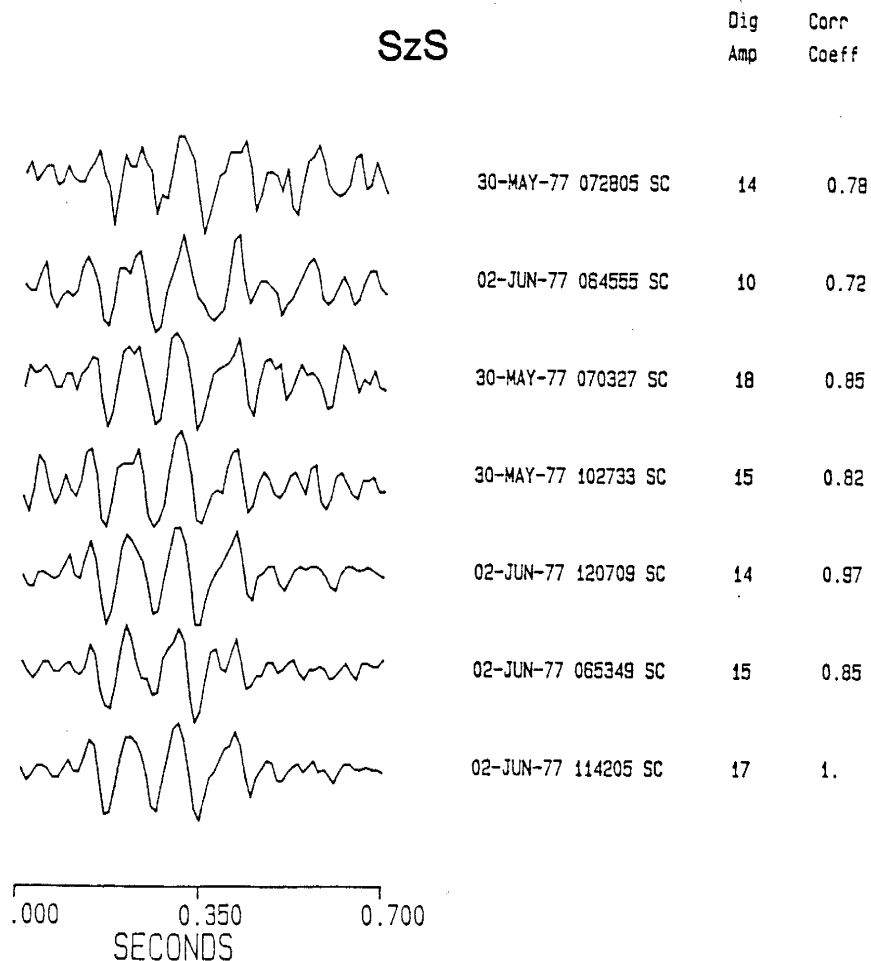


Figure 20. Windows of the reflected S-phase (S_zS) from the seismograms in Figure 19. Maximum digital amplitudes and cross-correlation coefficients relative to the bottom waveform are listed. The average correlation coefficient is 0.83.

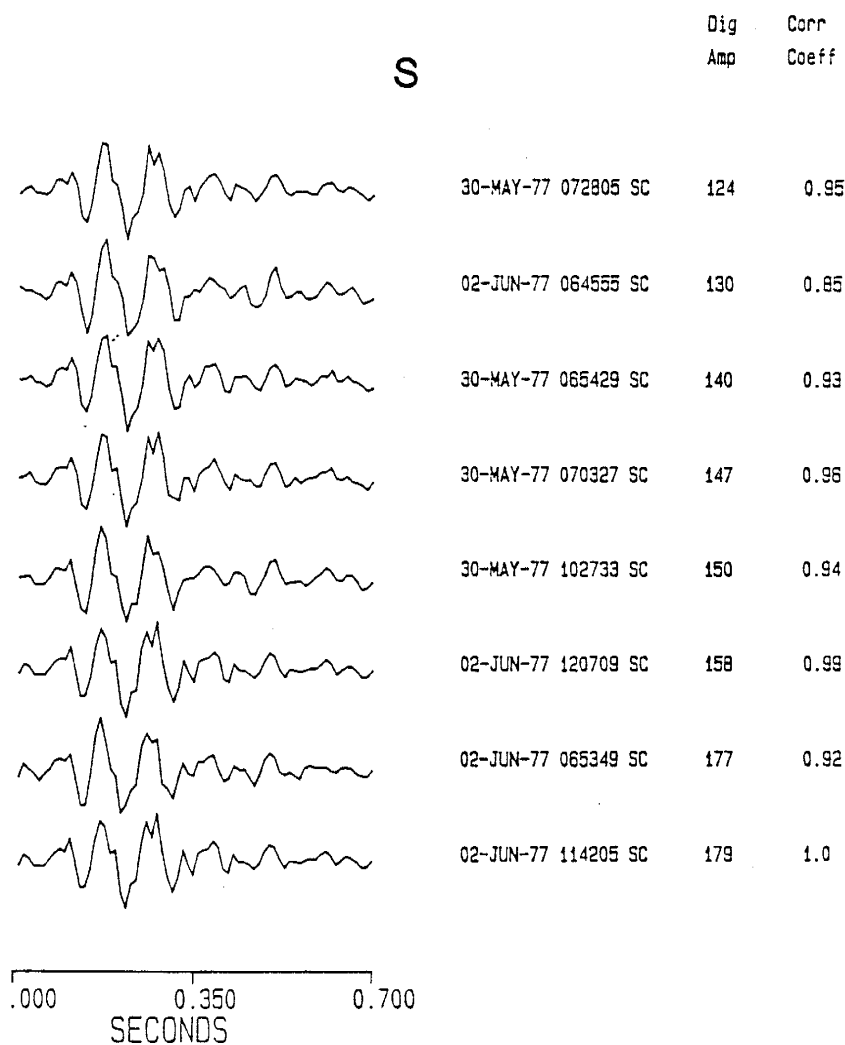


Figure 21. Windows of the direct S-phase (S) from the seismograms in Figure 19. Maximum digital amplitudes and cross correlation coefficients relative to the bottom waveform are listed. The average correlation coefficient is 0.93. Correlation coefficients between S and S_z are all positive and average 0.79.

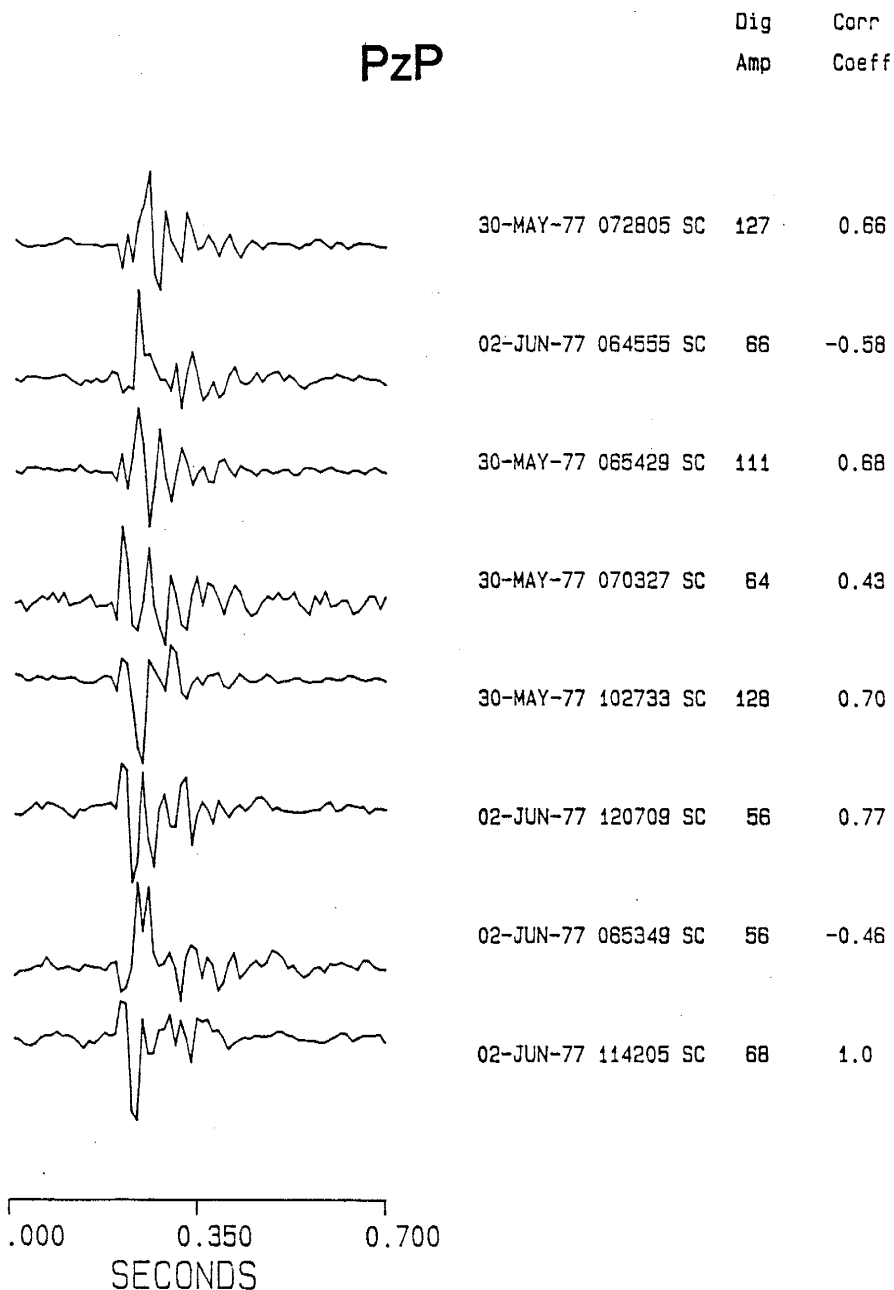


Figure 22. Windows of the reflected P-phase (P_zP) from the seismograms in Figure 19. Maximum digital amplitudes and cross-correlation coefficients relative to the bottom waveform are listed. Although much larger in amplitude than S_zS , cross-correlation coefficients are lower (avg = $^z0.60$) for these waveforms.

P

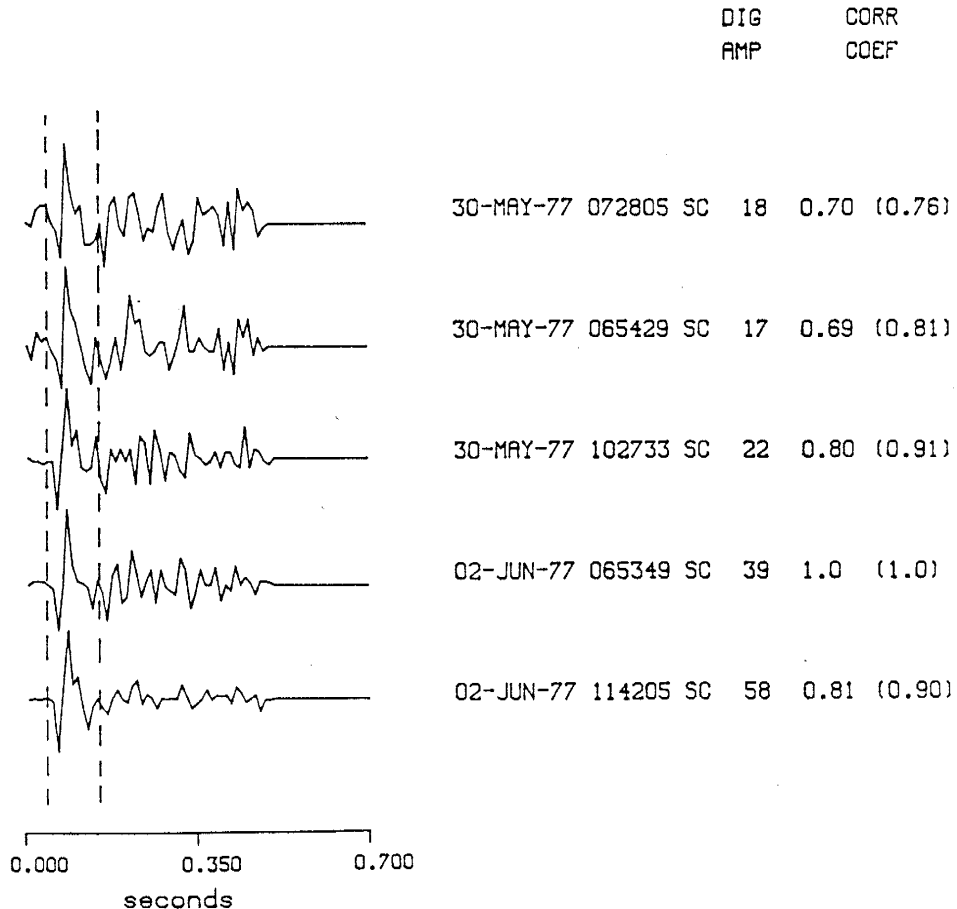


Figure 23. Windows of the direct P-phase (P) from the seismograms in Figure 19. Maximum digital amplitudes and cross-correlation coefficients relative to the bottom waveform are listed. The average correlation coefficient is 0.75. The correlation coefficients in parentheses are for the shorter windows (dashed on figure). Correlation coefficients between P and P_P average 0.45 and with one exception all are negative.

Figure 19 which will be discussed in detail in the following section.

DATA ANALYSIS

S_zS Reflections

The S_zS phase is a weak but clear phase (averaging about 10 percent of the S phase amplitude) that occurs 7.66 seconds after the direct P (Figure 19). Figure 20 shows 0.7 second windows of the Figure 19 microearthquake seismograms containing the S_zS phases. Correlation coefficients of all traces relative to the last trace in the panel vary from 0.72 to 0.97 (with an average of 0.83). This is exceptionally high considering the digital amplitudes of the reflected phases. These amplitudes range from 10 to 18, which is not large relative to background noise and/or to the strength of the coda at the arrival time of the S_zS phase (Figure 19). The S/N ratio of the S_zS phase ranges from 3:1 to 6:1 for the events included here.

Figure 24a is an illustrative cross-correlation function for two S_zS phase waveforms. The exceptional similarity of the S_zS waveforms is evident from comparison of Figure 24a with Figure 24b, which is the auto-correlation function of one of the S_zS waveforms. The sharp peak and symmetric envelope of the cross-correlation is almost identical to what is observed in the auto-correlation function. The evenly spaced, multicyclic nature of the correlation functions is indicative of multicyclic, nearly monochromatic data.

The spectral characteristics of the S_zS phase are illustrated in Figure 25 for two events. A dominant peak near 11 hz is present in all of the S_zS spectra. Little energy is present beyond about 30 hz in any of the spectra. The spectra in Figure 25 have been smoothed with a

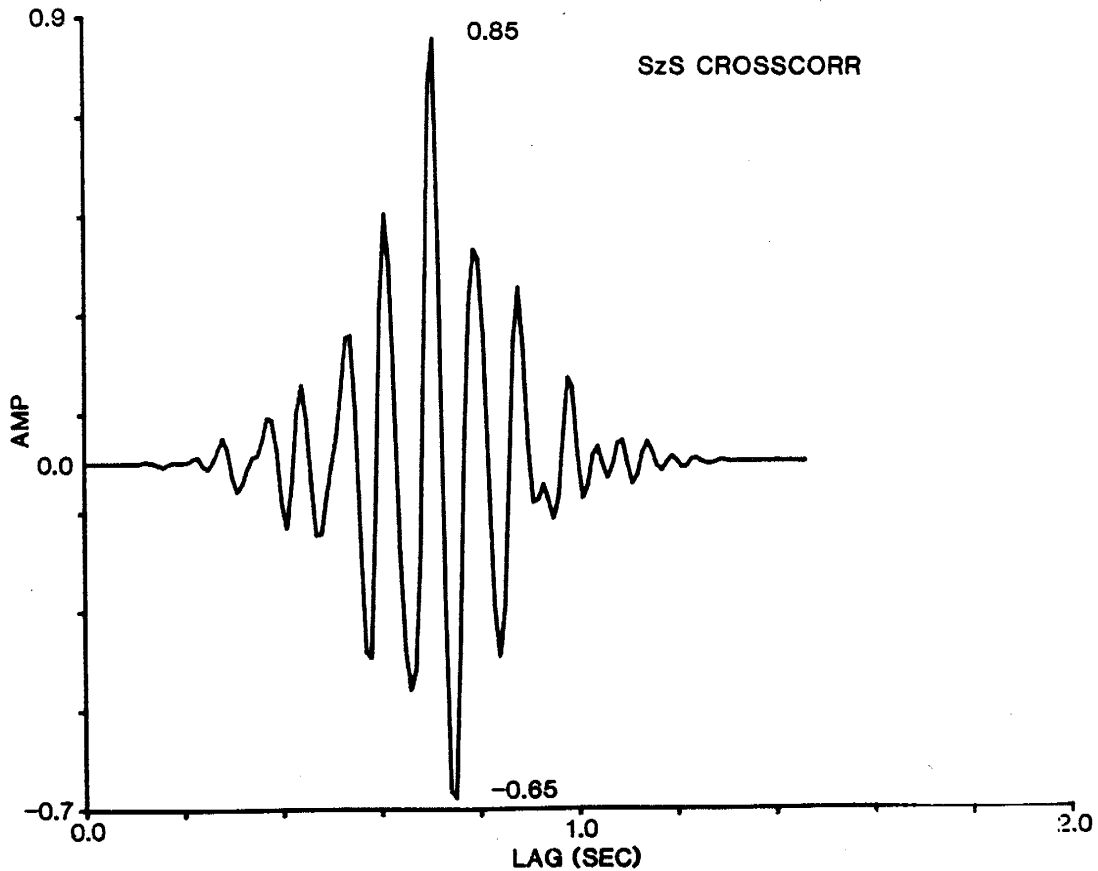


Figure 24a. Cross correlation function between S S phase windows of 114205 and 102733 events. Maximum and minimum values of function are indicated. Notice the almost perfect symmetry of the function about the maximum value.

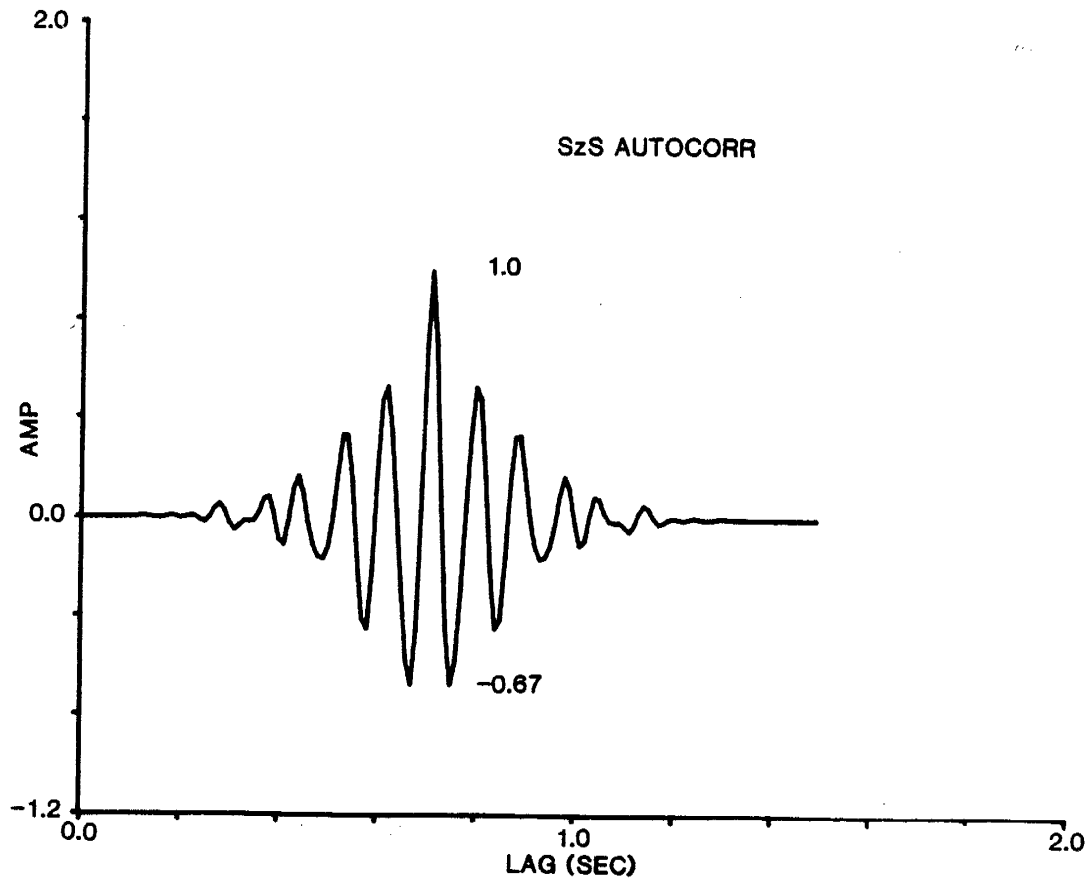


Figure 24b. $S_z S$ phase autocorrelation function, 102733 event.

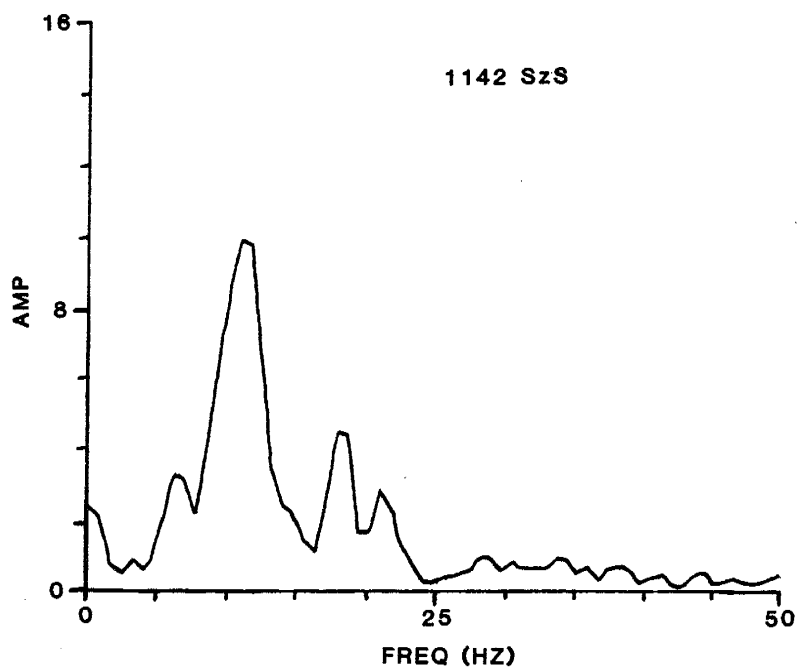
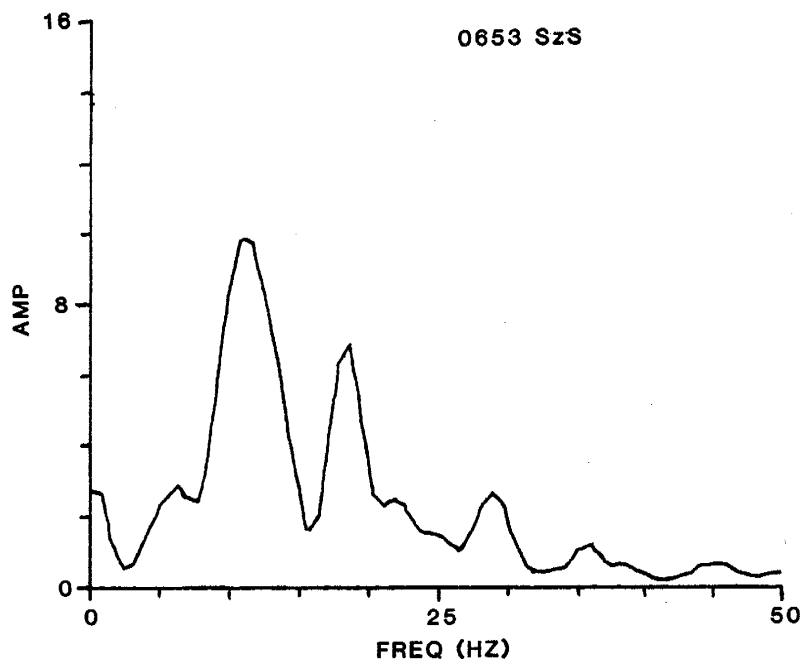


Figure 25. Spectra of S_zS phase windows, 114205 and 102733 events. Spectra are for ground velocity. A three point moving average filter has been applied.

three point moving average filter. The consequences of spectral smoothing are discussed in Appendix I.

Shown in Figure 21 are 0.7 second windows of the Figure 19 seismograms containing the direct S phase. Correlation coefficients relative to the last trace on the figure range from 0.85 to 0.99 (with an average of 0.93). Figure 26a is an illustrative cross-correlation function for two S phase waveforms. Again comparing the cross-correlation function to the auto-correlation function for one of the waveforms (Figure 26b), the overwhelming similarity in the waveforms is evident. The spectral characteristics of the S phase are shown in Figure 27 for two of the S phases. As with the S_zS phase, a strong peak near 11 hz is noted with very little energy present beyond about 20 hz.

A visual comparison of Figures 20 and 21 indicates significant similarity exists between the S and S_zS phases. Cross-correlation between these phases yields coefficients which are positive and large in absolute value. For example, cross-correlation of the S-phase of the event at 11:42 on June 2 (last trace on Figure 21) with all seven of the S_zS phases in Figure 20 yields values that are all positive and range from 0.76 to 0.81 with an average of 0.79. Figure 28 is an example of the cross-correlation function between the S and S_zS phases. Notice how similar the function appears to the S and S_zS cross and auto-correlation functions shown in Figures 24 and 26. Figure 29 further demonstrates this similarity by comparing the amplitude spectra of the S and S_zS phases.

The large correlation coefficients between S and S_zS phases indicate that for incident shear energy, the reflection must come from a

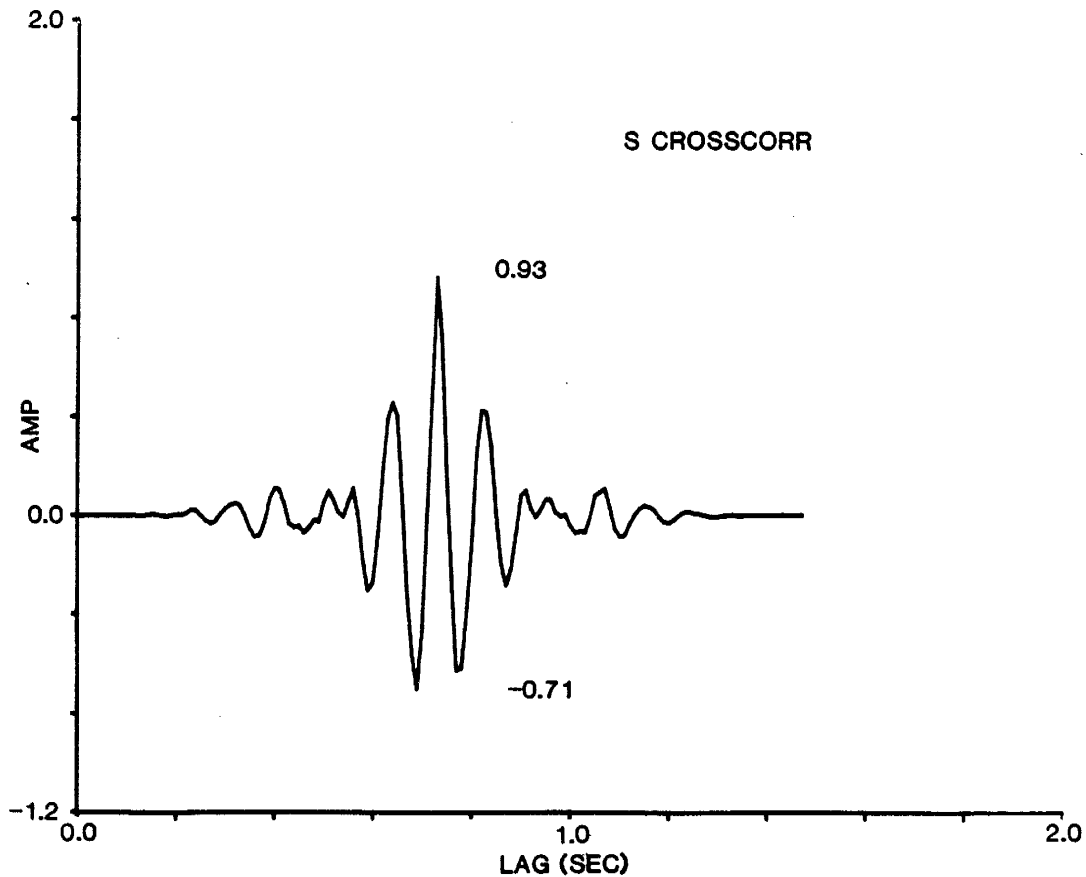


Figure 26a. Cross correlation function between S phase windows of 0653429 and 114205 events. Maximum and minimum values of function are indicated. Notice the almost perfect symmetry about the maximum-value.

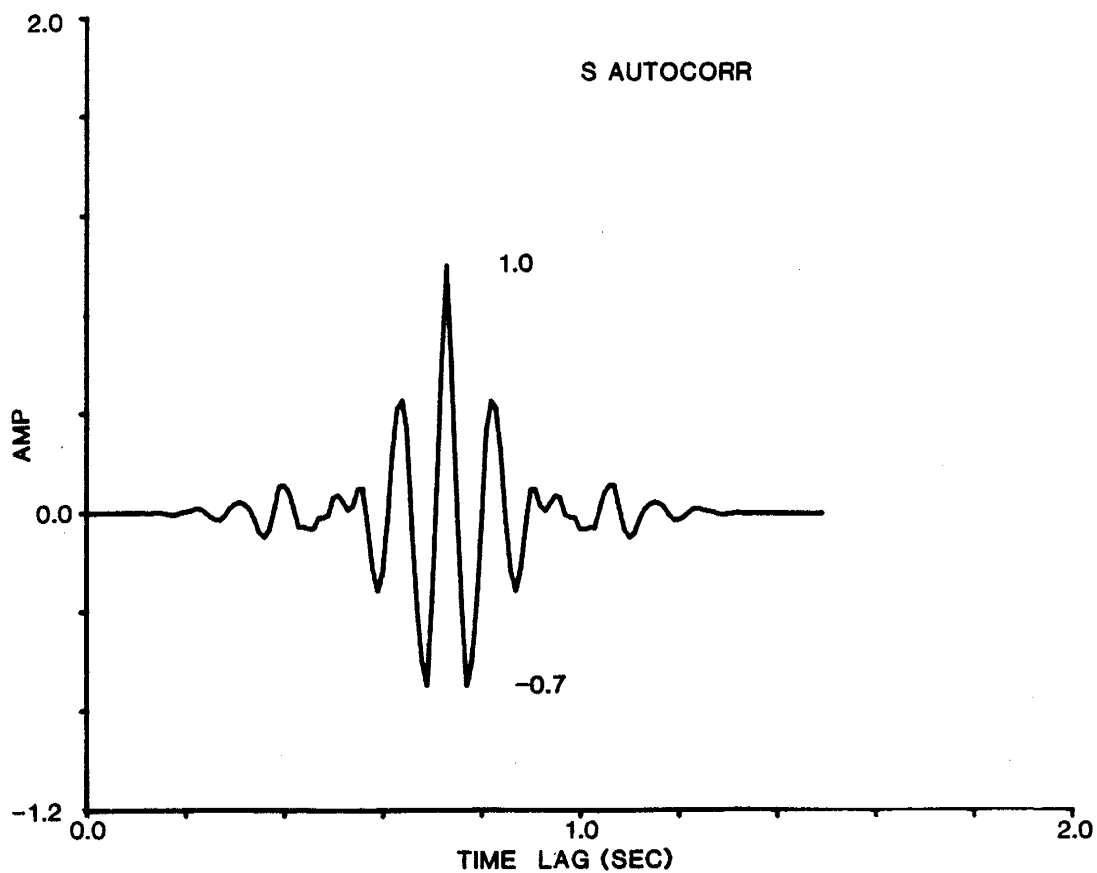


Figure 26b. S₂S phase autocorrelation function of 065429 event. Maximum and minimum values of function are indicated.

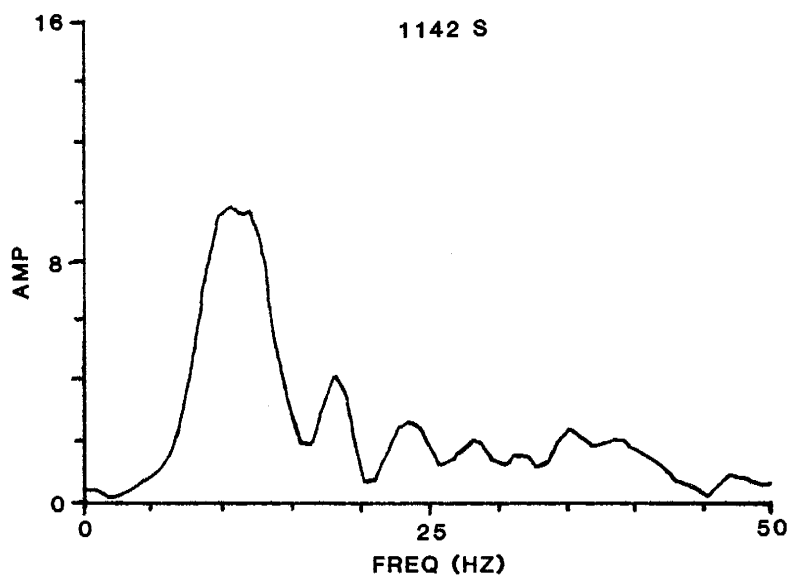
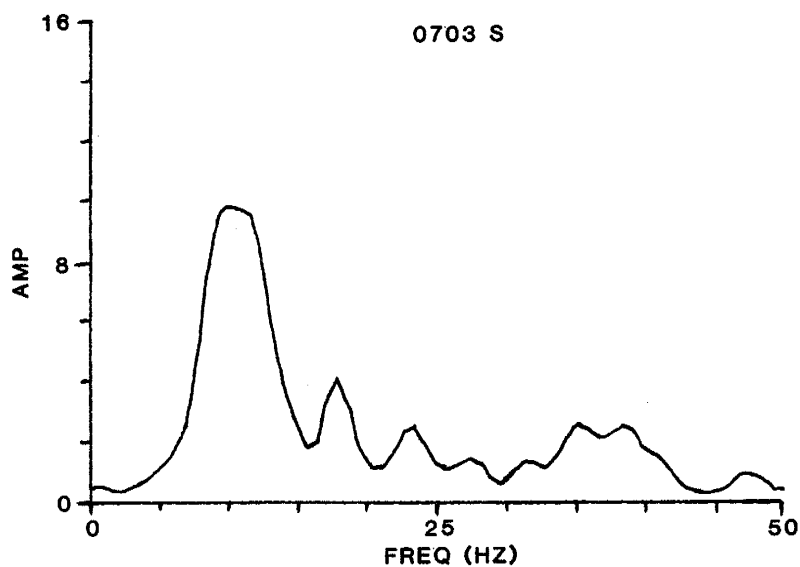


Figure 27. Spectra of S phase windows, 114205 and 070327 events. Spectra are for ground velocity. A three point moving average filter has been applied.

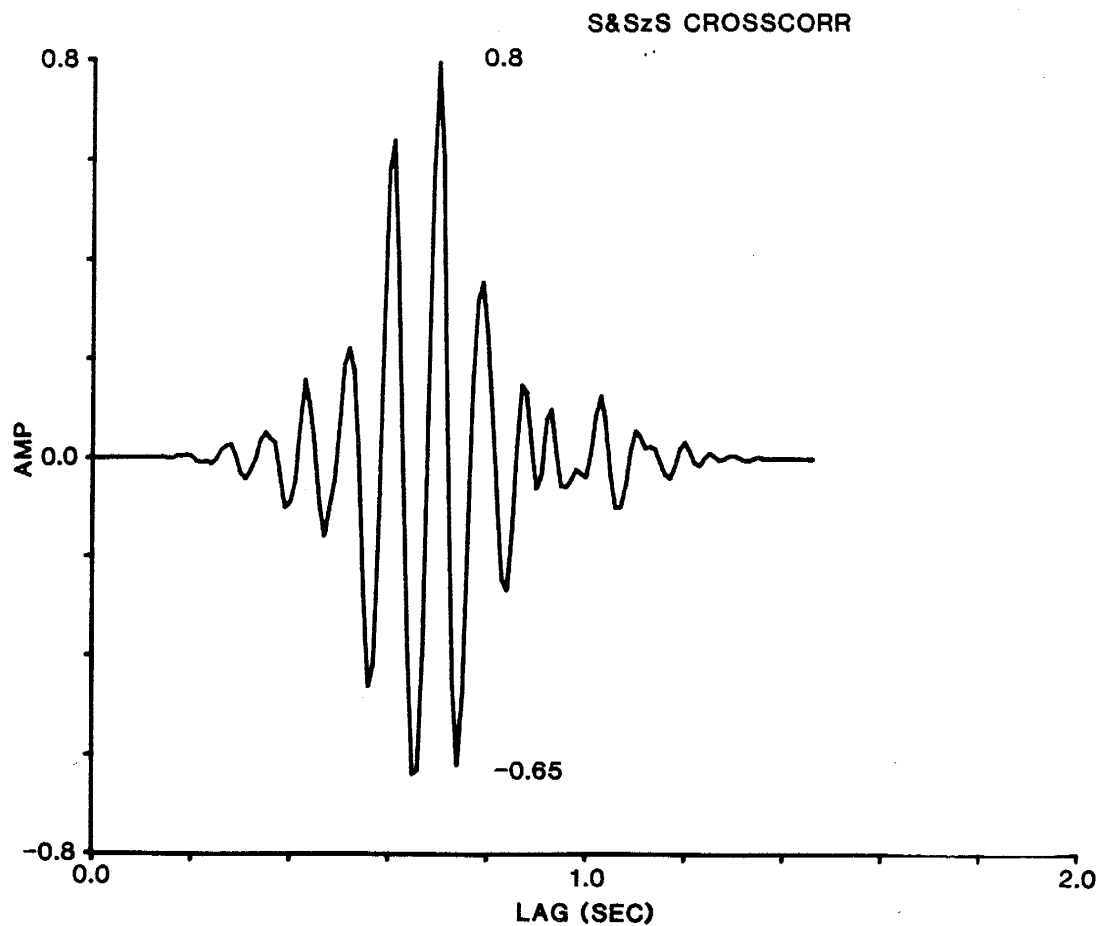


Figure 28. Cross correlation function between S and S_zS phase windows of 102733 event. Maximum and minimum values of function are indicated. Notice similarity to Figures 24a and 26a.

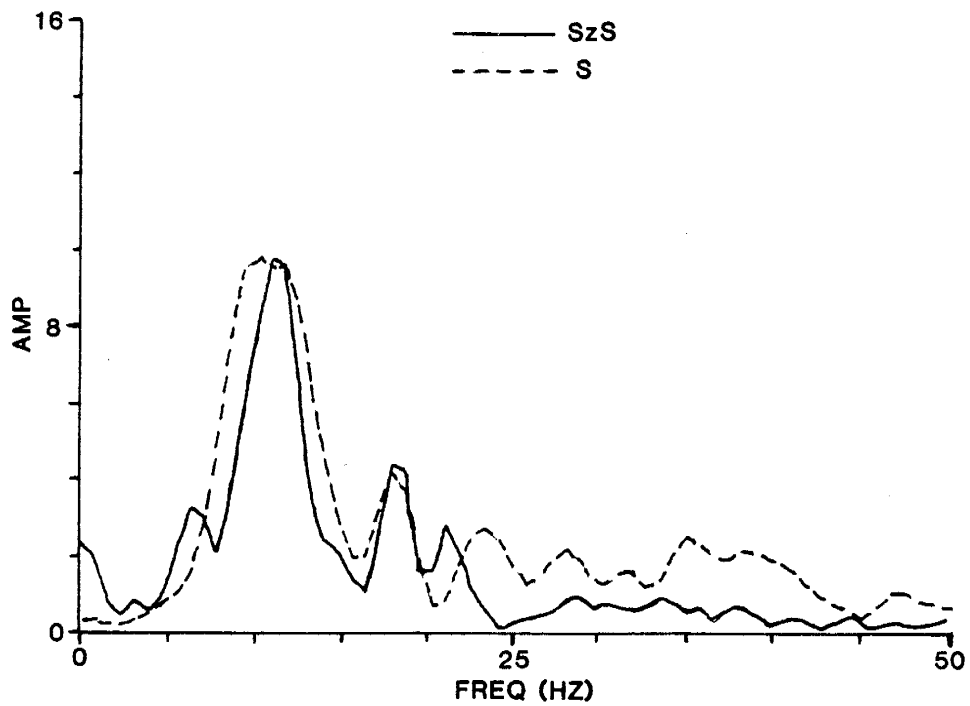


Figure 29. Comparison of S and S_zS phase spectra, 114205 event. Spectra are for ground^z velocity. A three point moving average filter has been applied.

single, simple velocity discontinuity. For instance, reflection from the top and bottom of a thin bed will tend to differentiate the waveform [Anstey,1979]. This differentiation will lead to a pronounced change in the appearance of the time series and the spectra.

Recall from Figure 18 that at the foci, the downgoing ray (S_zS) has the opposite sense of shear to the upgoing, direct ray (S). Therefore the positive correlation coefficients between these two phases require that the velocity discontinuity at depth produce a 180° phase shift of the reflected waveform. For shear energy this phase reversal can be shown to occur at the boundary between a rigid and non-rigid material, as at a solid-magma interface (see Appendix III).

The strengths of the S_zS reflections relative to the direct S phase (Figure 19) are nearly the same for all events within the swarm (Table 2). This implies that focusing effects are minimal and the upper surface of the magma body is essentially planar over a distance somewhat greater than that covered by the reflections. Hypocentral migration appears to be minimal for this swarm and the spread in reflection points no more than ≈ 500 m [Sanford et al., 1984].

Table 2. Amplitude Ratios of Direct and Reflected Phases at Station SC with Associated Standard Deviations

$\frac{S}{S_z S}$	10.52 ± 1.68 (n=7)	8.92 ± 2.84 (n=18)*
$\frac{P}{P_z P}$	0.04 ± 0.34 (n=5)	—
$\frac{P}{S}$	0.19 ± 0.08 (n=5)	0.32 ± 0.21 (n=18)*

*These ratios are for all well recorded events at station SC even if magnitudes are >0 and $P_z P$ not well recorded.

P_zP Reflections

The P_zP phase is a strong phase that occurs 3.66 seconds after the direct P phase (Figure 19). A panel showing the P_zP reflections of Figure 19 in detail is presented in Figure 22. Correlation coefficients between the last trace and all other traces are generally lower than those for the direct and reflected S phases, 0.43 to 0.77 (average of 0.60). Several features of the waveforms in Figure 22 are notable. First, although this phase clearly has a greater signal to noise ratio than does the S_zS phase (Figure 19), the cross-correlation coefficients for the P_zP waveforms are much lower than those of the S_zS phase. Second, notice the high-frequency, low-amplitude oscillations at the beginning of six of the eight traces in Figure 22. These may be the result of reflection from a thin, high-velocity lid to the discontinuity. Third, note the relatively high frequency content of the P_zP phase.

Figure 30 contains illustrative cross and auto-correlation functions for the P_zP waveforms. Notice the distinct change in character of the P_zP correlation functions relative to those of the direct and reflected S phases. The very spike-like appearance and poorly developed decay envelope of the P_zP correlation functions are indicative of impulsive waveforms with a broader spectral bandwidth present.

Shown in Figure 23 are windows of the direct P phase for five events shown in Figure 19. Correlation coefficients for the 0.40 second windows are all positive and range from 0.69 to 0.81 with an average of 0.75. Also indicated on the figure are correlation coefficients for shorter windows, neglecting most of the P phase coda. Correlation coefficients for these shorter (0.13 second) windows average 0.85. The

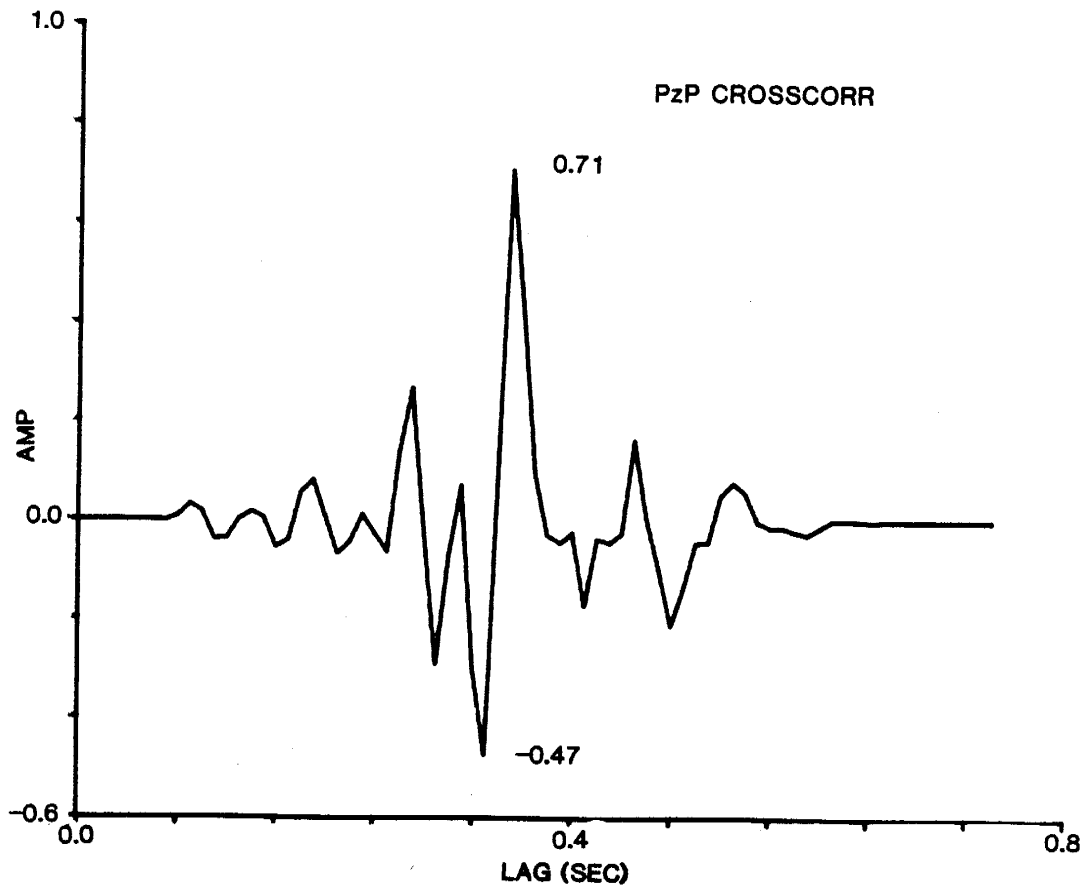


Figure 30a. Cross correlation function between P_zP phase windows of 102733 and 120709 events. Maximum and minimum values of function are indicated. Notice the lack of symmetry of the function about the maximum value.

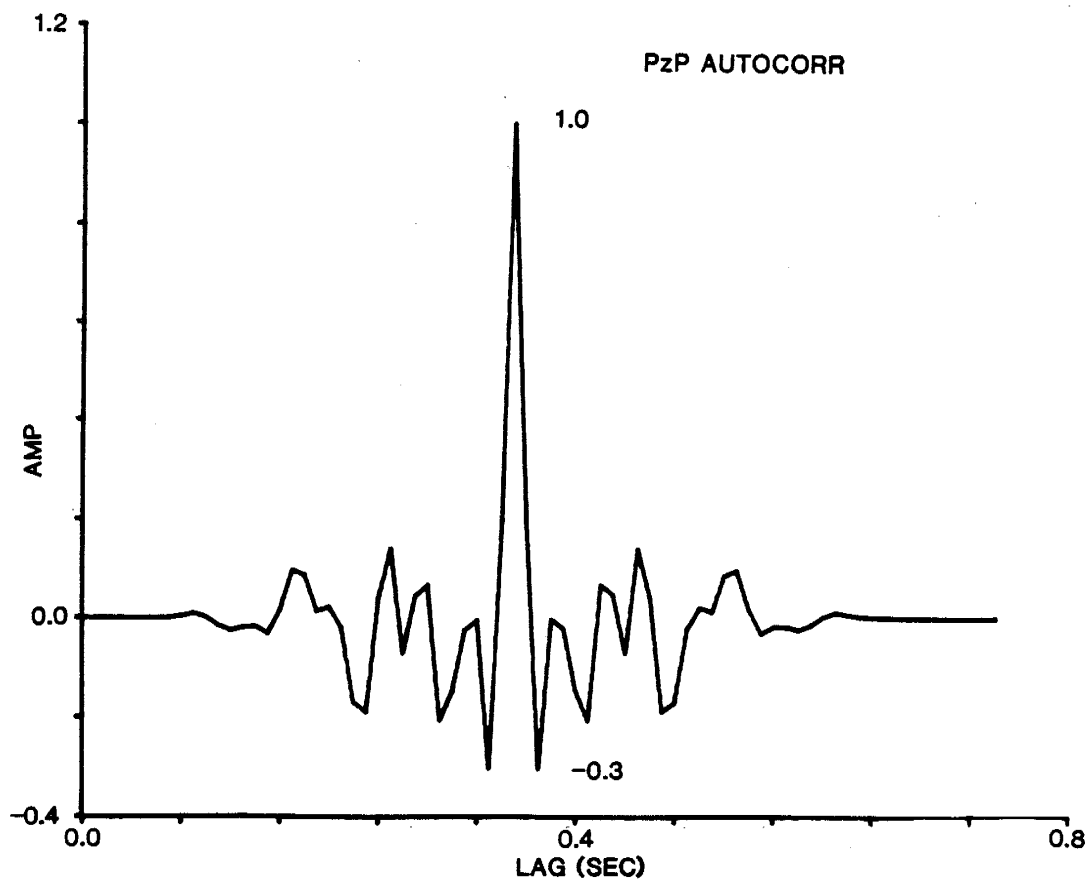


Figure 30b. P_zP phase autocorrelation function, 120709 event. Maximum and minimum values of function are indicated.

characteristics of the cross and auto correlation functions for the P-phase are illustrated in Figure 31. The character of the P-phase correlation functions is more like the P_zP correlation functions than those of either the direct or reflected S phases. The amplitude spectra of the direct P phase shows a distinctly broader band than does the S and S_zS spectra (Figure 32a). The spectra of the P and P_zP phases are compared in Figure 32b.

A visual comparison of Figures 22 and 23 does not show the remarkable similarity between the P and P_zP phases that was obvious between the S and S_zS phases. Cross-correlation between these two phases yields absolute values that range from 0.26 to 0.60 and with one exception, the signs of the correlation coefficients are negative, average=0.45 (see Figure 33). Recall from Figure 18 that the direct P and reflected P leave the focus in phase, the negative correlation coefficient indicates that they are out of phase at the recording station. This implies a reflecting interface that produces a 180° shift in phase, i.e., a discontinuity across which the acoustic impedance decreases. For normal incidence of compressional energy upon a plane interface, the reflection coefficient, r , simplifies to:

$$r = \frac{V(2) \rho(2) - V(1) \rho(1)}{V(2) \rho(2) + V(1) \rho(1)} \quad (1)$$

where V and ρ indicate velocity and density respectively. For values appropriate for mid-crustal depths in the Socorro area (to be discussed in more detail later) r becomes negative, i.e. a phase reversal occurs.

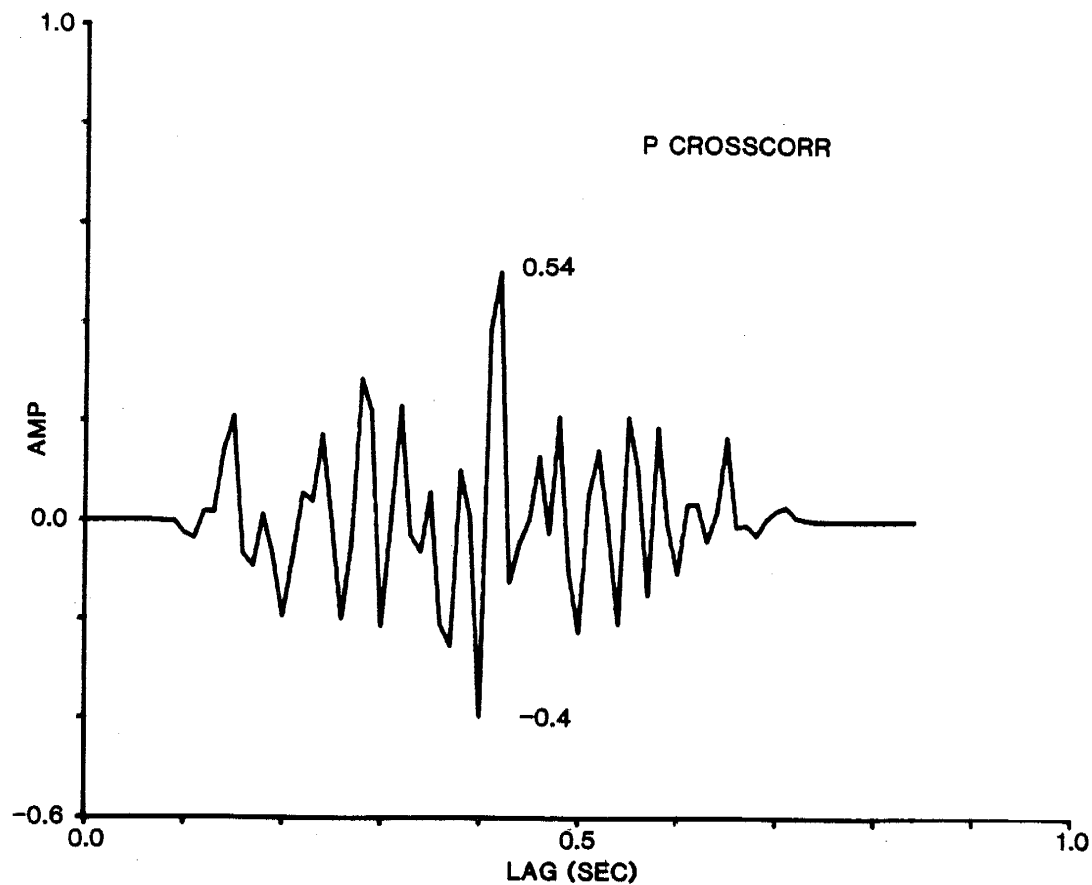


Figure 31a. Cross correlation function between P phase windows of 102733 and 072805 events. Maximum and minimum values of function are indicated. Notice the complexity of the function relative to the S phase cross correlation function (Figure 26a).

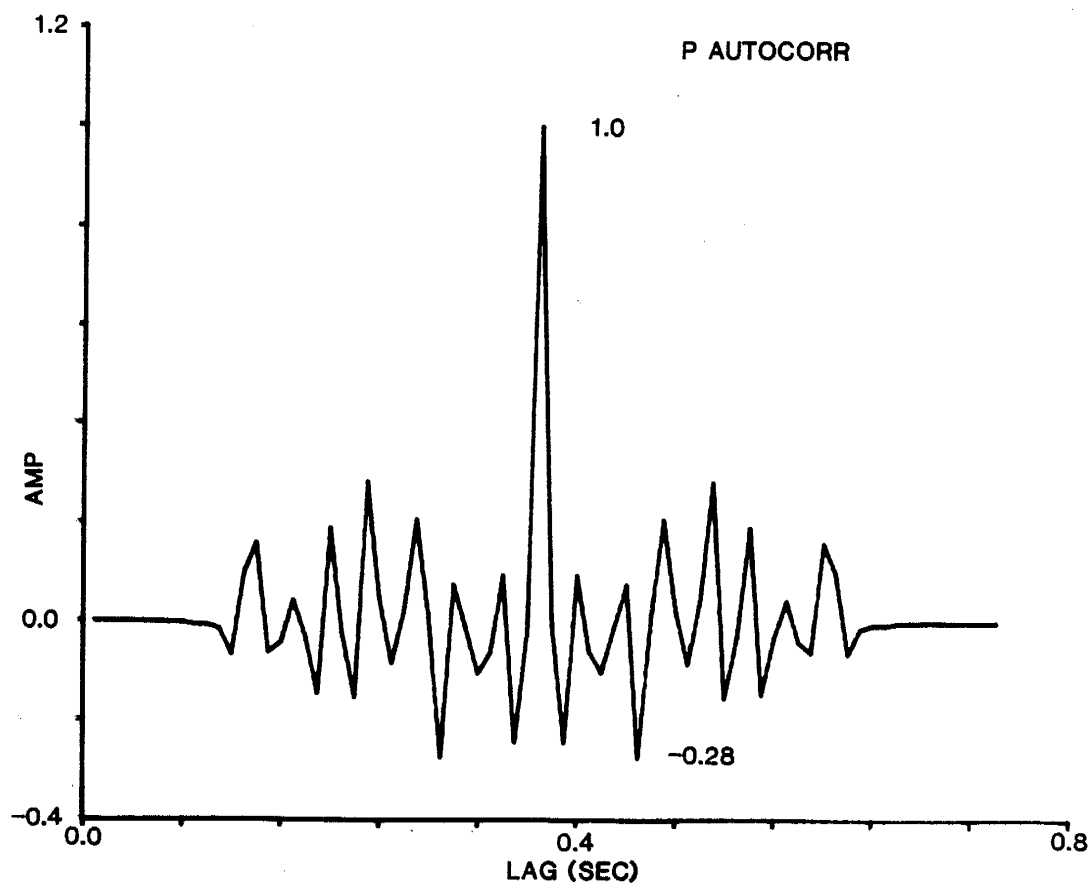


Figure 3lb. P phase autocorrelation function, 102733 event.
Maximum and minimum values of the function are indicated.

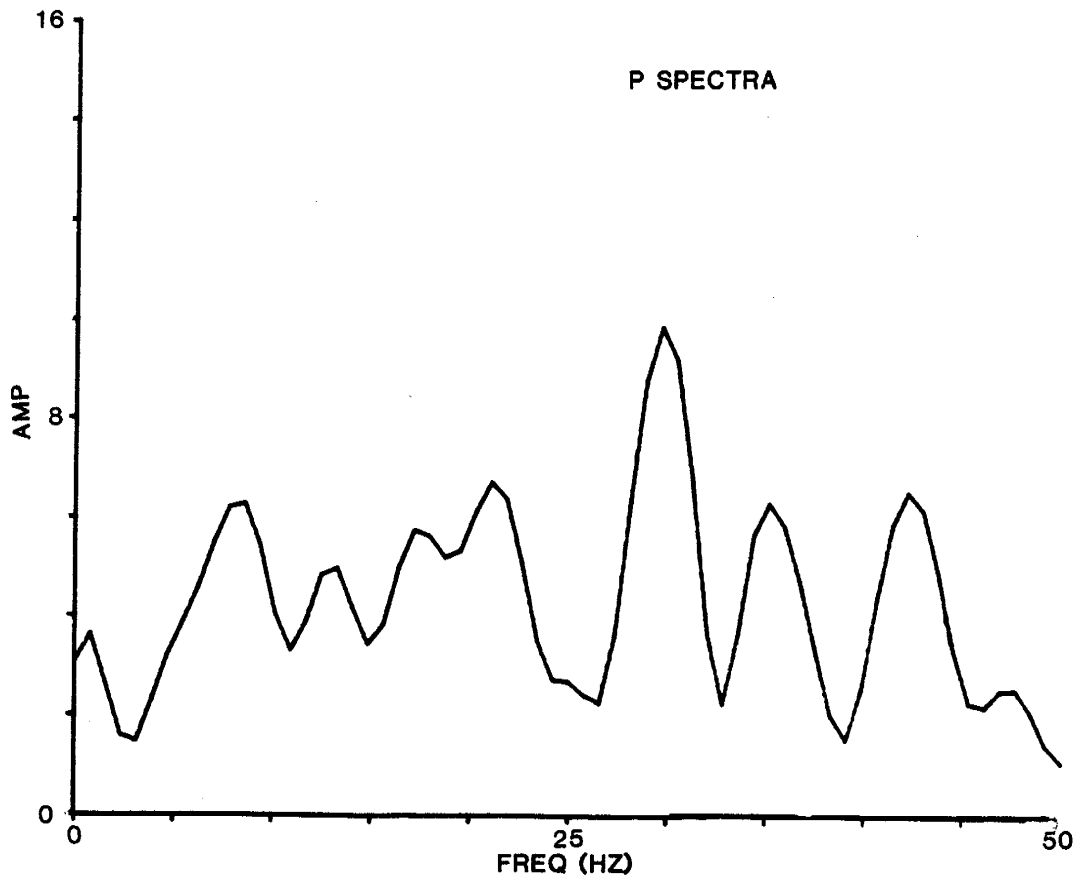


Figure 32a. Spectra of P phase, 102733 event. Spectra is of ground velocity. Window length is 0.4 seconds. A three point moving average filter has been applied.

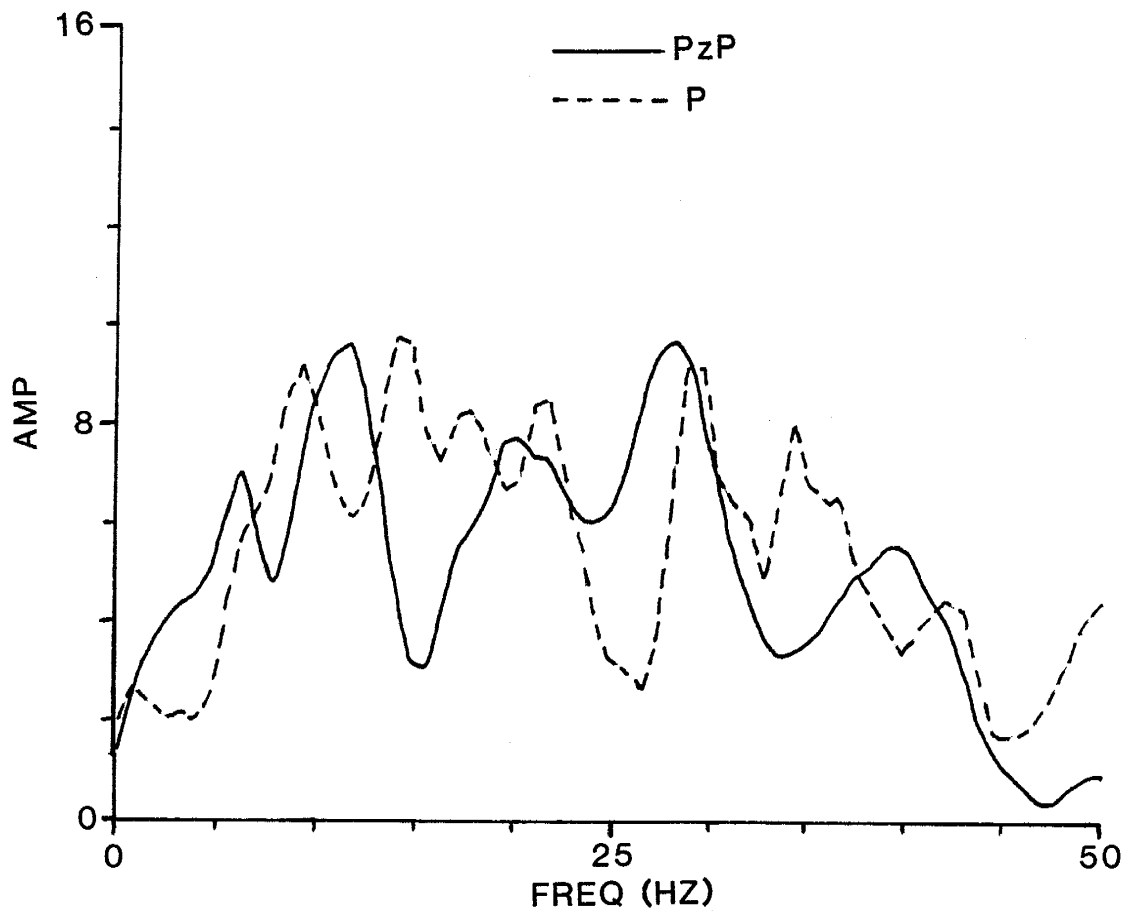


Figure 32b. Comparison of P and P_zP phase spectra, 114205 event. Spectra are of ground velocity. Window length is 0.7 seconds. A three point moving average filter has been applied.

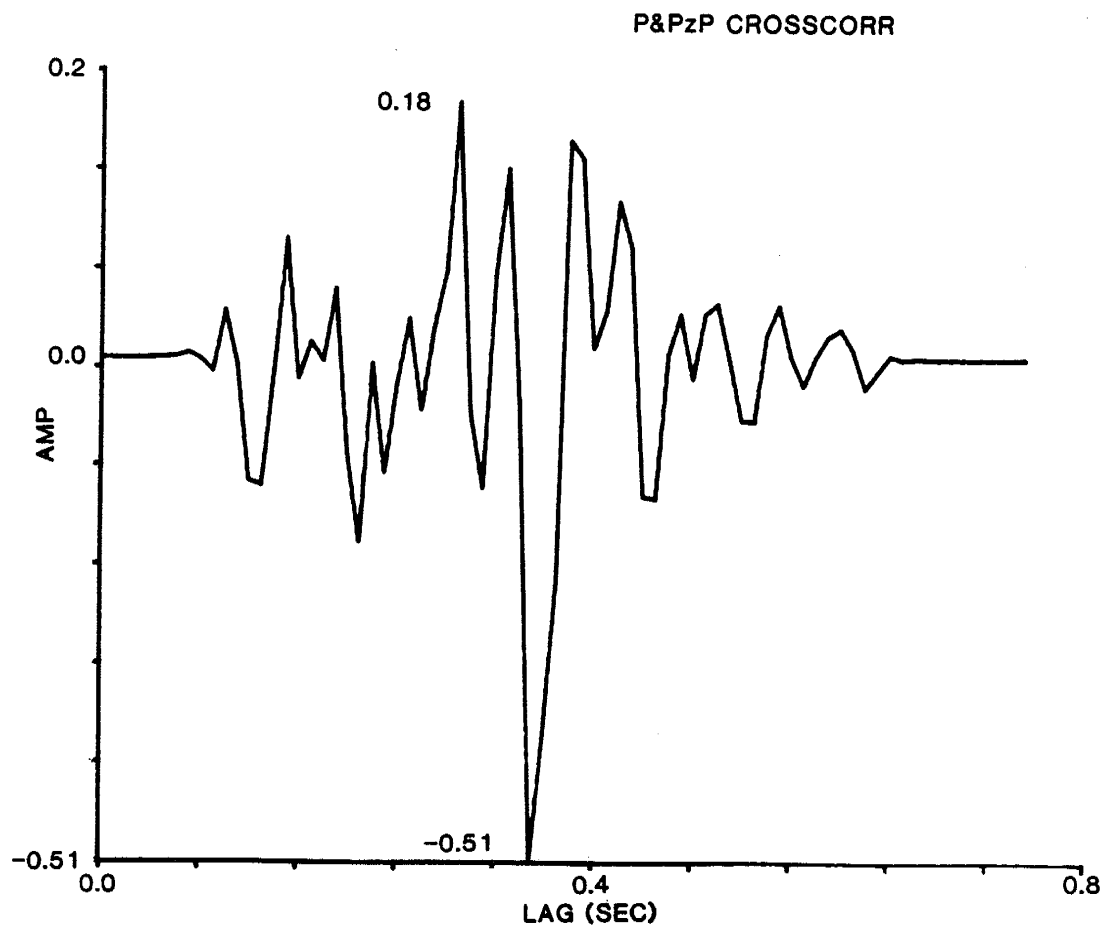


Figure 33. Cross correlation function between P and P_zP phase windows of 102733 event. Maximum and minimum values of function are indicated. Notice the lack of similarity to the S and S_zS cross correlation function (Figure 28).

Structure of the Magma Body

The high degree of similarity between S_zS phases indicates the upper surface of the reflector is relatively flat and planar. If any topography exists, its wavelength (curvature) is greater than the Fresnel zone of the incident S waves. The generally small correlation coefficients between P and P_zP phases for the same event and between P_zP for several events (Figure 23) indicates the P reflection is from a layered structure which has lateral variations in the thicknesses and/or velocities of the layer or layers. However, to be consistent with the observed characteristics of the S_zS waveforms, the first significant layer in any layered sequence must be non-rigid. This layer blocks further downward travel of shear energy and produces a reflected waveform expected for normal incidence on a singular discontinuity.

Regardless of change in internal structure, maximum limits can be placed on the thickness of the magma body from the duration of the P_zP waveform. Assuming a velocity of 2.6 km/sec for a full basaltic melt [Murase and McBirney, 1973], the duration of the P_zP signal, 0.30 ± 0.03 seconds (1 s.d.), indicates a maximum thickness of 380 meters for a single layer full-melt model. If the body is assumed to consist of alternating layers of full melt and solid rock in about equal proportions, the maximum thickness is about 500 meters.

Comparison With COCORP Reflections

The general characteristics of COCORP reflections and microearthquake P_zP reflections are similar. When clearly recorded in the COCORP studies, the mid-crustal reflection consists of four to six oscillations in a time span of 0.25 to 0.38 seconds (Figure 34) [Brown et al., 1980; Brocher, 1981]. The P_zP microearthquake waveforms which appear to be most similar to those in the COCORP studies are traces presented in Figure 35. These traces have four to six oscillations in 0.27 to 0.32 seconds. The COCORP studies considered reflections recorded 30 and 45 km northeast of Socorro, whereas the microearthquake P_zP reflections considered in this study were recorded 16 km southwest of Socorro (see Figure 5) in an area where the initial interpretation of COCORP line 3 failed to show the presence of a strong mid-crustal reflector. The reflection data reported here are interpreted as indicating that the mid-crustal magma body is present in the South Canyon area southwest of Socorro. As defined by the earlier results of Rinehart et al. [1979] however, both the COCORP study areas mentioned above and the microearthquake data reported here are near the margin of the magma body. The similarity in the durations of the microearthquake and COCORP reflected signals implies a similar overall thickness for the magma body at these widely separated points.

The amplitude spectra for the microearthquake P_zP reflections (Figure 36) are somewhat more complex than those for the COCORP reflections [Brown et al., 1980; Brocher, 1981]. This is in part due to the different bandwidths of the two sources. Figure 37 contains exam-

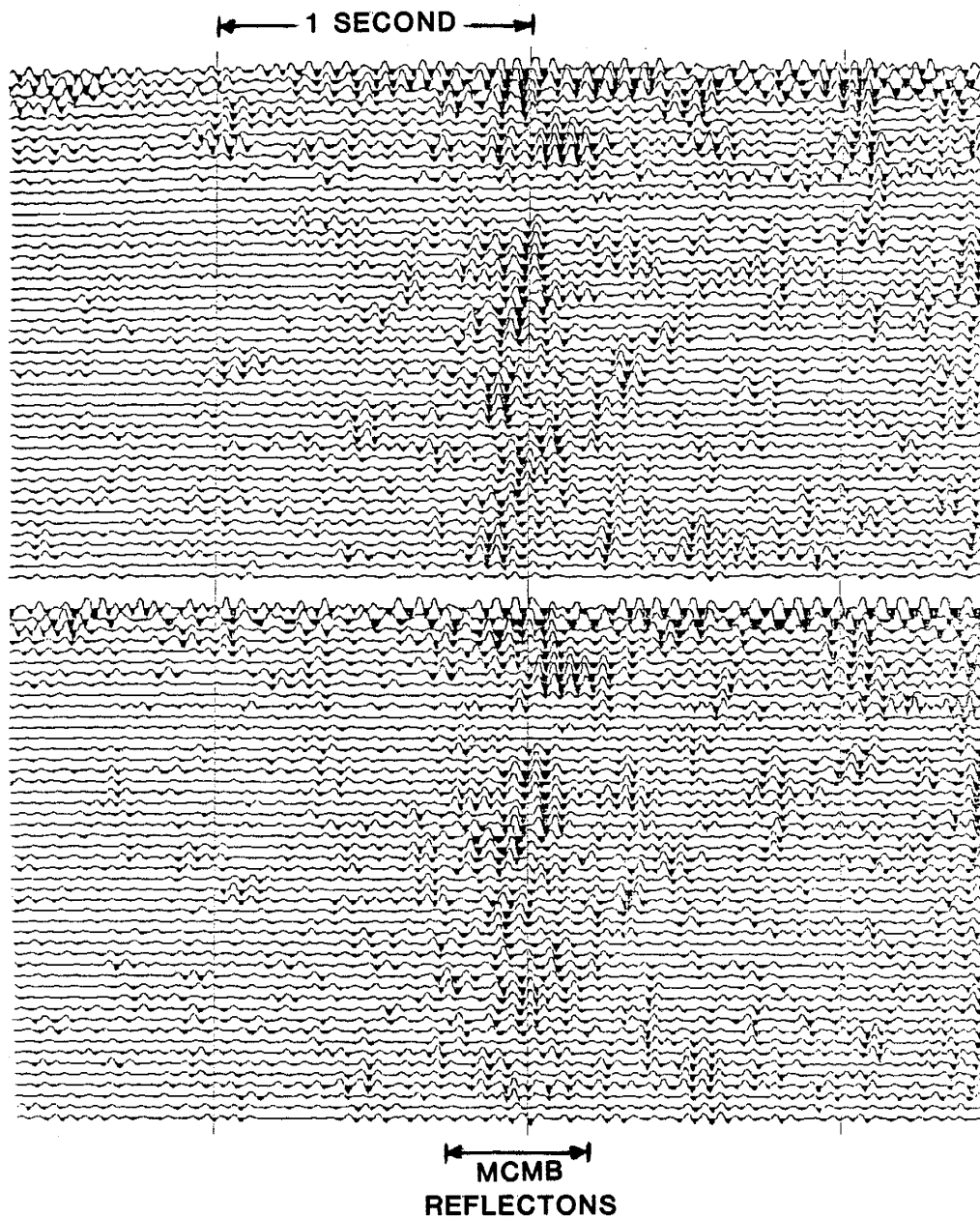


Figure 34. Example traces showing COCORP mid-crustal magma body reflections. One second of record time is indicated on figure. Data has been stacked but is otherwise unprocessed. Data is from line 1A, near VP130 (data provided by L. Brown).

PzP

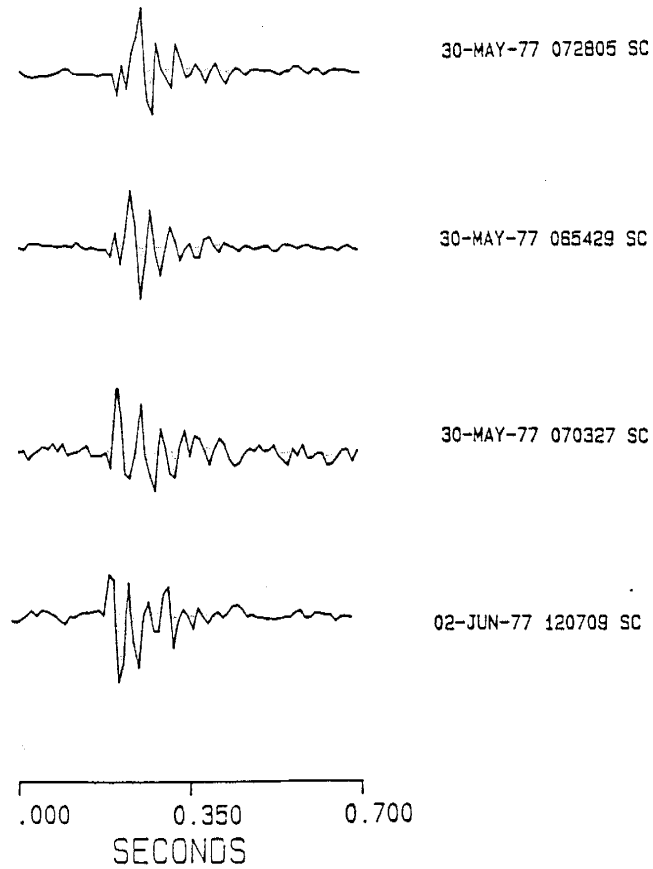


Figure 35. P_zP microearthquake reflections for comparison to COCORP mid^z-crustal magma body reflections.

ples of raw spectra of the mid-crustal reflection from unstacked COCORP line 2A data for comparison with Figure 36. Six of the eight spectra for the P_zP reflections have three well-defined peaks with average frequencies of 9.6 ± 2.5 (1 s.d.), 25.5 ± 1.6 (1 s.d.), and 43.9 ± 3.1 (1 s.d.) hz. The first and third peaks of these microearthquake P_zP spectra would not be observed in the spectra of the COCORP reflections because the VIBROSEIS (Trademark Continental Oil Co.) sweeps were from 10 to 32 hz. However, the COCORP reflectivity data does show an obvious spectral peak at ≈ 26 hz [Brocher, 1981] which agrees with the middle of the three peaks in the microearthquake spectra. The general character of the raw COCORP spectra (Figure 37) is also similar to the microearthquake P_zP spectra. At least some of the differences in the raw spectra are due to the signal passing through different near surface materials.

The most detailed analysis to date of the COCORP Socorro mid-crustal magma body reflections is by Brocher [1981]. Brocher calculated the reflectivity of the mid-crustal reflector as a function frequency for COCORP data on line 1. An estimate of the frequency dependence of the magma body reflectivity can be obtained from the microearthquake data for comparison if certain assumptions are made. First, assume that no significant reflectors exist between the focal depth and the depth of the mid-crustal reflector. Both the COCORP and microearthquake data tend to support this assumption [A. Sanford pers. comm.]. Next, assume that the events are small enough that source directivity will only effect the relative spectral levels of the direct and reflected phases, i.e., ignore any source induced spectral shifts. The validity of this assumption will be examined in

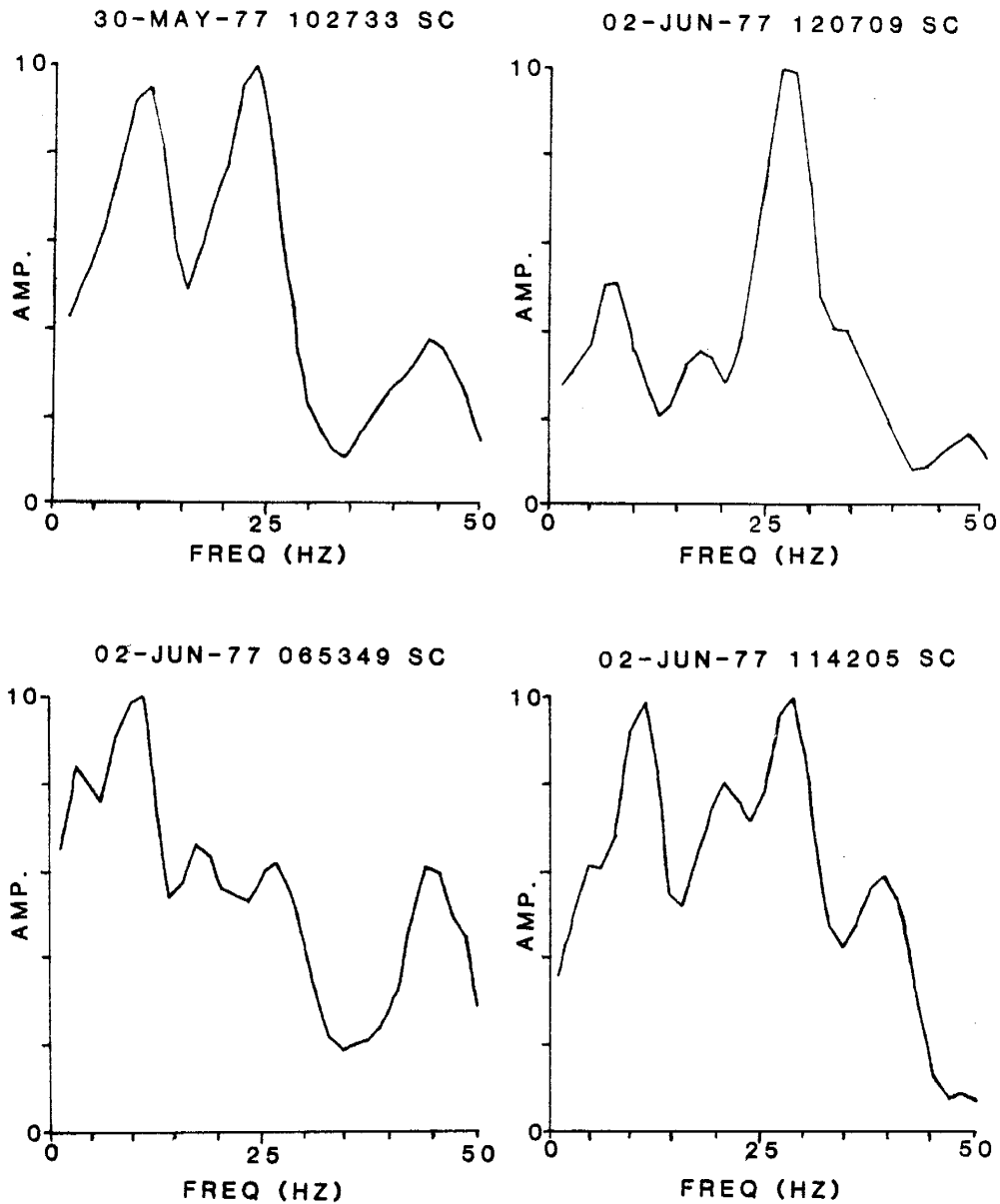


Figure 36a. Spectra of the reflected P phase waveforms shown in Figure 22. Spectra are for ground velocity. A three point moving average filter has been applied.

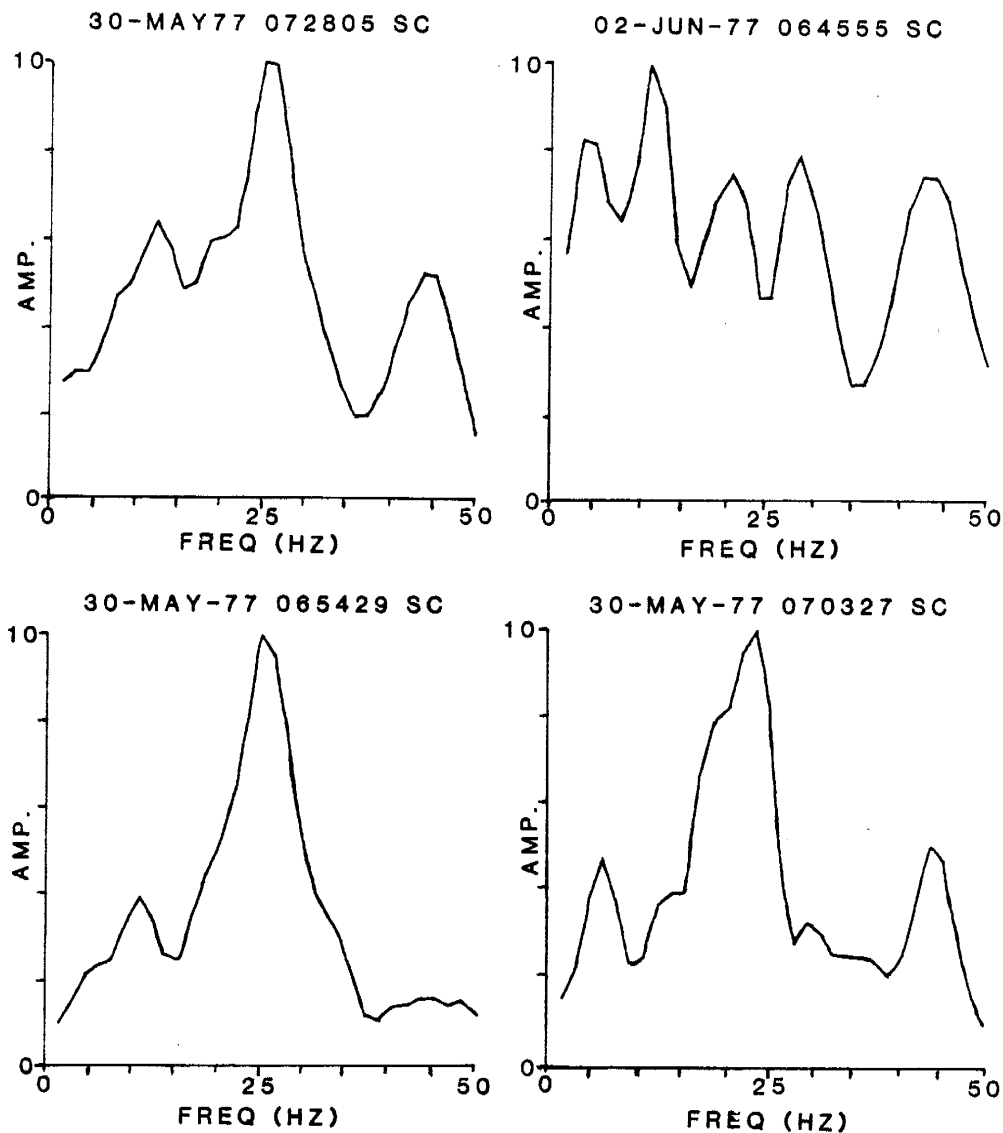


Figure 36b. Spectra of the reflected P phase waveforms shown in Figure 22. Spectra are for ground velocity. A three point moving average filter has been applied. Six of the eight spectra show a prominent peak near 26 hz.

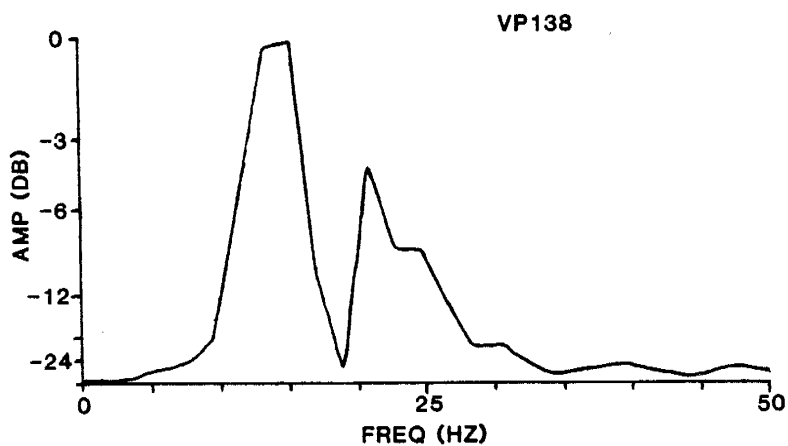
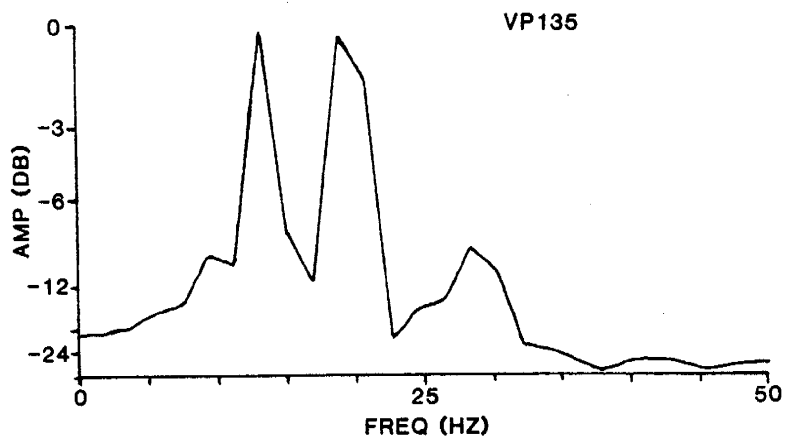


Figure 37. Spectra of COCORP mid-crustal magma body reflections from line 1A. Spectra are for ground velocity. Notice the similarity to the microearthquake P_zP spectra in Figure 36 (from L. Brown).

detail in the next section. The ratio of the normalized amplitude spectra of the reflected to direct P-phase will then be an estimate of the reflectivity function ($R_f(f)$). No absolute reflection coefficients can be obtained in this fashion of course, but the shape of the reflectivity curve should be accurate. This estimate of the magma body reflectivity in the South Canyon area is compared to the reflectivity on COCORP line 1 (from Brocher, 1981) in Figure 38. Within the limits of the vibrator sweeps, considerable similarity is evident between the two spectra. This implies a similar response for the magma body at two different locations.

The peak frequencies in the P_zP spectra occur roughly in the arithmetic sequence 1:3:5. This suggests the mode sequence of an open pipe which occurs in the resonance of a layer, i.e. a constructive interference phenomena [Cranswick et al., 1986]. The simplest model to explain the interference would be a single layer of magma with a thickness equal to one-quarter wavelength of the signals at the lowest frequency peak. The expression for the quarter wavelength resonance frequency of a resonating layer is given by:

$$f_{\frac{1}{4}} = \frac{V_l}{4 d_l} \quad (2)$$

where V_l and d_l are the velocity and thickness of the layer respectively. A 75 meter thick layer with a velocity of 2.6 km/sec, the full-melt velocity of Columbia River basalt (CRB) [Murase and McBirney, 1973], would produce spectra with peaks at 8.7, 26.0, and 43.3 hz. This method of obtaining the spectral response of a layer or layers,

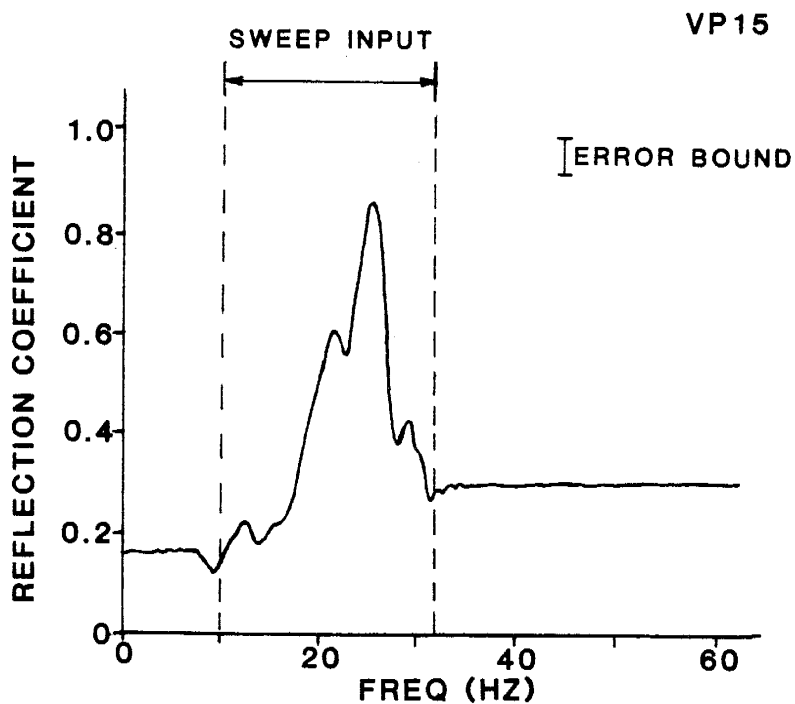
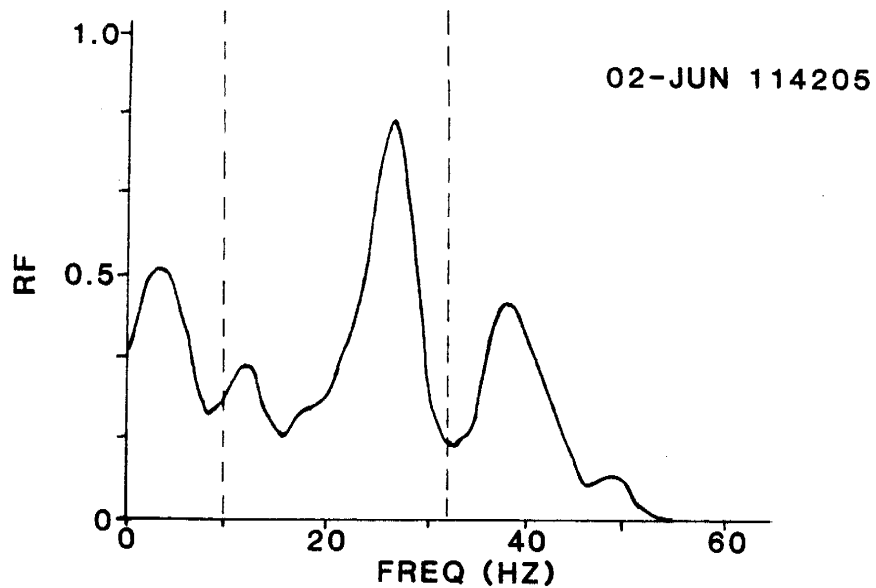


Figure 38. Comparison of estimated reflectivity of micro-earthquake (top) and COCORP (bottom) P_zP data. Note the prominent peak near 26 hz in both spectra. Lower figure is from Brocher (1981). Location of VP15 is indicated in Figure 5.

suggests a different method of calculating synthetic spectra.

A possible problem with the single layer interpretation is whether the observed relative strengths of the spectral peaks fit the model. This can only be confirmed after corrections are made for (1) attenuation over the total travel path, (2) attenuation within the magma body, (3) interaction of the reflected signal with large velocity discontinuities near the surface, (4) instrument response, and (5) the effect of any high velocity lid to the magma body. These factors were addressed by constructing synthetic seismograms to compare to the observed data.

SOURCE INDUCED SPECTRAL SHIFTS

The question of whether or not the direct and reflected phases have the same spectral characteristics leaving the source region arises in both the calculation of reflectivity (previous section) and in modelling (next section). This question will be addressed by consideration of both theory and observational results.

Failure on a fault surface emits seismic energy that travels outward from the point of failure with an amplitude and spectra which depend on the characteristics of the rupture process. Specifically, whether the rupture proceeds in a bilateral or unilateral fashion can have a pronounced azimuthal effect on the energy leaving the focal region. Analysis of the amplitude ratio of S and S_zS phases for this swarm indicates that rupture was dominantly unilaterally downwards [Sanford et al., 1984]. The question of whether or not the azimuthal effects are significant enough to consider when comparing the direct and reflected phases must then be considered.

It can be shown [Aki and Richards, 1980] that the far field radiation of both the P and S waves depends on the integral of the time derivative of the slip function on the fault. For the P wave it is possible to write the radiation function (Ω) as:

$$\Omega(x,t) = \int_{\Sigma} \mu \left[\dot{u}(x, t - \frac{r}{V_p}) \right] d\Sigma \quad (3)$$

where μ is the rigidity, r is radial distance and V_p is the P wave velocity. The far field radiation depends on both the temporal variation of the slip function (u) as well as its spatial variation over the rupture

surface Σ . It can also be shown that the long period portion of the displacement spectrum is proportional to the seismic moment, M_0 [Brune, 1970]. The moment is a fundamental parameter describing earthquakes.

Consider the case of unilateral rupture as diagrammed in Figure 39. The fault length is L and width is W (with $L \gg W$). Assume that the rupture front includes the entire fault width and ruptures in the $+x$ direction. Assuming that the displacement on the fault is the same function everywhere and that the rupture velocity is V_r , then the rupture propagation can be specified by

$$u = d\left(t - \frac{x_1}{V_r}\right). \quad (4)$$

Then the final offset is

$$\lim_{t \rightarrow \infty} d(t) = D, \quad (5)$$

and the moment is given by

$$M_0 = \mu \cdot L \cdot W \cdot D. \quad (6)$$

Define

$$r_o \gg L, \quad (7)$$

then the far field P wave radiation becomes

$$\Omega(x, t) = \mu W \int_0^L \left[d\left(t - \frac{x_1}{V_r} - \frac{r_0}{V_p} + \frac{x_1 \cos(\theta)}{V_p}\right) \right] dx_1 \quad (8)$$

[Aki and Richards, 1980]. The spectral description of the P-wave

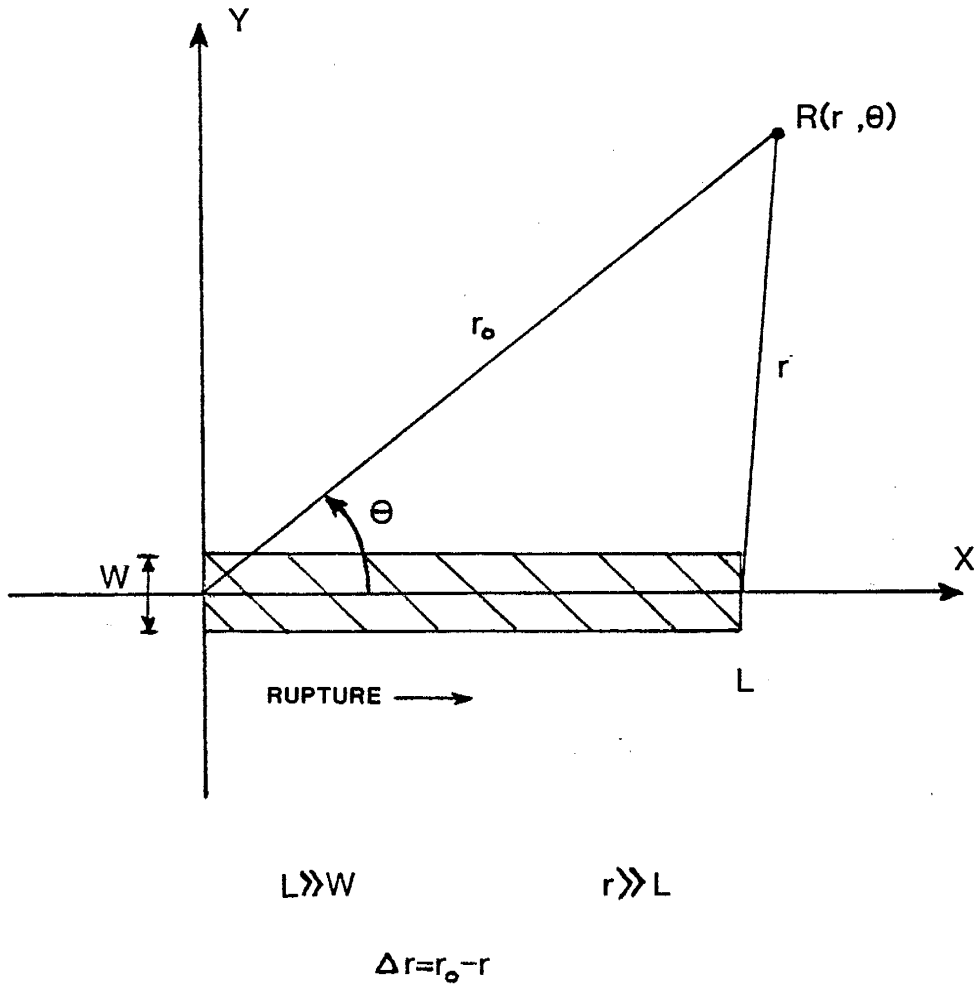


Figure 39. Geometrical relationships used in the derivation of spectral shift due to unilateral rupture. Rupture begins at the origin and proceeds in the +x direction. R is the observation location.

radiation is then

$$\Omega(\bar{x}, \omega) = \omega d(\omega) W L \frac{\sin(X)}{X} \exp \left[i \left(\omega \frac{r_0}{V_r} - \frac{\pi}{2} + X \right) \right] \quad (9)$$

where

$$X = \left(\frac{\omega L}{2} \right) \left[\frac{1}{V_r} - \frac{\cos(\theta)}{V_p} \right] \quad (10)$$

[Aki and Richards, 1980]. The finite nature of the fault leads to the sinc function behaviour (see Figure 40). This behaviour gives rise to a smoothing effect on the waveform [Bullen and Bolt, 1986].

Consider the time function

$$\tau_p = L \left[\frac{1}{V_r} - \frac{\cos(\theta)}{V_p} \right] \quad (11)$$

The function τ_p may be regarded as the time required to propagate down the fault at velocity V_r minus the time lag due to Δr at the velocity V_p (Figure 39) [Bullen and Bolt, 1986; Aki and Richards, 1980]. Consider the two extremal cases, when the receiver is in front of the fault, $\theta=0^\circ$, and

$$\tau_p = L \left[\frac{1}{V_r} - \frac{1}{V_p} \right]. \quad (12)$$

When the receiver is behind the fault (rupturing away from the receiver), $\theta=180^\circ$ and

$$\tau_p = L \left[\frac{1}{V_r} + \frac{1}{V_p} \right]. \quad (13)$$

If the displacement on the fault is a step function, then the

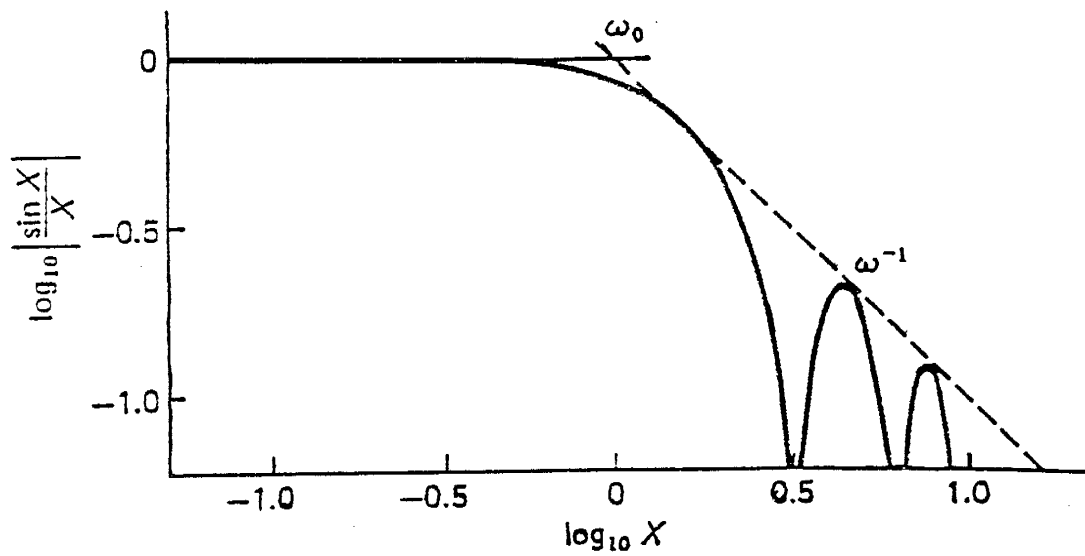


Figure 40. Illustration of the frequency domain behaviour of the sinc function. Notice the $1/\omega$ rolloff above ω_0 (after Bullen and Bolt (1986)).

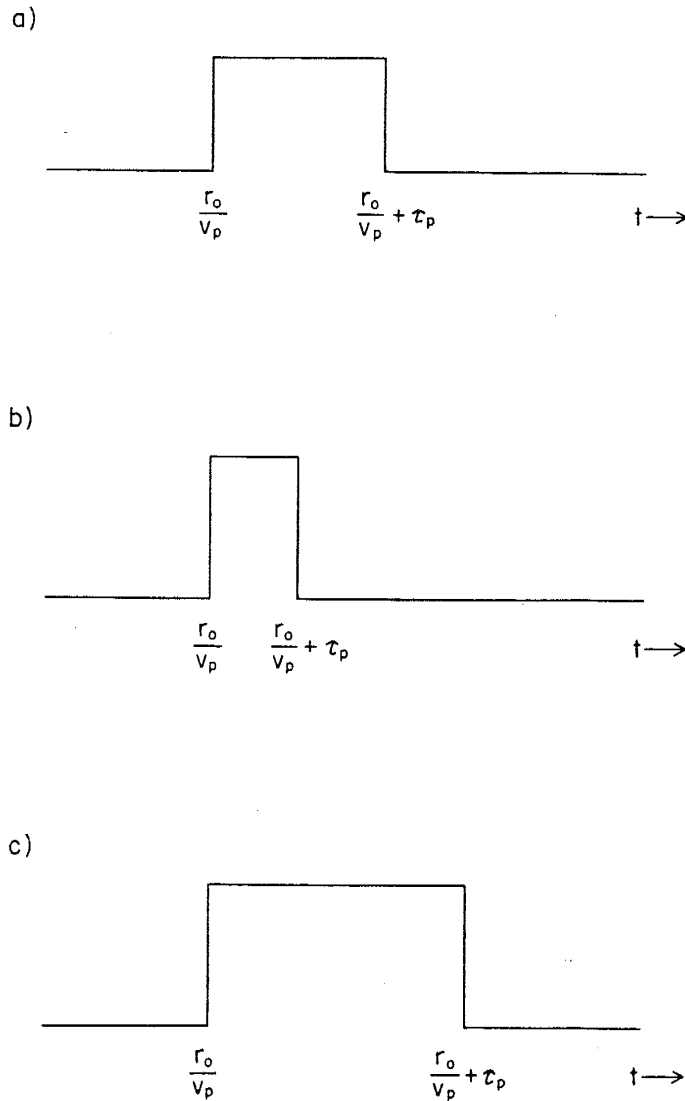


Figure 41. Dependence of pulse width on source-reciever azimuth. Step function displacement gives rise to a boxcar (a). When $\theta=0^\circ$, the function is narrow (b). When $\theta=180^\circ$, the function is broad (c).

observed signal from the fault is a boxcar (Figure 41a). Out in front of the fault where rupture is propagating towards the receiver a narrow boxcar signal will be observed (Figure 41b). Behind the fault a broader signal will be observed (Figure 41c). Define the rupture velocity (V_r) to be one half of the P- wave velocity (V_p), then when $\theta=0^\circ$,

$$\tau_p = \frac{L}{V_p} \quad (14)$$

and when $\theta=180^\circ$

$$\tau_p = 3 \frac{L}{V_p}. \quad (15)$$

For microearthquakes in the Socorro area, fault radii on the order of a few meters are observed [Ake and Sanford, 1984; Sanford et al., 1984]. If $L=10\text{m}$ (certainly an upper bound), then the difference between the pulse widths for $\theta=0^\circ$ and $\theta=180^\circ$ becomes fairly small at relevant frequencies (Figure 42a). If attenuation along the path is considered, the effective pulse that would interact with the magma body is essentially the same for the two cases considered above. The effect of attenuation on the two pulses in Figure 42a was calculated for a propagation distance of 10.5 km and a Q value of 650; the results are plotted in Figure 42b. All observable differences between the two pulses are now absent. Theory would thus predict that it is safe to ignore any azimuthal effects on the spectra of microearthquakes of the size considered by this study.

The observational results are certainly consistent with the theoretical arguments outlined above. Consider the spectra of the S and S_zS waveforms in Figures 25 and 43. The striking similarity observed between direct and reflected phases strongly suggests that no

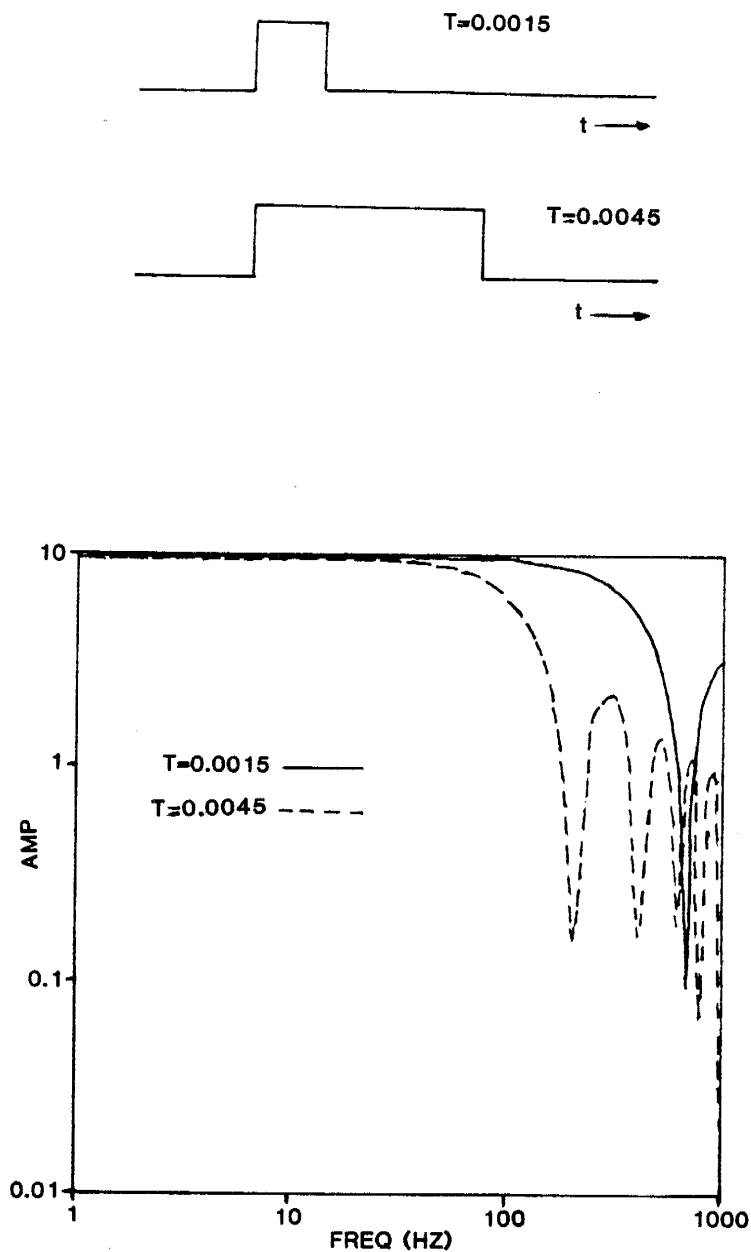


Figure 42a. Frequency domain behaviour of representative source pulses (bottom). Time domain pulse width varies by a factor of three (top). Infinite Q value assumed.

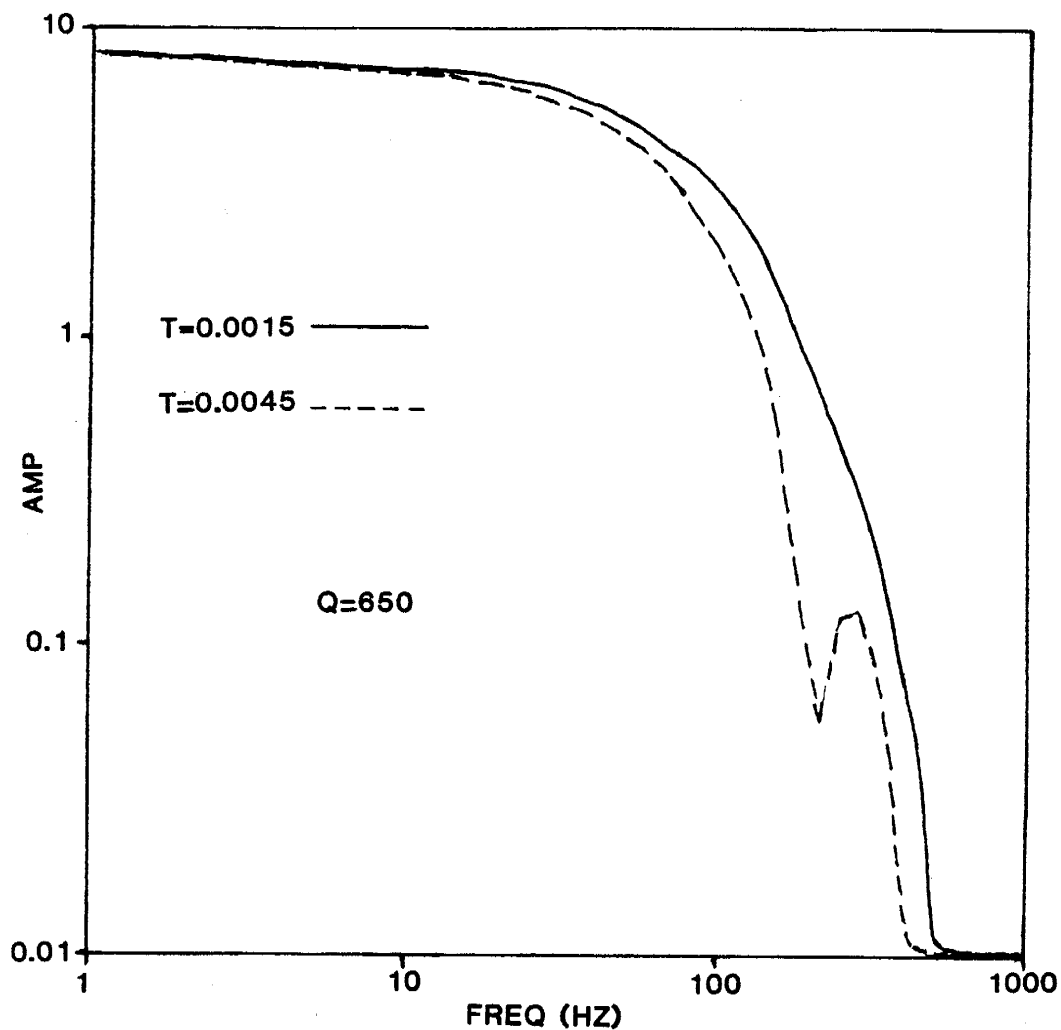


Figure 42b. Frequency domain behaviour of representative source pulses in Figure 42a if attenuation is considered. A propagation distance of 10.5 km and a Q of 650 have been assumed.

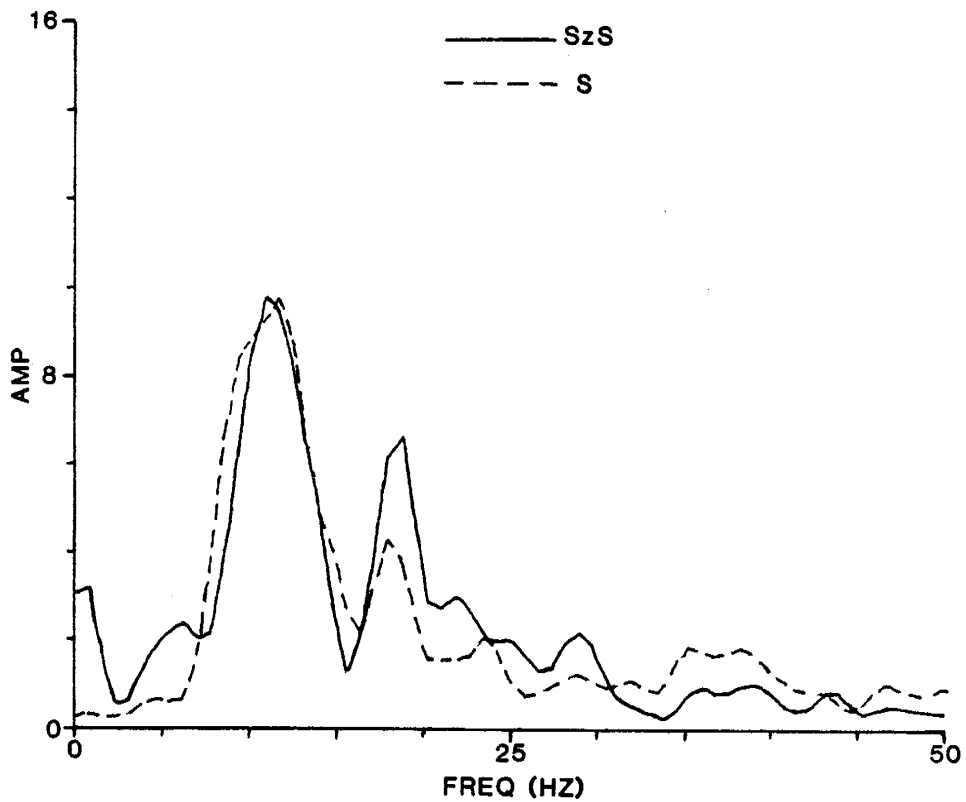


Figure 43. Comparison of S and S_zS phase spectra, 065349 event. Spectra are for ground ^zvelocity. A three point moving average filter has been applied.

significant spectral shifts exist between rays travelling upwards and downwards from the source.

MODELLING

Outline of Method

The recorded seismic trace can be expressed as the convolution of a series of linear operators representing source effects, path and instrument response and attenuation. In the frequency domain this may be expressed as

$$A(f) = S(f) \cdot H \cdot G(f) \cdot I(f) \cdot W(f) \quad (16)$$

where $A(f)$ is the spectra of the recorded time series, $S(f)$ is the source function, H represents geometrical spreading, $G(f)$ is the path response function, $I(f)$ represents instrumental response and $W(f)$ is an attenuation operator. Before realistic synthetic seismograms can be calculated and desired properties isolated, the unknown factors on the right side of equation (16) must be specified or removed.

Geometrical spreading is frequency independent and so does not affect spectral shape, only spectral level. Since this study will only try to reproduce spectral shape, and not absolute amplitudes, geometrical spreading, H , can be ignored.

Differential source effects can be the cause of considerable confusion when comparing waveforms. The similarity of the direct S and P phases for the events studied here (Figures 21 and 23) would imply no differential source effects are present. Previous studies in the Socorro area have shown that microearthquakes are often source independent below threshold magnitudes on the order of 1 [Sanford et al., 1984; Ake and Sanford, 1984]. For these weak events, the corner frequency of the source lies above the high frequency cutoff imposed by

attenuation along the travel path (Figure 44). This phenomena is discussed in some detail in Ake [1984], and has been shown to extend to at least $M_L=0.4$ for the swarm studied here [Sanford et al., 1984]. This is illustrated in Figure 45 with a panel of duplicate S phase waveforms over a wide range in event strengths, the strongest in the figure having a magnitude of 0.3. Using conventional source theory it is impossible to explain this behaviour without requiring unreasonable physical properties. Specifically, to preserve the observed pulse widths, the source radius must not become too large. This requires the resulting stress drop to become unreasonably large for the larger magnitude events. To insure source independence, all events considered in this study have local magnitudes less than zero (i. e., below the threshold of duplication).

To actually construct the seismograms, some known delta-like source pulse must be assumed. The solution for a pressure pulse in a sphere [Sharpe, 1942] was utilized. Following O'Neill and Healy [1973], the far field displacement-time function for the source may be written as

$$u(t) = \frac{4\sigma}{\sqrt{2}\mu} \frac{a^2}{r} \exp\left(\frac{-Wt}{\sqrt{2}}\right) \sin(Wt) \quad (17)$$

where

$$\begin{aligned} u &= \text{radial displacement,} & r &= \text{radial distance,} \\ a &= \text{cavity radius} & \mu &= \text{rigidity,} \\ T &= \text{time,} & \sigma &= \text{step function pressure pulse,} \\ W &= \frac{2\sqrt{2}V_p}{3a}, & V_p &= \text{P wave velocity,} \\ \text{and } t &= \text{reduced time} = T - \frac{(r-a)}{V_p}. \end{aligned}$$

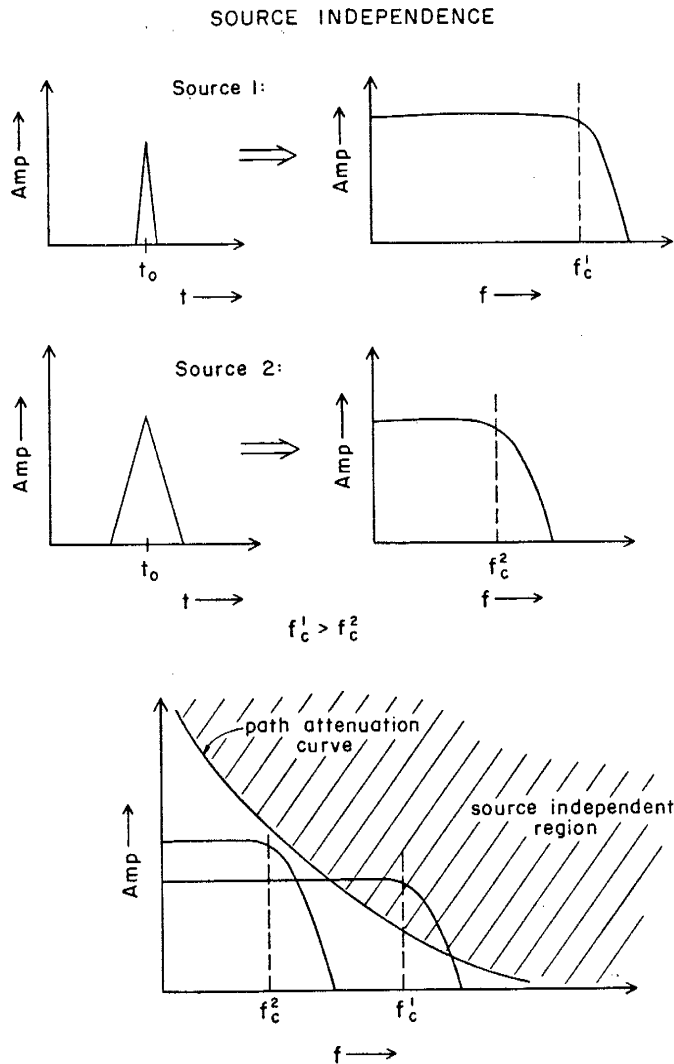


Figure 44. Illustration of source independence in the frequency domain. The location of the corner frequency, f_c , depends on the "deltaness" of the source function (top). Source independence depends on the position of the corner frequency relative to the path attenuation curve (bottom).

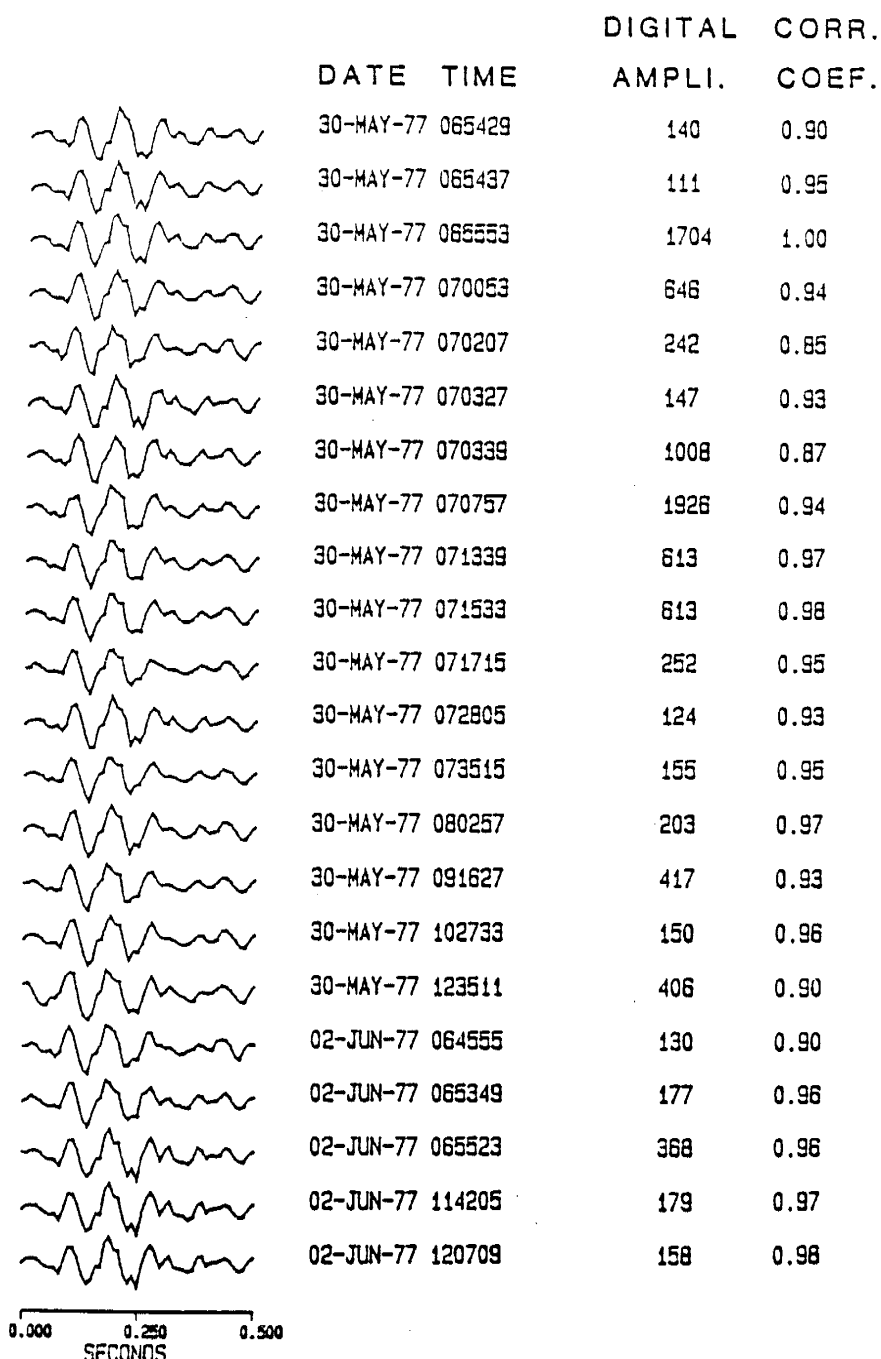


Figure 45. Comparison of S phase waveforms from the May 30-June 4, 1977 microearthquake swarm. Magnitudes range from -0.9 (maximum digital amplitude 111) to +0.3 (maximum digital amplitude 1926). Average correlation coefficient for these waveforms is 0.90.

Higher order, near field terms have been omitted. Previous work in the Socorro area based upon pulse width studies [Sanford et al., 1984; Ake, 1984] has indicated source radii of a few meters to a few tens of meters to be appropriate for small ($M_L < 1$) microearthquakes. To evaluate equation (17) the following quantities were defined:

$$\begin{aligned} V_p &= 5850 \text{ m/sec [Ward et al., 1981],} \\ a &= 10 \text{ m [Ake et al., 1984; Sanford et al., 1984],} \\ r &= 10 \cdot a, \\ \frac{\sigma}{\mu} &= 1 \text{ (enters only as a scale factor).} \end{aligned}$$

The resulting delta-like displacement pulse is plotted in Figure 46a. Figure 46b shows the spectra of the source pulse at ranges of $r=100$ m and $r=10,500$ m with a Q value of 700. Notice that although the true corner frequency of the source pulse is no longer evident for $r=10,500$ m, significant high frequency ($f > 30$ Hz) energy is still present. Hence considerable resolution would still be expected at this distance.

Attenuation effects were modelled by assuming a function of the form

$$W(f) = W_o \exp\left(\frac{-\pi r f}{Q V_p}\right) \quad (18)$$

where r is the propagation distance, Q is the seismic quality factor and V_p is the average P wave velocity along the travel path. Given a known propagation distance and average Q and velocity values along the path, the attenuation function can be determined. Following the results of Carpenter [1984], Q is assumed to be independent of frequency. The Q_p value used in equation (18) is for the depth range

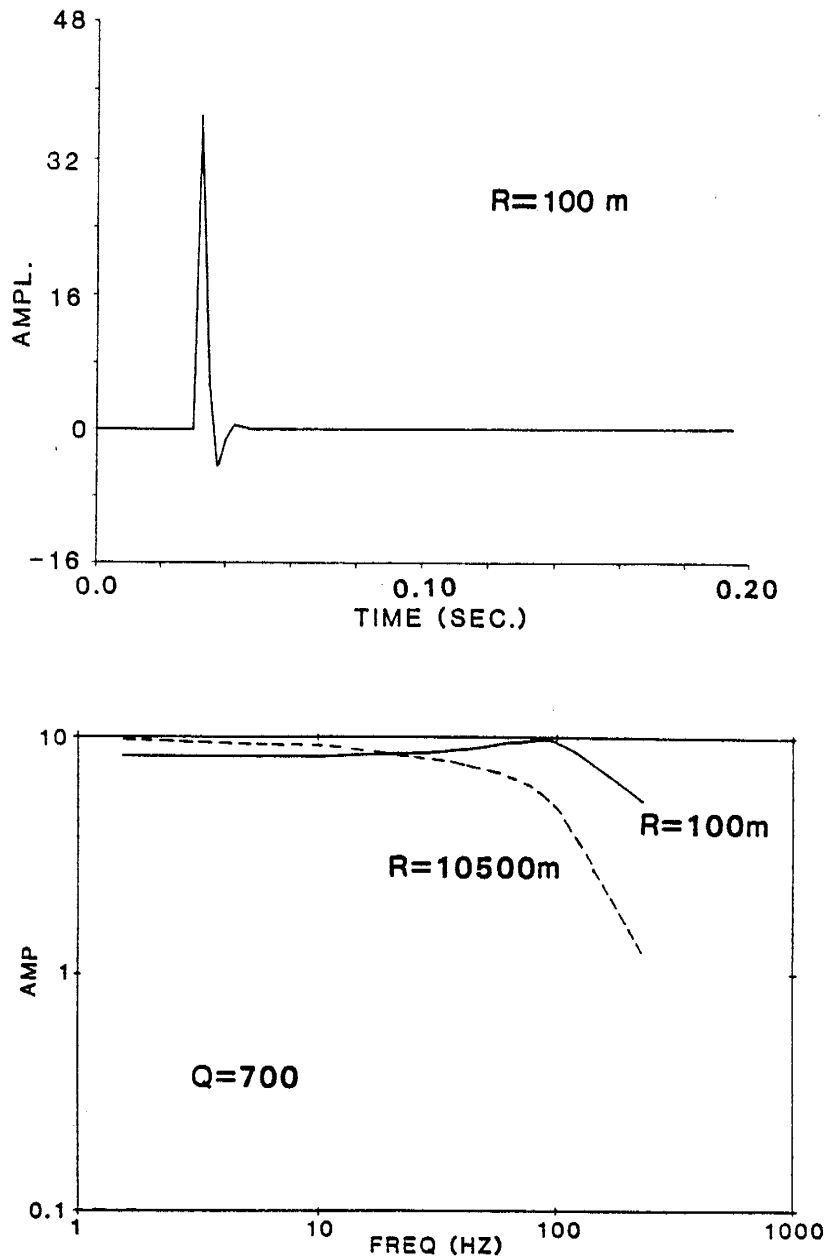


Figure 46. a) Time domain source pulse. Pulse shape is from equations derived by Sharpe (1941). Source radius is 10m and range is 100m.
 b) Spectra of source pulse at ranges of 100m and 10,500m.

10-20 km. Parametric studies indicate that a Q_p value of ≈ 700 was needed to duplicate the pulse width of the initial portion of the P_zP waveform (Figure 47). This value is consistent with the results of Carpenter and Sanford [1985] who found a Q value of ≈ 400 for the upper 10 km of the crust in the Socorro area. Carpenter [pers. comm.] determined a Q value of roughly 600 for the middle crust using spectra of earthquakes and explosions. Brocher [1981] obtained a Q value of 625 for the mid-crust north of Socorro from analysis of the COCORP reflection data.

When studying small earthquakes it is often difficult to isolate the pronounced effect of low-velocity, low- Q , near-surface materials on the seismic signal from other factors [Frankel, 1982; Carpenter and Sanford, 1985; Cranswick et al., 1985]. In order to account for the near surface path effects on the reflected phase, an empirical Green's function was found to model these effects. Notice in Figure 18 that the path of the direct and reflected P phases is almost identical through the near-surface materials. Since the events considered here are quite small, and the source function is quite impulsive (delta-like), the initial portion of the direct P phase may be considered as the path impulse response function for the upper crust and near-surface materials. To improve the signal to noise ratio, minimize digitization effects and "average-out" short wavelength near-station effects, the short P windows in Figure 23 were stacked and a single path and instrument response function determined (Figure 48). Effectively, this breaks the path response function ($G(f)$) into two parts, one for upper crustal and instrumental effects ($R(f)$) and one for the mid-crustal and magma body effects ($L(f)$). It is now possible to rewrite equation (16)

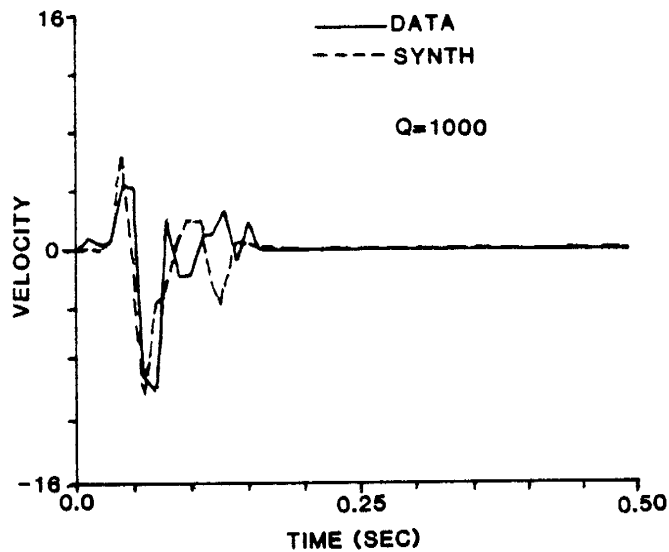
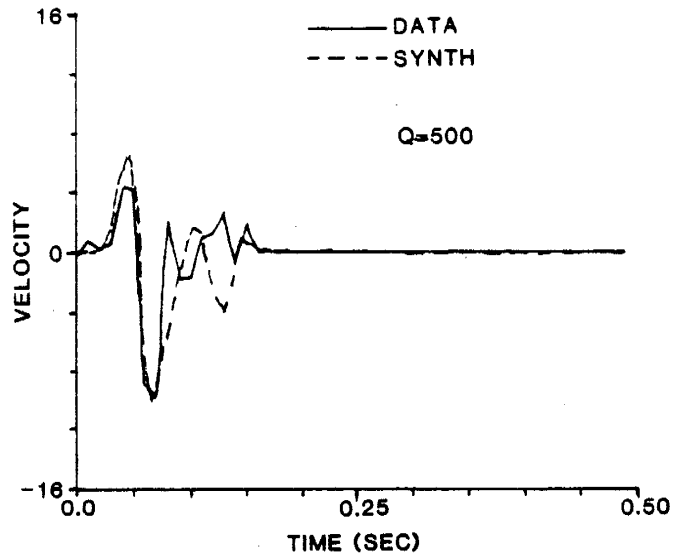


Figure 47. Illustration of attenuation on P_zP waveform. The figure compares synthetic and observed waveforms to illustrate the effect of mid-crustal attenuation upon pulse widths. Observed waveform is 114205 event. Single layer model (to be discussed later) is used to generate synthetics. Q value of medium is 500 (top) and 1000 (bottom).

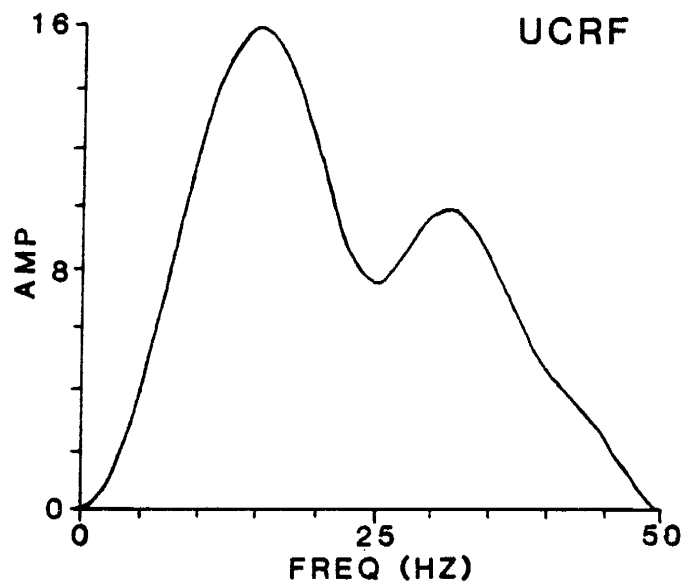
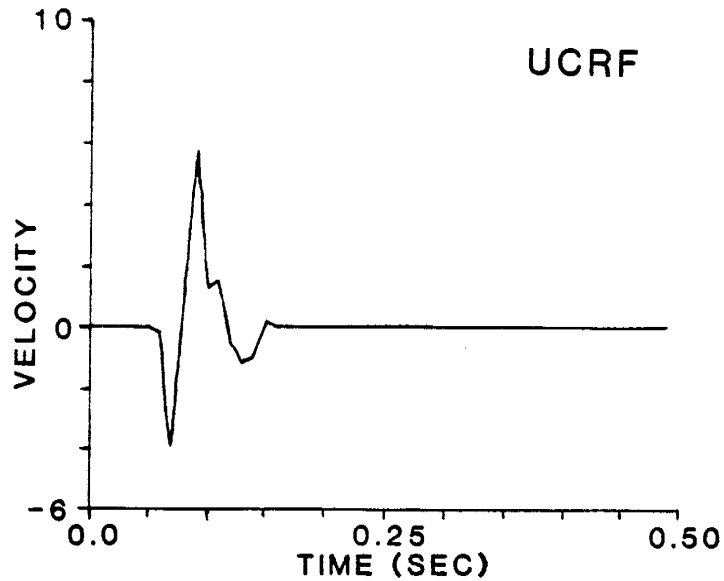


Figure 48. a) Stacked direct P phase from dashed windows in Figure 23. The spectrum of this waveform was used as the upper crustal response function, $R(f)$, in equation (19).
 b) Spectrum of upper crustal response function (UCRF). Note that 26 hz is not a peak in the UCRF spectra.

as

$$A(f) = S(f) \cdot W(f) \cdot L(f) \cdot R(f) \quad (19)$$

where $R(f)$ is the spectra of the stacked short P-window in Figure 48, $S(f)$ is the spectrum of the source pulse defined by equation (17), and $W(f)$ is defined by equation (18). For the sake of simplicity normal incidence is assumed.

All quantities on the right side of equation (19) are now defined except for $L(f)$, the magma body effects. Once a model for the magma body is chosen, the response ($L(f)$) can be defined and synthetic spectra or time series calculated.

Calculation of Synthetic Seismograms

This section presents a brief description of the method used to construct the synthetic seismograms in this study. For normal incidence upon plane-layered structures at large distances from the source, synthetic seismograms can be calculated by simply convolving a known source wavelet with a reflection coefficient series. This amounts to time shifting, rescaling and superimposing a sequence of input source functions (Figure 49). This neglects attenuation however. The present study has incorporated attenuation effects by calculating a series of attenuation modified source wavelets which are then time-shifted, multiplied by the appropriate reflection coefficient and superimposed. This will then represent lower crustal and magma body effects, $L(f)$, and attenuation effects, $W(f)$, in equation (19).

Attenuation within the earth may be thought of as a filtering process. In the frequency domain, the modulus of the filter may be written as:

$$|W(f)| = A_o \exp\left(\frac{-\pi r f}{QV_p}\right) \quad (20)$$

where Q is assumed to be frequency independent. To completely define the filter function, the phase spectrum must be specified. Earth materials are generally assumed to behave in a minimum phase fashion [Ricker, 1953; Anstey, 1977]. A minimum phase filter is one that produces the minimum possible phase shift for its gain [Kanasewich, 1981]. For a physically realizable, minimum-phase filter the logarithm of the modulus and the phase spectra are Hilbert transform pairs [Kanasewich, 1981]:

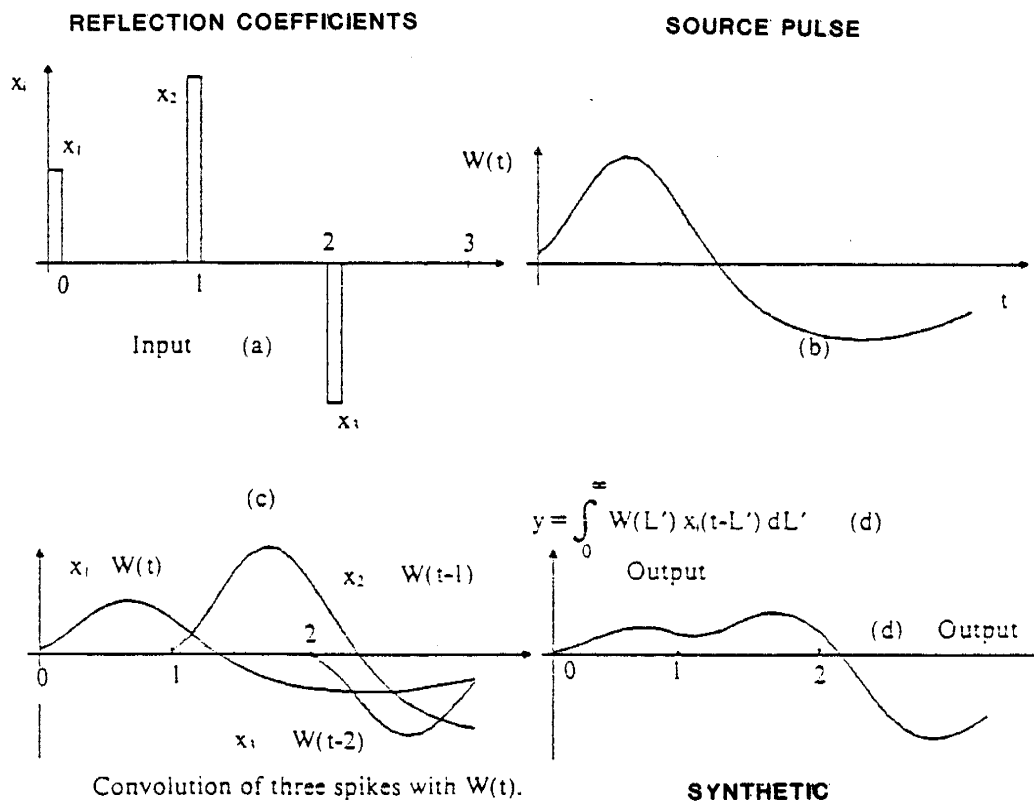


Figure 49. Illustration of source pulse convolution. The source is convolved with a reflection coefficient series to produce a simple synthetic waveform (after Kanasewich (1981)).

$$\log |Y(f)| = \frac{-1}{\pi} \int_{-\infty}^{\infty} \frac{\phi(f)}{Y-f} dw \quad (21)$$

and

$$\phi(f) = \frac{1}{\pi} \int_{-\infty}^{\infty} \frac{\log |Y(f)|}{Y-f} dw \quad (22)$$

Substitution of equation (20) into equation (22) allows the calculation of the phase spectra. Specification of the effective values of Q and V along a path of length r in equation (20) allows the computation of $|W(f)|$ and hence $\phi(f)$. This filter is then applied to the source wavelet (equation (17)), the result is a causal, minimum phase wavelet. Figure 50 illustrates the effect on the source wavelet (defined by equation (17)); of transmission through 10.5 km of material with a Q value of 700. This filtering process is applied for each significant arrival, the resulting attenuation modified wavelets are then time-shifted, rescaled and superimposed. The resulting waveform ("primitive synthetic") is then convolved with the upper-crustal response function to create the final synthetic seismogram.

As previously noted, the general shape of the P_zP microearthquake spectra is suggestive of constructive interference, hence the effect of interbed multiples were included in the synthetic seismogram calculations. Interbed multiples may be a significant factor in this case, as a low-velocity magma layer sandwiched between high velocity materials may act as an effective energy trap. Both first and second order multiples were included in the magma body modelling. Examples of the number of multiples included, ray paths and reflection coefficients are illustrated in Figure 51.

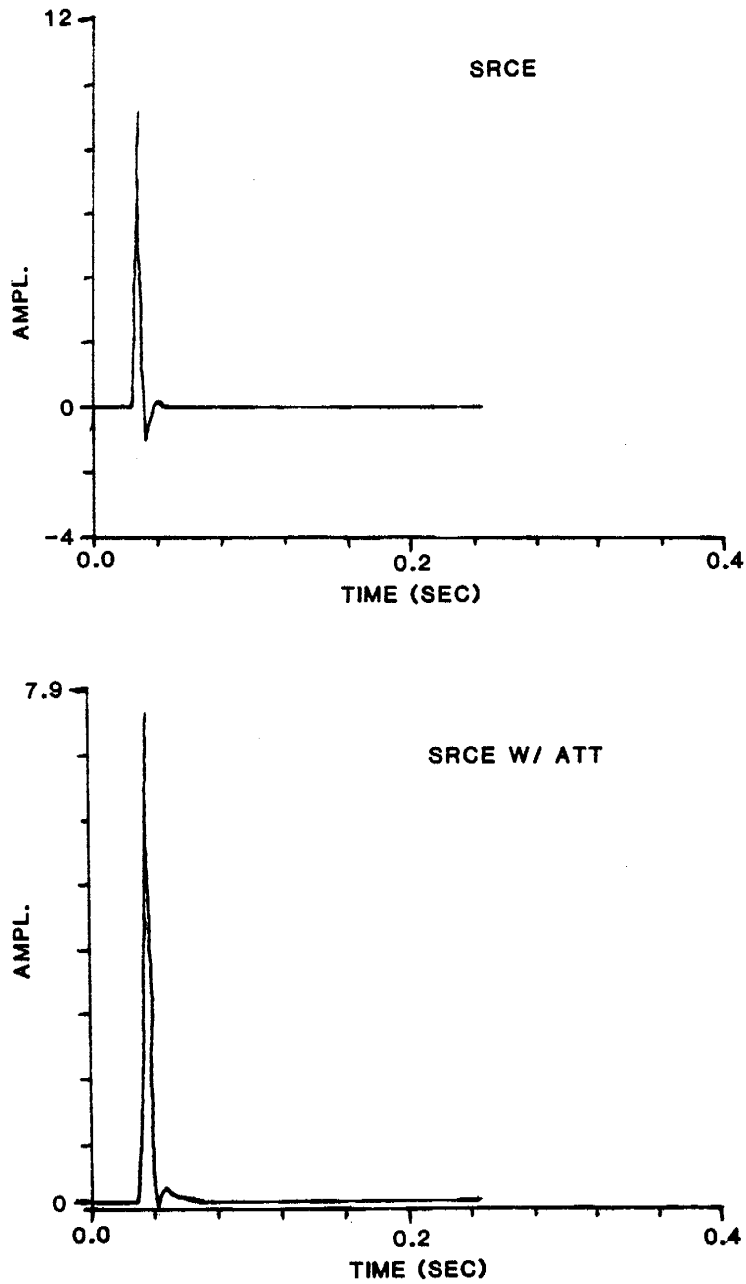
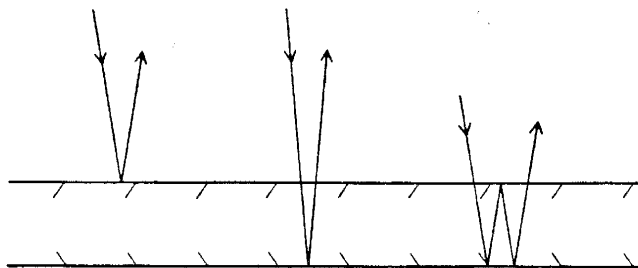
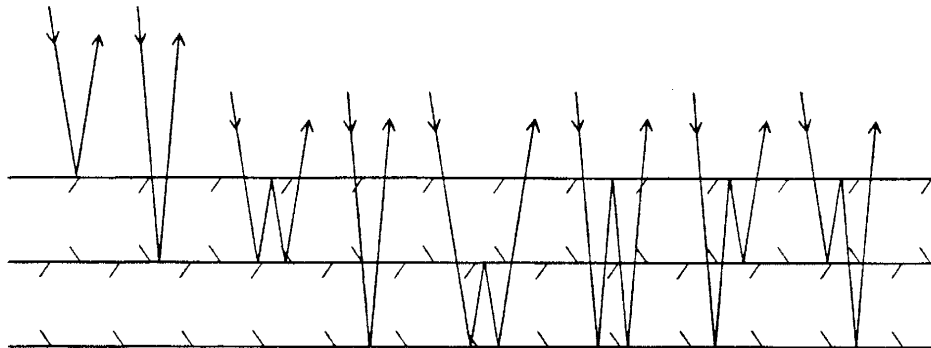


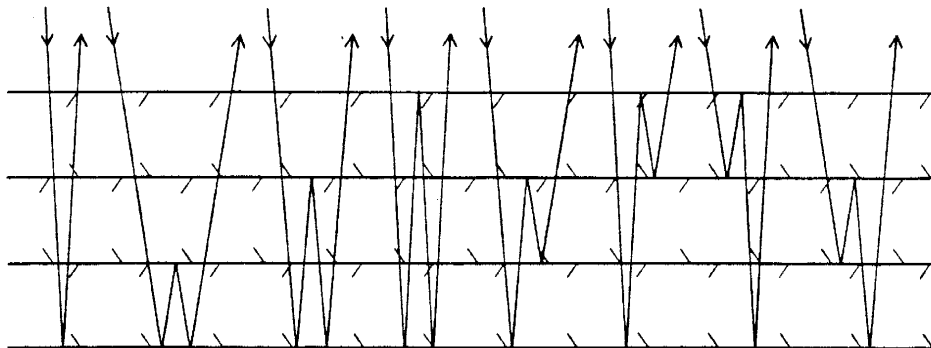
Figure 50. Comparison of source wavelet before (top) and after (below) transmission through 10.5 km of material with a Q value of 700.



single layer case: 3 arrivals.



two layer case: 8 arrivals.



three-layer case: all of the arrivals for the two layer case plus those shown above, total of 16 arrivals.

Figure 51. Illustration of number of arrivals considered for one, two and three layer case. Only "second order" multiples (i.e. three or fewer reflection points) or less were considered.

Resolution

The question of vertical and horizontal resolution will invariably arise in a study where relatively thin beds are being investigated at large distances from the source. This section will discuss the implications of resolution on the present study.

Lateral resolution must be considered because of the geometrical divergence that occurs as waves travel outward from the source region. Consider normal incidence upon a flat reflecting surface. Rather than use conventional ray theory (Figure 52a), describe the situation with geometrically spreading wavefronts (Figure 52b). By Huygen's principle, each point along the boundary will act as a source of wavelets travelling outward from that point. Thus, signals being reflected from point B will be delayed by some amount relative to signals being reflected from point A. A "spatial delay" is introduced, constructive interference will occur if the spatial delay is $< \frac{\lambda}{2}$. This length along the reflector is the first Fresnel zone. The diameter of the Fresnel zone is given by

$$d \approx 2\sqrt{\frac{VZ}{2f}} \quad (21)$$

where f is the frequency of the signal, V is the velocity above the reflector and Z is the distance from source to reflector (see Figure 53).

How large an area of the reflector contributes to the observed microearthquake reflected signals? The answer not only depends upon the frequency under discussion, but will differ for P and S waves. To evaluate equation (21) recall that most of the energy that actually

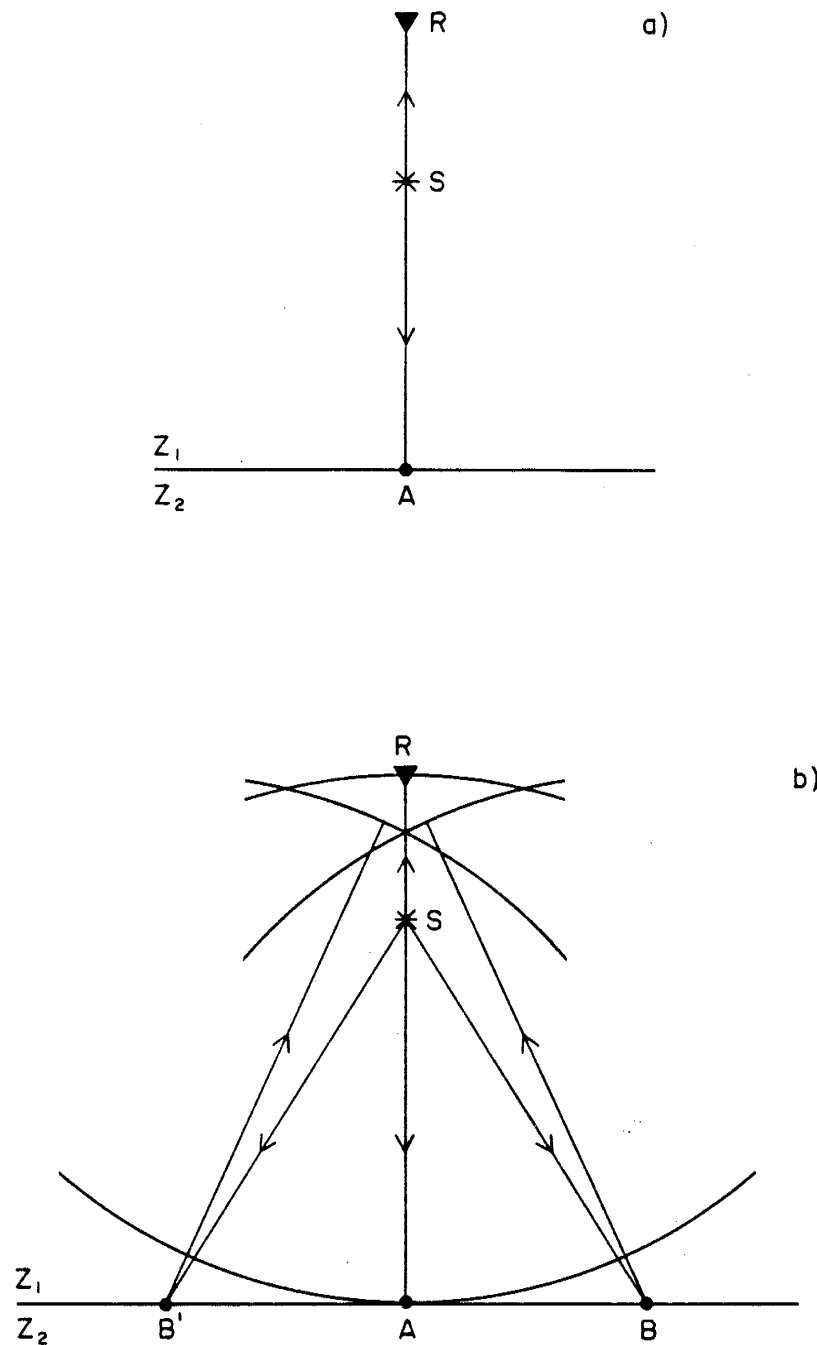
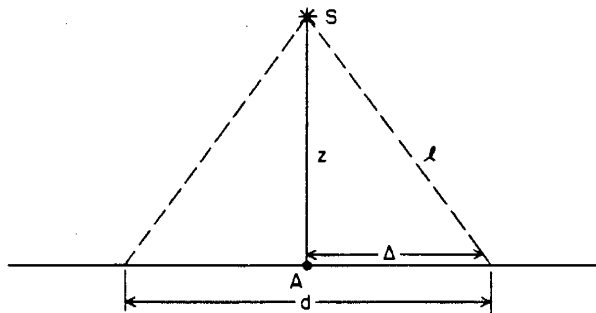
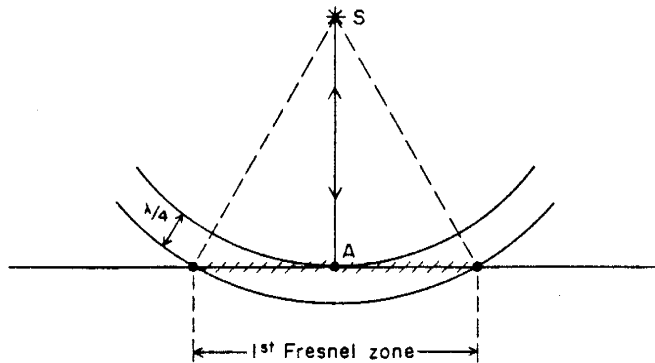


Figure 52. a) Geometrical relationship for reflection from point A on reflecting surface when source is at S and receiver at R using ray theory.

b) Geometrical relationship between arrival from reflection points at normal incidence (A) and non-normal incidence (B' and B) using geometrically spreading wavefronts. Source is at S , receiver at R and Z is acoustic impedance.



$$d = 2\Delta \text{ and } l = z + \frac{\lambda}{4}, \text{ hence } \Delta = \sqrt{l^2 - z^2}$$

$$\text{also } v = f\lambda \text{ and } z \gg \lambda \quad = \sqrt{z^2 + 2z \frac{\lambda}{4} + \frac{\lambda^2}{16} - z^2}$$

so,

$$= \sqrt{\frac{z\lambda}{2} + \frac{\lambda^2}{16}}$$

$$d \approx 2 \sqrt{\frac{vz}{2f}}$$

Figure 53. Relationships used in derivation of Fresnel zone diameter. Source is at S and zero offset reflection point is at A. A monochromatic source is assumed. The wavelength of emitted signals is λ in the upper media.

reaches the surface to be sensed by the seismometer lies in the spectral band of 10-40 hz (see Figure 43). As a result, the calculation of Fresnel zone diameter will be carried out with $f=10$ and 40 hz to produce an estimate of the minimum and maximum effective diameter. To calculate the S wave Fresnel zone diameter assume a Poisson's ratio of $1/4$, a P wave velocity of 5.85 km/sec and a distance of 10.5 km (see Figure 53).

The consistency in S to S_zS amplitude ratio, and the constant S_zS and P_zP arrival times for all swarm events, implies that the upper surface of the magma body is flat over a region at least as large as the S wave Fresnel zone (2.7 km) plus the uncertainty in reflection point locations (≈ 500 m), i.e. about 3.2 km. This suggests a minimum size for the magma body in this area. The data set considered here cannot prove or disprove the postulated discrete, pod-like geometry proposed by Brocher [1981], however, it suggests the minimum size of the pod(s) is at least 3.2 km in diameter.

While the S_zS observational data is consistent with the Fresnel zone theory, a discrepancy is evident between observation and theory for the case of the P_zP phase. Carrying out a similar calculation as above for the P wave, the resulting Fresnel zone diameter ranges from 1.75 to 3.5 km. The flat upper surface of the body requires that the observed changes in the P_zP waveforms are due to changes in thickness and structure of the body. Recall that the uncertainty in reflection points is only ≈ 500 m. Since the Fresnel zone is considerably larger than the variation in reflection point position (Figure 54), all of the waveforms will be "averaging" essentially the same portion of the reflector. In other words very little change in waveform

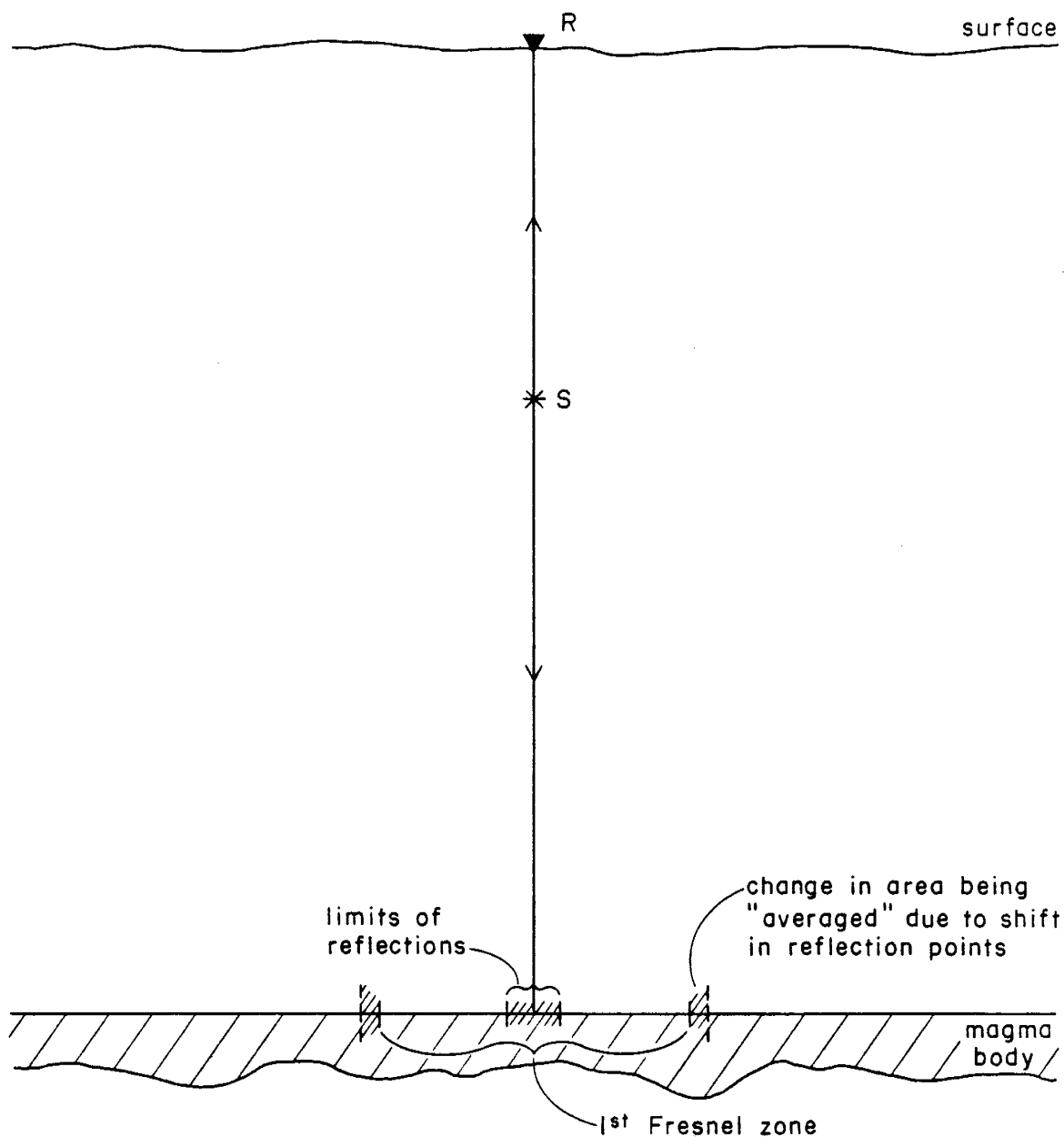


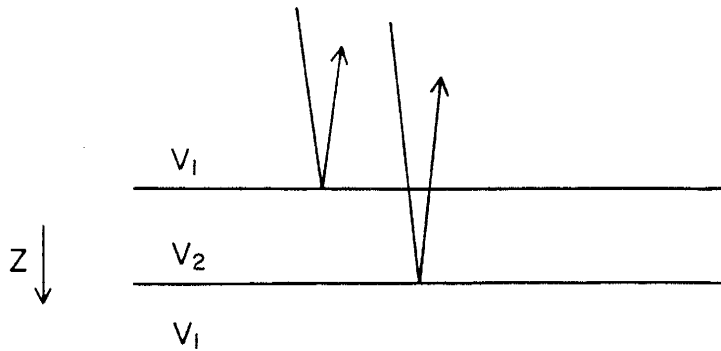
Figure 54. Comparison of lateral dimensions of Fresnel zone and shift in reflection points. According to Fresnel zone theory, all the reflections are "seeing" the same portion of the reflector.

character should be observed unless significant changes in internal structure are occurring. This discrepancy has also been noted, but not resolved, in the exploration geophysics field (J. Knapp, pers. comm.). In this study the observations are felt to be compelling evidence that only a small portion of the reflecting surface is actually making a significant contribution to the observed waveform. However, it is necessary to remember that some spatial averaging is occurring.

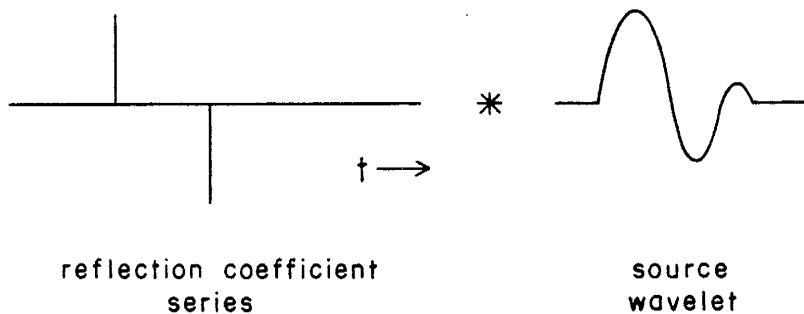
Vertical resolution is the ability to distinguish that more than one reflecting interface is contributing to an observed event on the seismogram. In other words, at what point is it impossible to resolve the top and bottom of a layer imbedded in two semi-infinite media? This question has been addressed in some detail in the exploration geophysics literature. In this study the method of Widess [1973] was used to estimate the limits of vertical resolution.

The reflection seismogram will show interference between the top and bottom of the bed (Figure 55). Widess showed that the character of separate top and bottom reflections becomes indistinguishable for beds with thicknesses less than $\lambda/8$. Beds thinner than this will not be well resolved. Again different wavelengths (or frequencies) have different levels of resolution. Utilizing a P wave velocity of 2.6 km/sec [Murase and McBirney, 1973] and frequencies of 10 and 40 hz, the limits of resolvable bed thicknesses become 8-32 m. Hence the postulated magma body thickness of 75 m should be easily resolved by the microearthquake signals. This points out one advantage of the microearthquake source. The abundance of high frequencies present in the wavetrain when it interacts with the discontinuity (Figure 46) allows for the resolution of very thin beds, especially if fairly high Q

Assume zero offset and ignore multiples.



This geological situation leads to mathematical description:



with the result:

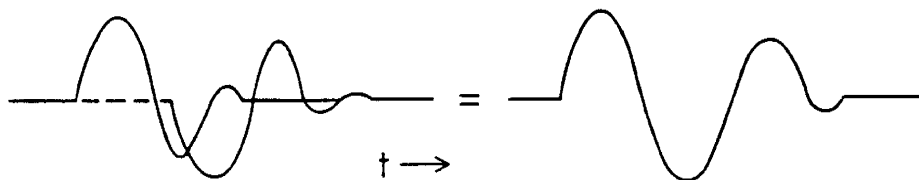


Figure 55. Schematic representation of interference phenomena that influence vertical resolution. Interference between waves reflected from the top and bottom of a bed (top) is equivalent to convolution of reflection coefficients with a source pulse (middle). For this example constructive interference occurs. If the reflection coefficients become too closely spaced (i.e. bed becomes too thin), destructive interference occurs.

surface materials are present.

RESULTS

This section presents a summary of the results of the modelling study. Due to the large number of models considered, only those results relevant to the points discussed previously, or to those discussed in the next section are presented.

The first model considered was a single layer of full melt basalt. The values used for velocity and density of the magma layer are for a full melt Columbia River basalt (CRB) [Murase and McBirney, 1973]. The values for velocities above and below the magma layer are appropriate for depths above and below the Conrad discontinuity in the Socorro area [Toppozada and Sanford, 1976; Rinehart and Sanford, 1981]. Density values outside the magma layer are from Daggett et al. [1986] and Cordell [1976].

Synthetic seismograms for the single layer model were calculated for a variety of layer thicknesses. Comparisons of the synthetic and observed waveforms indicated that the initial estimates of layer thickness based on spectral peaks (equation 2) were reasonable. Synthetic P_zP seismograms and spectra for layer thicknesses of 50, 75 and 100 m are compared to the observed P_zP waveform and spectra of the 114205 02-June event in Figure 56. It is clear from Figure 56 that the synthetic P_zP seismograms and spectra are quite sensitive to small changes in layer thickness.

Overall the fit for $z=75$ m in Figure 56 is reasonable, as indicated by the correlation coefficient of 0.75. In general the correlation coefficient of high frequency waveforms do not tend to be very large. Values of roughly 0.7 are sometimes felt to represent duplication

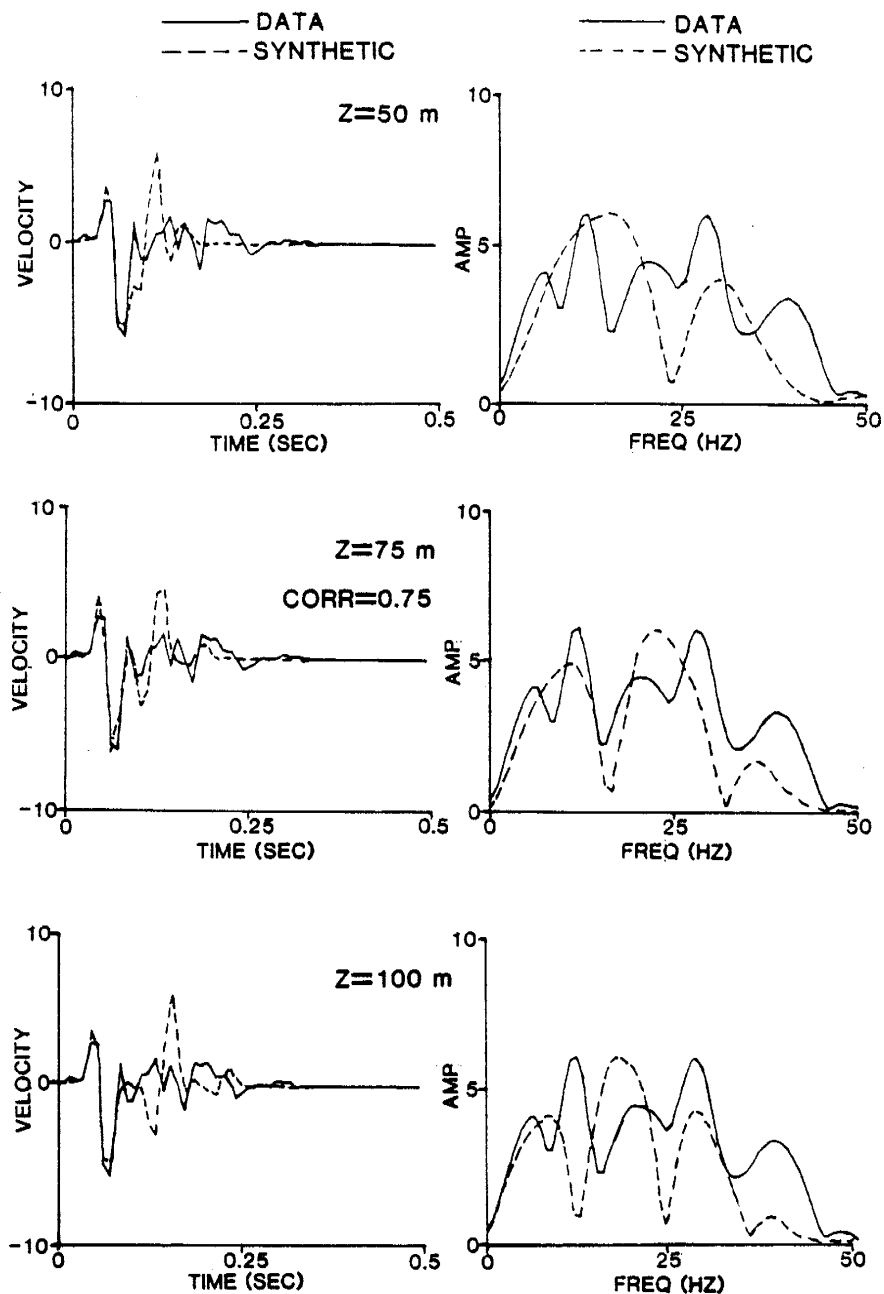


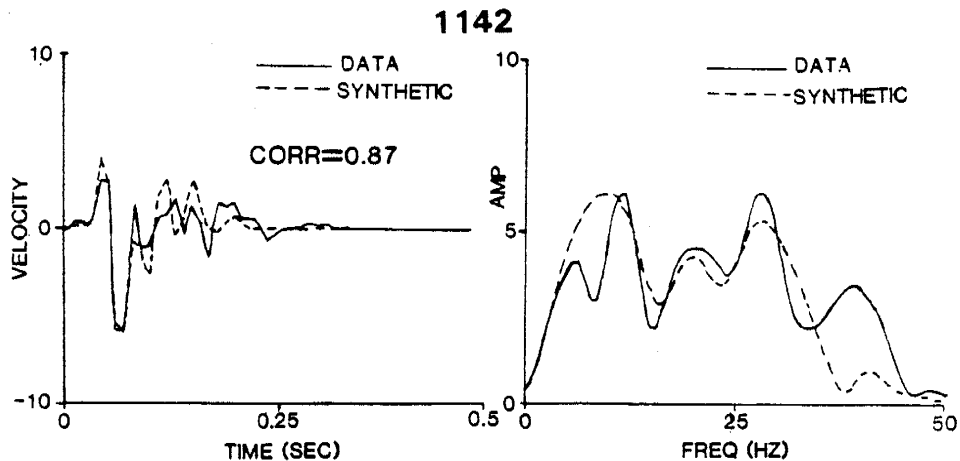
Figure 56. Comparison of observed and synthetic P reflection time series and spectra, single layer model. Observed waveform is from event of 114205 02-June-77. Model used was a single layer of full-melt basalt with a thickness of 50, 75 and 100m. The assumed melt velocity is 2,500 m/sec. Spectra are for ground velocity.

[Frankel and Kanamori, 1979]. The initial portion of the waveform is well duplicated by the single layer synthetic, but the large secondary arrival on the synthetic waveform is not nearly as pronounced on the observed waveform. This mismatch was interpreted as indicative of overestimation of the reflection coefficient at the bottom of the magma layer. To improve the fit between the observed and synthetic waveforms, an intermediate velocity layer beneath the non-rigid layer was added to the model. A velocity of 4.5 km/sec was defined for this layer. This velocity is consistent with a partially solidified CRB [Murase and McBirney, 1973].

The addition to the model of a layer of partial melt beneath a layer of full melt has physical justification. If a sill-like intrusion cools in a gravity field a layer of semi-solidified mafic crystal mush will accumulate at the bottom of the sill. The Palisade sill complex is an example of this type of geometry [Hyndman, 1972]. Another plausible model would be the intrusion of young, fully molten magma above a slightly older, partially solidified sill. The amount of mixing is expected to be minimal if the aspect ratio of the chamber is low [Huppert et al., 1986].

The effect of adding the intermediate velocity layer on the model synthetic P_zP seismograms is illustrated in Figure 57. This addition results in a significant improvement in the fit between the observed and synthetic waveforms; an increase in correlation coefficient from 0.75 to 0.87.

All of the observed waveforms could be fit reasonably well with this simple two layer model. The only changes in the model necessary to fit each observed waveform were small adjustments of layer



$$V_1=5850 \text{ m/s}, \rho_1=2700 \text{ kg/m}^3, Q_1=700, Z_1=10500 \text{ m}$$

$$V_2=2500 \text{ m/s}, \rho_2=2600 \text{ kg/m}^3, Q_2=5, Z_2=64 \text{ m}$$

$$V_3=4500 \text{ m/s}, \rho_3=2650 \text{ kg/m}^3, Q_3=300, Z_3=58 \text{ m}$$

$$V_4=6500 \text{ m/s}, \rho_4=2850 \text{ kg/m}^3, Z_4=\infty$$

Z_1 =Difference in depth of top of magma layer, $\approx 20 \text{ km}$,

and average focal depth of swarm, $\approx 9.5 \text{ km}$.

Figure 57. Comparison of observed and synthetic P-reflection time series and spectra, two layer model. Observed waveform is from the event at 114205 on 02-Jun-77. Spectra are for ground velocity. Model values are indicated.

thicknesses. Shown in Figure 58 are the 102733 30-May and 120709 02-June waveforms and spectra with the corresponding best fit models. Although values of velocity, density, and Q are indicated for each layer in Figures 57 and 58, parametric studies indicate that all density values, and Q values for the magma layers are poorly constrained. The full and partial melt layers are thin enough that any non-zero Q will have little effect on the synthetic waveform. Q values of the full melt were varied between 1 and 200 with no significant change evident in the resulting waveforms. The values chosen for the modelling were consistent with values in the literature. Since the density of a full melt basalt [Murase and McBirney, 1973] varies by only a few percent from that inferred for the country rock [Rinehart and Sanford, 1981] density has little or no effect on the reflection coefficient at the interface between the two materials. Hence densities are not resolved by the modelling. In all cases the cross-correlation coefficient between the observed waveform and respective best-fitting synthetic waveform was greater than the correlation coefficient for that waveform with respect to the other observed waveforms (shown in Figure 22).

A high frequency, low amplitude oscillation appears at the beginning of several of the P_zP waveforms in Figure 22. It was hypothesized that these could be due to reflection from a thin, high velocity lid to the magma body. This idea was tested by adding a thin (50-90 m), high velocity ($V_p=6500$ km/sec) layer to the two layer model. The results are shown for the 065455 02-June and 072805 30-May events in Figure 59a. The best fitting waveforms for the 065349 02-June and 070327 30-May events are shown in Figure 59b. The

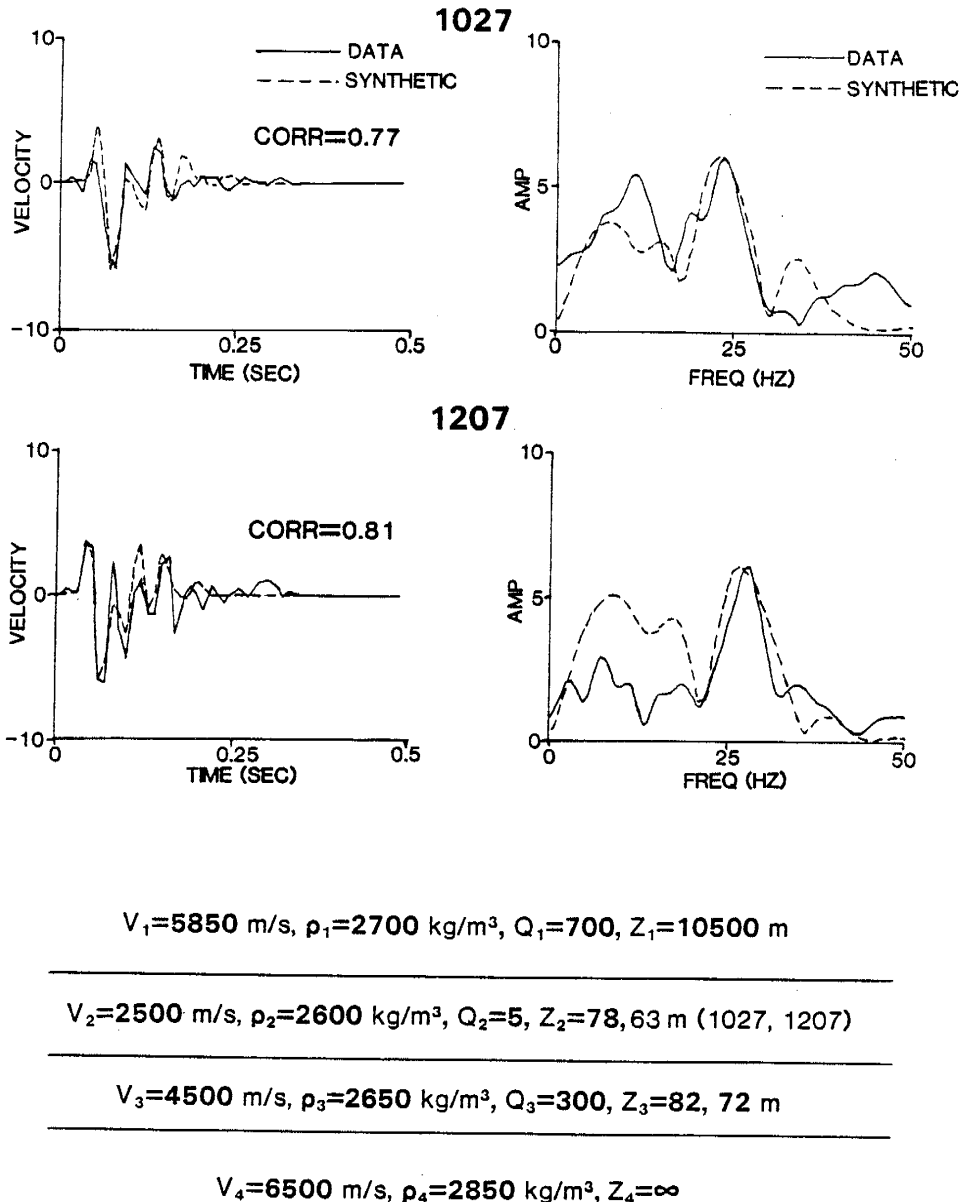
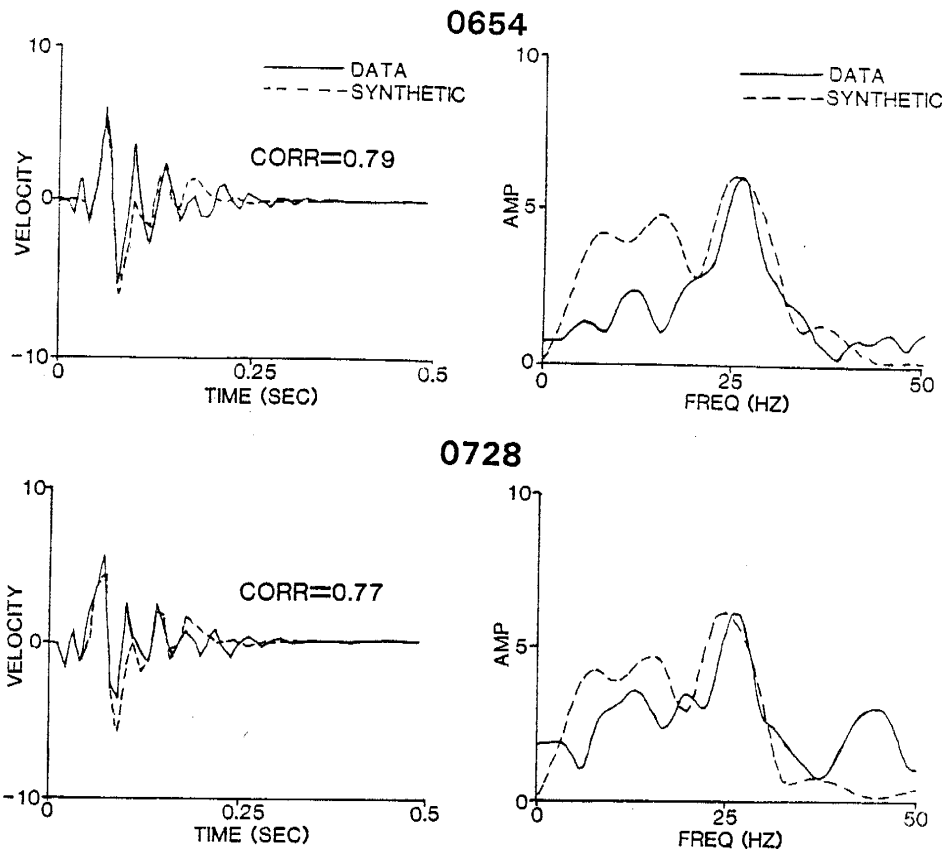


Figure 58. Comparison of observed and synthetic time series, two layer model. Observed P P waveforms are from events at 102733 30-May and 120709 02-Jun 1977 (Figure 22). The two models used to produce synthetic data had the same velocities, densities and Q values, only the thickness of the full and partial melt layers differ. Correlation coefficients between observed and synthetic time series are given. Model values are indicated at the bottom of the figure.



$$V_1=5850 \text{ m/s}, \rho_1=2700 \text{ kg/m}^3, Q_1=700, Z_1=10500 \text{ m}$$

$$V_2=6500 \text{ m/s}, \rho_2=2850 \text{ kg/m}^3, Q_2=800, Z_2=65, 85 \text{ m (0654, 0728)}$$

$$V_3=2500 \text{ m/s}, \rho_3=2600 \text{ kg/m}^3, Q_3=5, Z_3=70, 70 \text{ m}$$

$$V_4=4500 \text{ m/s}, \rho_4=2650 \text{ kg/m}^3, Q_4=300, Z_4=78, 78 \text{ m}$$

$$V_5=6500 \text{ m/s}, \rho_5=2850 \text{ kg/m}^3, Z_5=\infty$$

Figure 59a. Comparison of observed and synthetic time series and spectra, two layer model with high velocity lid. Observed P_p waveforms are from events at 065429 02-June and 072805 30-May 1977 (Figure 22). The two models used to produce synthetic data had the same velocities, densities and Q values, only the layer thicknesses differ (indicated on figure). Correlation coefficients between synthetic and observed time series are shown.

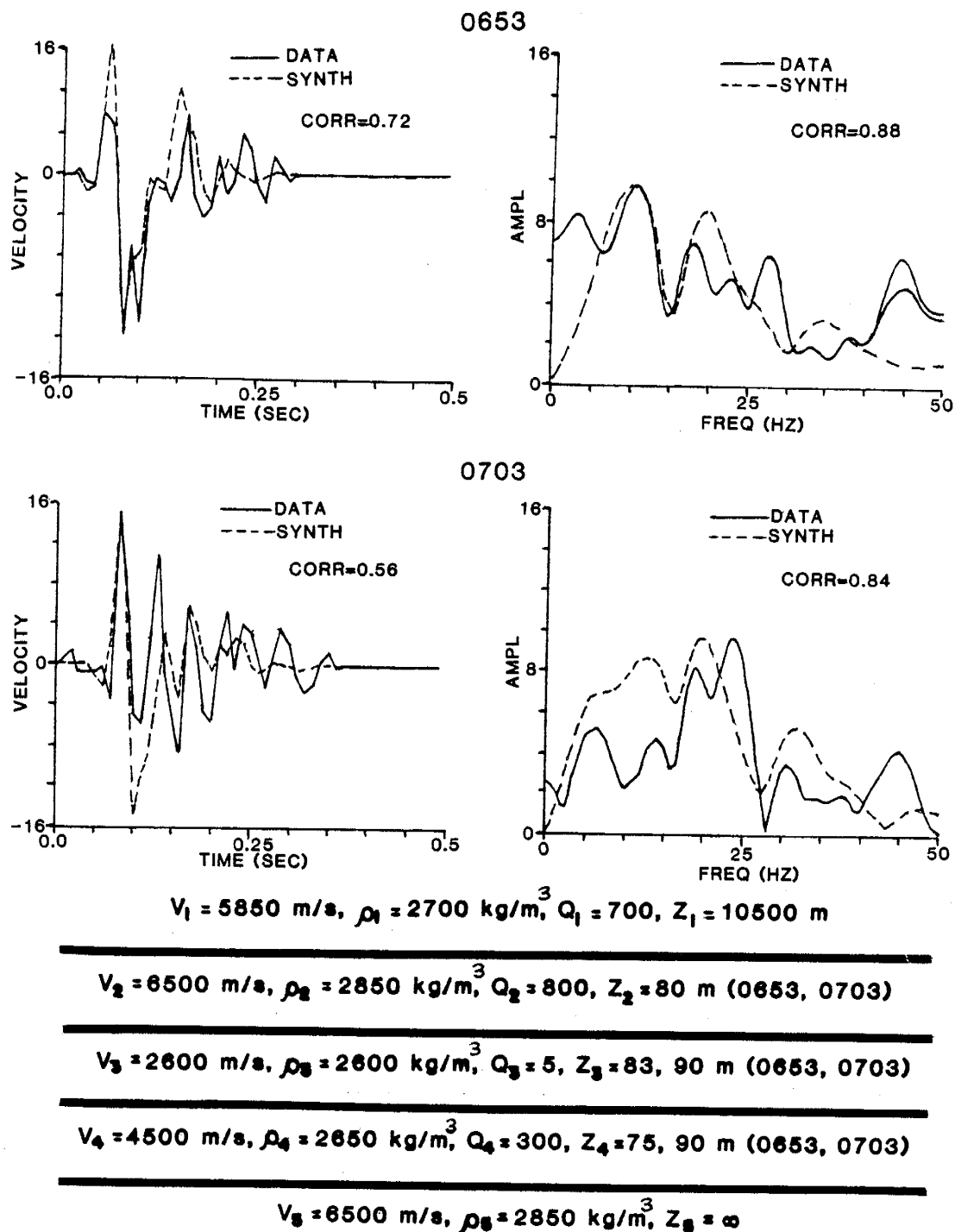


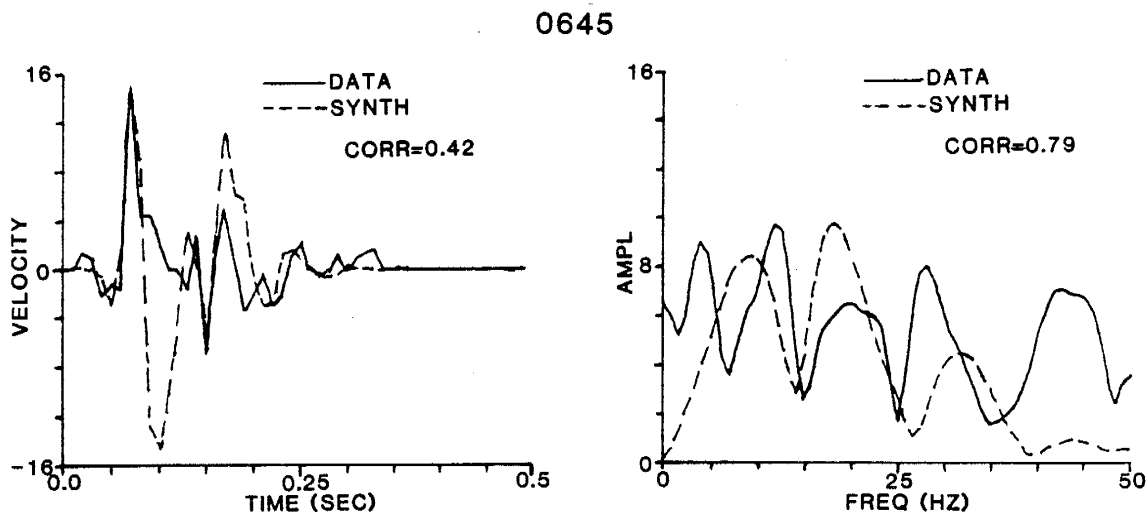
Figure 59b. Comparison of observed and synthetic time series and spectra, two layer model with high velocity lid. Observed P-P waveforms are from events at 065349 02-June and 070327 30-May 1977 (Figure 22). The two models used to produce synthetic data had the same velocities, densities, and Q values, only the layer thickness differ (indicated on figure). Correlation coefficients between synthetic and observed time series are shown.

addition of the high velocity lid resulted in a slight improvement in the quality of fit (as defined by the maximum value of the cross correlation function) relative to the two layer model for these events.

The 064555 02-June event proved to be a bit more difficult to fit. In this case the observed waveform is atypical when compared to the other observed waveforms (see Figure 22). Notice the "one-sided" nature to the waveform. This is surmised to be a near surface interference effect that results in the cancellation of the large negative portion of the waveform. The modelling program is unable to reproduce these effects as it uses a constant upper crustal response function. The best fitting synthetic time series and spectra are compared to the 0645 waveform in Figure 60. Except for the missing cycle of motion, the synthetic reproduces most of the features of the observed waveform.

Since the maximum spread in reflection points for this swarm is about 500 m and the gross spectral characteristics of the P_zP data are similar, a fit to a stacked or "average" spectra was made. Stacking the spectra will tend to average out the small scale lateral changes in magma body structure. This method allows the determination of general magma body characteristics in the South Canyon area. Figure 61 compares the stacked spectra and the best-fitting model to the stacked spectra. The synthetic spectrum reliably reproduces all of the major features of the observed stacked spectrum.

One of the advantages of stacking the observed spectra is that an estimate of the uncertainties in the spectra may now be obtained. The uncertainties contain both the usual uncertainties associated with background noise and sampling as well as uncertainties due to differences in internal structure of the magma body. A general idea of



$$V_1 = 5850 \text{ m/s}, \rho_1 = 2700 \text{ kg/m}^3, Q_1 = 700, Z_1 = 10500 \text{ m}$$

$$V_2 = 6500 \text{ m/s}, \rho_2 = 2850 \text{ kg/m}^3, Q_2 = 800, Z_2 = 80 \text{ m}$$

$$V_3 = 2600 \text{ m/s}, \rho_3 = 2600 \text{ kg/m}^3, Q_3 = 5, Z_3 = 90 \text{ m}$$

$$V_4 = 4500 \text{ m/s}, \rho_4 = 2650 \text{ kg/m}^3, Q_4 = 300, Z_4 = 78 \text{ m}$$

$$V_5 = 6500 \text{ m/s}, \rho_5 = 2850 \text{ kg/m}^3, Z_5 = \infty$$

Figure 60. Comparison of observed and synthetic time series and spectra, two layer model with high velocity lid. Observed P_p waveform is from event at 064555 02-June 1977 (Figure 22). The model used to generate synthetic data indicated. Correlation coefficient between synthetic and observed time series is shown.

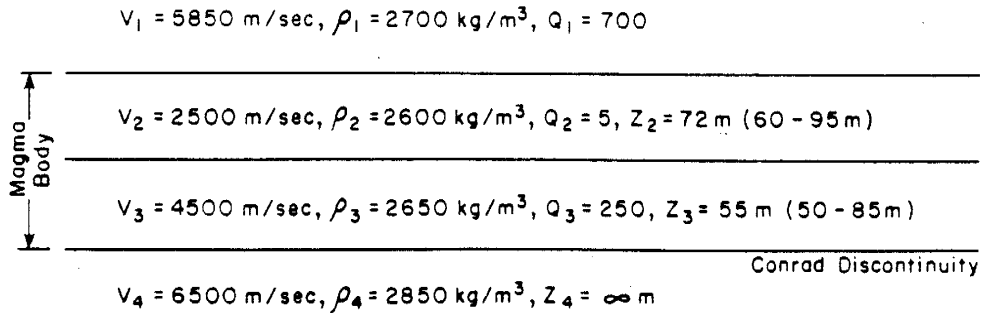
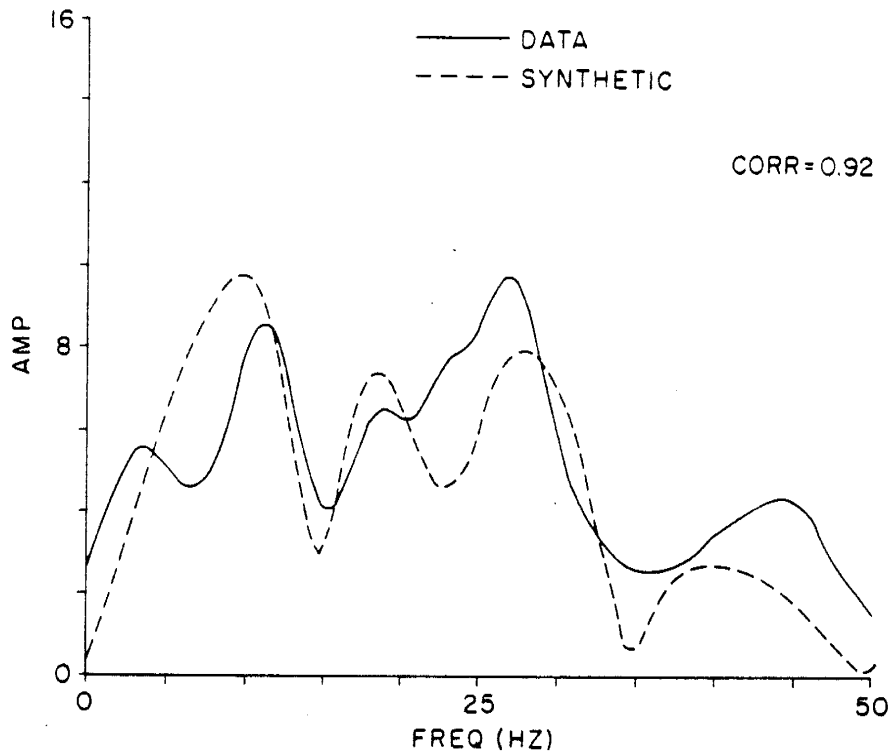


Figure 61. Comparison of stacked P-P spectra with best fitting synthetic spectra. Model used to generate synthetic shown beneath spectra. Spectra have been smoothed with three-point moving average filter.

how well the synthetic models fit the average properties of the magma body is now obtainable. If there are n observed waveforms being stacked, then the mean value of the i^{th} spectral value is given by

$$\bar{x} = \frac{1}{n} \sum_{j=1}^n x_j(i) \quad (22)$$

and the associated standard deviation in the i^{th} spectral value is

$$s.d. = \sqrt{s^2} \quad (23)$$

where

$$s^2 = \frac{1}{(n-1)} \sum_{j=1}^n (\bar{x}(i) - x_j)^2. \quad (24)$$

For the data considered here $n=8$. Figure 62 compares the ± 1 standard deviation envelope to the best fit synthetic spectra shown in Figure 61. The only places where the synthetic deviates noticeably from the observed is at local minimum points in the synthetic spectra. This is to be expected as the observed P_zP waveform occurs within a background of S phase coda which will have a generally "white" character. The synthetics are noise free, hence the local minima in the synthetic spectra will be more pronounced. The characteristics of the best fitting model are summarized in Figure 63.

The preferred model proposed by this study appears to fit the data quite well (Figures 57 and 61). Brocher [1981] studied COCORP magma body reflections north of Socorro and proposed quite a different, multilayered model to explain the spectral character of the reflections. One of the primary questions posed by this study was whether or not the the magma body in this area is a simple single layer of melt or a more complex multilayer intrusion. To try to

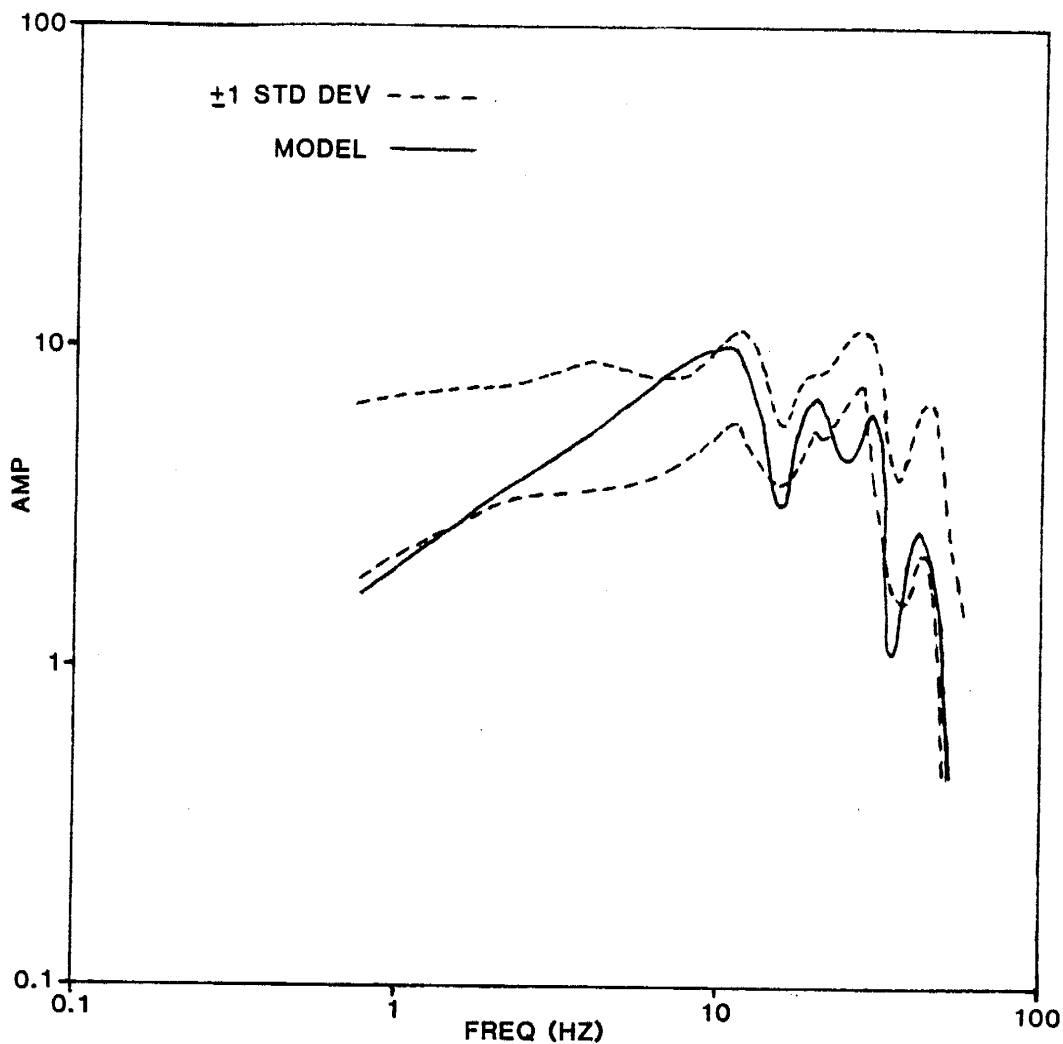


Figure 62. Comparison of synthetic spectra from the best-fitting two layer model (Figure 61) and the ± 1 standard deviation envelope to the stacked observed spectra. The dashed envelope represents uncertainties due to change in magma body structure as well as sampling and noise. The preferred model appears to be a good representation of the average magma body properties.

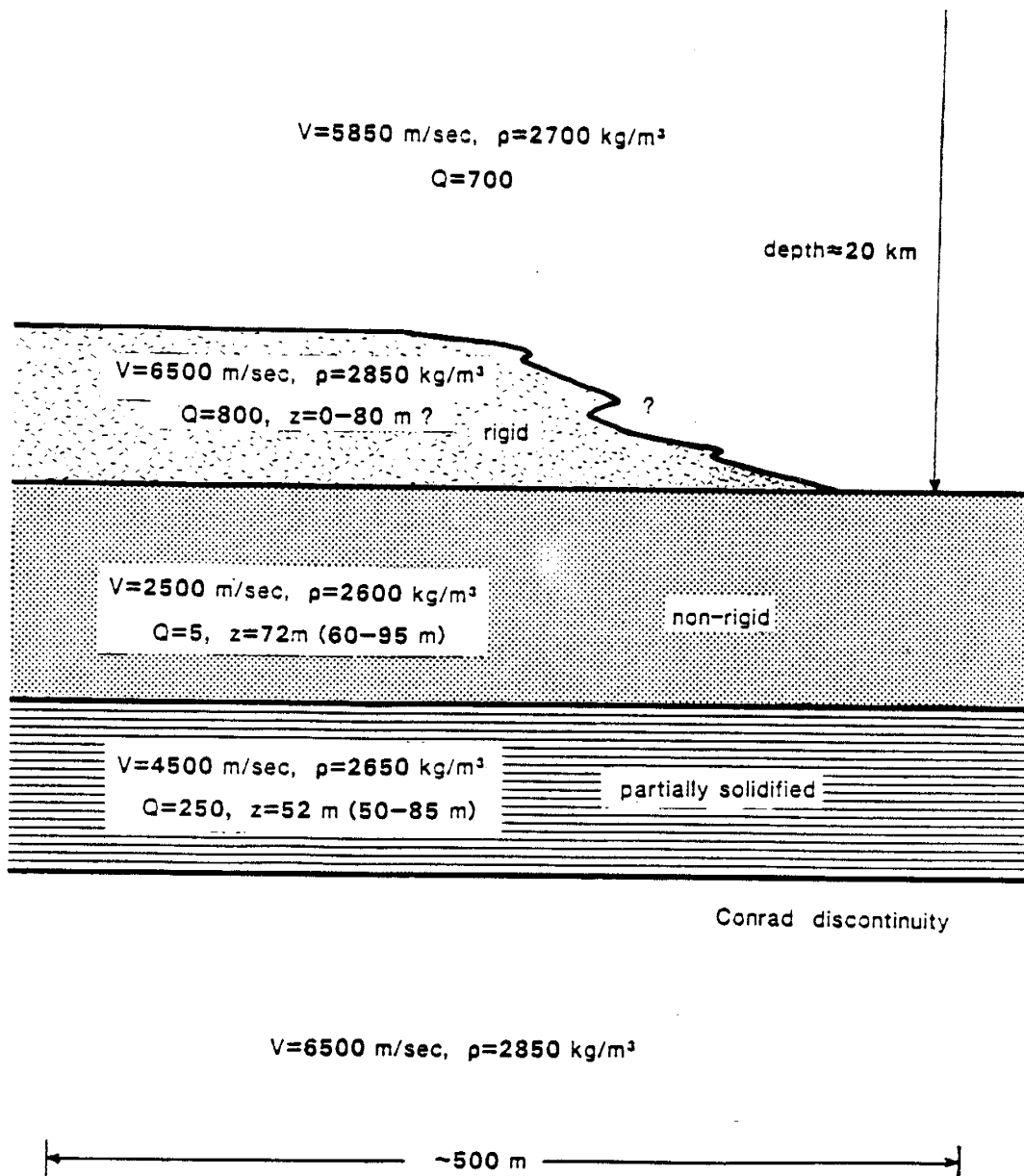
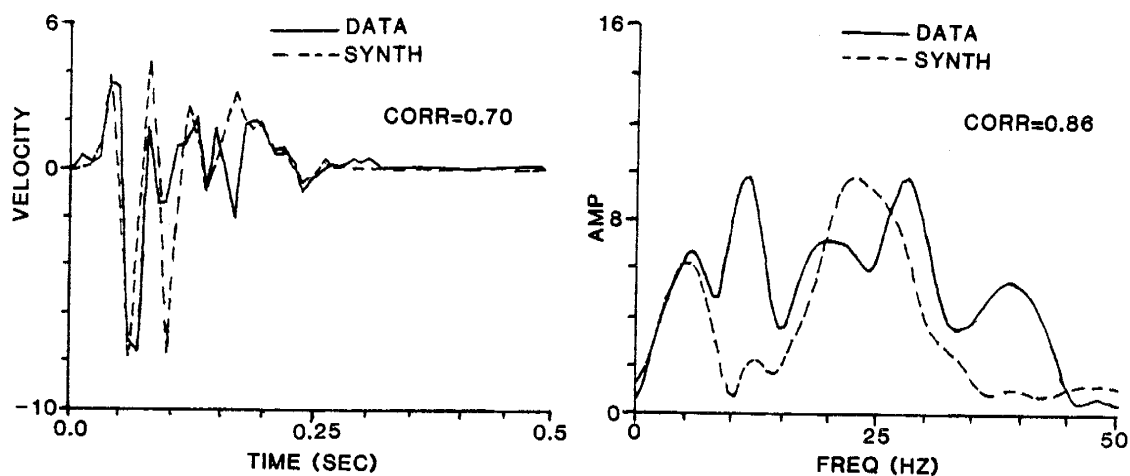


Figure 63. Model of magma body in the South Canyon area based on best fitting model to stacked $P_z P$ spectra.

answer this question, the model proposed by Brocher was compared to the observed data (Figure 64). Although the quality of fit to the observed data is somewhat lower than the two layer model, the multilayer model does reproduce many of the features of the observed data.

The favored multilayer models proposed by Brocher have a common characteristic: the layer thicknesses are all the same and the velocities of magma and country rock differ by a factor of two. This leads to the maximum constructive interference of a fundamental wavelength (or mode) with "overtones". If the velocities and thicknesses are perturbed slightly, the constructive interference begins to break down and the quality of fit is reduced (Figure 65). The relative merits of the proposed models will be discussed in the next section.



$$V_1 = 6000 \text{ m/s}, \rho_1 = 2700 \text{ kg/m}^3, Q_1 = 625, Z_1 = 10500 \text{ m}$$

$$V_2 = 3000 \text{ m/s}, \rho_2 = 2600 \text{ kg/m}^3, Q_2 = 5, Z_2 = 38 \text{ m}$$

$$V_3 = 6000 \text{ m/s}, \rho_3 = 2700 \text{ kg/m}^3, Q_3 = 625, Z_3 = 38 \text{ m}$$

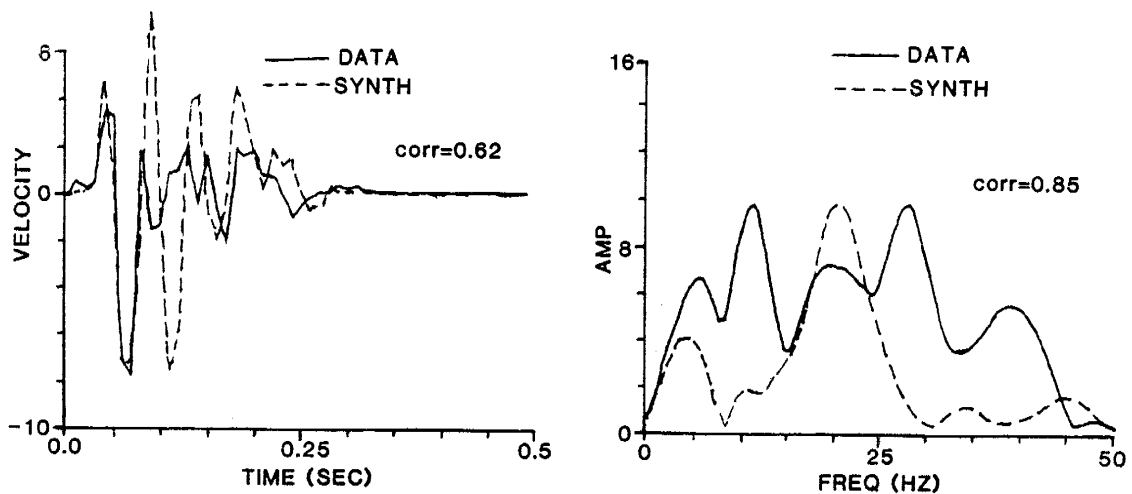
$$V_4 = 3000 \text{ m/s}, \rho_4 = 2600 \text{ kg/m}^3, Q_4 = 5, Z_4 = 38 \text{ m}$$

$$V_5 = 6000 \text{ m/s}, \rho_5 = 2700 \text{ kg/m}^3, Q_5 = 625, Z_5 = 38 \text{ m}$$

$$V_6 = 3000 \text{ m/s}, \rho_6 = 2600 \text{ kg/m}^3, Q_6 = 5, Z_6 = 38 \text{ m}$$

$$V_7 = 6000 \text{ m/s}, \rho_7 = 2700 \text{ kg/m}^3, Z_7 = \infty \text{ m}$$

Figure 64. Comparison of observed and synthetic time series and spectra, multi-layer (cyclic) model. Observed P_z waveform is from 114205 02-June 1977 event^z (Figure 22). The model used to generate the synthetic data is indicated. Correlation coefficient between synthetic and observed data is shown.



$$V_1 = 6000 \text{ m/s}, \rho_1 = 2700 \text{ kg/m}^3, Q_1 = 625, Z_1 = 10500 \text{ m}$$

$$V_2 = 2700 \text{ m/s}, \rho_2 = 2600 \text{ kg/m}^3, Q_2 = 5, Z_2 = 36 \text{ m}$$

$$V_3 = 5800 \text{ m/s}, \rho_3 = 2700 \text{ kg/m}^3, Q_3 = 625, Z_3 = 37 \text{ m}$$

$$V_4 = 2800 \text{ m/s}, \rho_4 = 2600 \text{ kg/m}^3, Q_4 = 5, Z_4 = 38 \text{ m}$$

$$V_5 = 5900 \text{ m/s}, \rho_5 = 2700 \text{ kg/m}^3, Q_5 = 625, Z_5 = 39 \text{ m}$$

$$V_6 = 3000 \text{ m/s}, \rho_6 = 2600 \text{ kg/m}^3, Q_6 = 5, Z_6 = 35 \text{ m}$$

$$V_7 = 6000 \text{ m/s}, \rho_7 = 2800 \text{ kg/m}^3, Z_7 = \infty$$

Figure 65. Comparison of observed and synthetic time series and spectra, multi-layer (cyclic) model with slight changes. Observed P-P waveform is from 114205 02-June 1977 event (Figure 22). The correlation coefficient between synthetic and observed data is shown.

DISCUSSION

All models of the type discussed here suffer from some degree of non-uniqueness. In this case the non-uniqueness manifests itself in two ways. First, it is always possible to "trade-off" layer thickness and velocity against each other to satisfy the constraints imposed by the data. However, once a velocity model is chosen, the permissible range of layer thicknesses is quite limited for this data set. This study has assumed that the published values for the velocities of basaltic melt at low pressures and temperatures are appropriate for the conditions considered here.

Christensen [1972; 1979] has published values for the change in compressional wave velocity with temperature and pressure for various types of solid rocks. It is assumed here that these values represent an upper bound to the actual changes in a full or mostly melt situation. The incompressible behaviour of a melt should ensure that changes in velocity would be less than those predicted for a solid material. Christensen's values for $(\frac{\partial V}{\partial P})_T$ and $(\frac{\partial V}{\partial t})_P$ for basalt at the depth of the magma body are summarized in Table 3. The value of pressure ($P=6$ kb) assumes a simple $P=\rho gh$ relation and the temperature at 20 km ($T=500^\circ\text{C}$) is calculated from values of geothermal gradient appropriate for the Basin and Range province [Blackwell, 1973]. The changes in velocity due to increasing pressure from 0 to 5 kbar and ambient temperature from 25 to 500°C tend to offset each other. The net change in velocity ($\Delta V_p = -0.02$ km/sec) for basalt at the pressure and temperature appropriate for a depth of 20 km in the Rio Grande rift would not seem to be significant. Unless the velocity

Table 3. Values of Temperature and Pressure Coefficients for Basalt (from Christensen [1979]).

Pressure Range	$(\frac{\partial V}{\partial P})_T$
(kbar)	(km/sec kbar)
3-5	0.035
5-8	0.016

Temperature Range	$(\frac{\partial V}{\partial T})_P$
(° C)	(km/sec ° C)
25-300	-0.39 E-3
300-400	-0.44 E-3
400-500	-0.52 E-3

$$\text{Net } \Delta V \approx -0.02 \text{ km/sec}$$

Assumed values: $Z=20$ km, $P=6$ kbar, $T=500^\circ\text{C}$, $(\frac{\partial V}{\partial P})_T$ from 0-3 kbar is the same as from 3-5 kbar.

values adopted for this analysis are greatly in error, the velocity-layer thickness tradeoff should not contribute to significant non-uniqueness.

The second manifestation of non-uniqueness that may arise is the inability to distinguish between two different models that produce similar seismic responses. This appears to be what is occurring with the two layer model favored by this study and the cyclic model favored by Brocher. The stacked P_zP spectra is compared to the best fit two layer model and the multi-layer model in Figure 66. Comparison of the maximum value of the cross-spectral function indicates the two layer model is a slightly better fit to the data. The simpler, two layer model is preferred here for a number of reasons. In addition to providing a somewhat better fit to the observed data than does the multilayer cyclic model, the simple two layer model is felt to be more geologically reasonable than a complex multi-layer "sandwich" with rather rigid constraints on layer thicknesses and velocities. Further, the similarity of seismic response at two widely separated locations argues strongly for the simpler model. Thin, extensive, sill-like intrusions (e.g. the Palisades sill or Bushveld complex) appear to more common in the geological record than the cyclic model. However, it should be kept in mind that there does not appear to be adequate evidence based just on the waveforms, to dismiss either of the two models.

The P_zP synthetics support the contention that the magma body is thin and has resolvable changes in internal structure over the areal extent of reflection points covered by this swarm. Although the general structure of sill-like bodies may be continuous over large lateral distances, the Bushveld complex for example [Hunter,1978], the

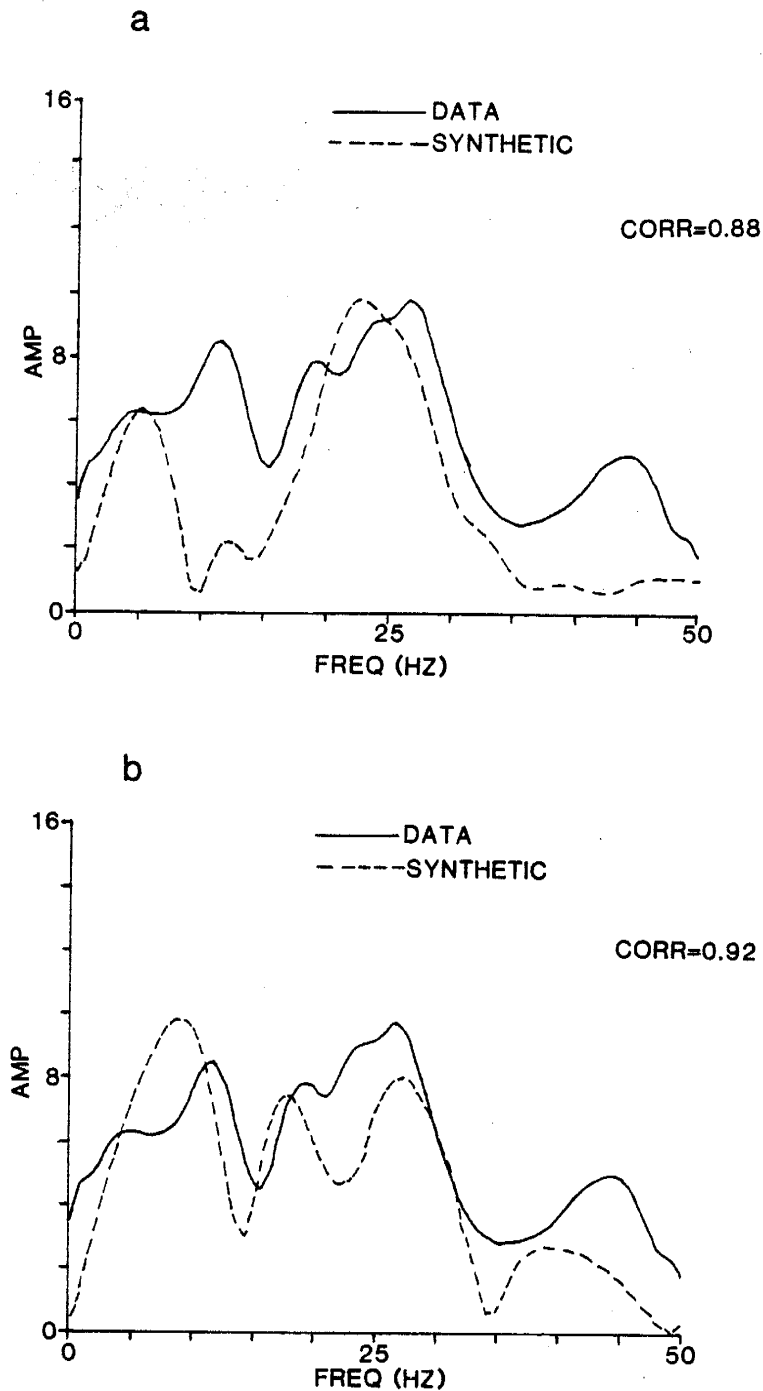


Figure 66. a) Comparison of stacked P-P spectra and spectra of multilayer (cyclic) model. Model parameters are identical to those in Figure 64. A three point moving average filter has been applied to both spectra.
 b) Comparison of stacked P-P spectra and spectra of best-fitting two layer model. Model parameters used to generate synthetic spectra are shown in Figure 63. A three point moving average filter has been applied to both spectra.

detailed characteristics may change vertically and horizontally. Rapid lateral changes in the structure of sill-like bodies are commonly observed [Tweto, 1951; Yoder, 1976; Hyndman, 1972].

All of the best-fitting two layer model synthetics have a pronounced 26 hz spectral peak (Figures 61, 62, 63, 64, and 65). Since the upper crustal response function does not show any sort of peak at this frequency, the 26 hz spectral peaks in the P_zP spectra are clearly a phenomena associated with the interaction of the downgoing seismic signal with the magma body.

In general, the quality of fit between the observed and synthetic data is excellent considering the simplifying assumptions that were made (see Figures 60 and 64). In particular, the P_zP phase occurs in a background of S phase coda which will contribute to any discrepancy between the synthetic and observed data. Also, the path of the direct and reflected P-phases between the focal depth and surface is quite similar but not identical. These factors may contribute to the slight misfit between the observed and synthetic data.

Recent reprocessing of COCORP lines 3 and 4 [de Voogd et al., 1987] has identified the mid-crustal magma body as far south as VP100 on line 3. Station SC lies about 8 km south and west of VP100. It is inferred that the magma accumulations imaged by the two studies are related. However, spatial continuity cannot be proven at this time. It is quite conceivable that the magma extends further to the south on lines 3 and 4. Source coupled noise in the bandwidth crucial to the magma body analysis (25-30 hz), prevented a detailed analysis of this portion of the COCORP data [de Voogd, 1986].

Microearthquake seismograms in the Socorro area do not indicate the presence of significant continuous reflectors at depths between 5 and 20 km [A. Sanford, pers. comm.]. This type of "transparent" seismic behaviour has also been observed for the upper crust in other deep reflection studies [Litak and Brown, 1987]. Below about 20 km however, COCORP data in both the Rio Grande rift and Death Valley reveal fairly strong, subhorizontal reflections [de Voogd et al., 1987]. The presence of a high-velocity lid and the simple layered structure of the mid-crustal magma body postulated by this study suggests how this type of layered lower crust could be created. The intrusion of sills of mafic material into sialic crust seems to produce the observed type of seismic response [Phinney et al., 1987]. This suggests that the mechanism for crustal growth in this area is by an underplating process. Intrusion of extensive, sill-like bodies into the crust along zones separating material of different rheological properties may be occurring.

The rather thin nature of the magma body suggests that it may be fortuitous to actually observe a sill in a molten state. Estimates of the thermal lifetime of such a body range from a few tens of years to hundreds of years (Appendix IV).

SUGGESTIONS FOR FURTHER RESEARCH

A number of questions that have arisen in the course of this study are worthy of further investigation. A few will be discussed below.

It would be of considerable interest to apply the modelling technique devised by this study to other microearthquake swarms in the Socorro area. Both the data set considered in the present study as well as that analyzed by Brocher were for locations near the margin of the magma body. It is still not clear that areas considerably away from the magma body margin will have a similar structure to that postulated here. If the magma body is really as thin as our results indicate over its entire areal extent, some interesting speculations regarding the mechanism of magma emplacement present themselves. The existing digital data base would need to be supplemented with additional recordings from either the permanent network or temporary installations.

The one dimensional modelling scheme may be applied to studying the problem of near surface reverberations. This phenomena produces a profound effect on the recorded seismograms at many sites. Because of the low velocity of near surface materials, the angle of incidence is often quite steep. Hence the one dimensional modelling should work well. Reverberations within one, two, or three layers above a half-space could be modelled with this technique.

Modelling of the direct P phase could also lead to estimates of Q_p . This is a completely different approach to estimating Q than the

spectral slope methods which have been utilized before. As such it could be used as both an independent estimator of upper crustal Q , or as corroboration for other methods.

The success in modelling the P_zP phase indicates that for this type of geometry and event strengths, utilizing the direct P phase as the upper crustal response function is a valid technique. This should also be useful in earthquake source studies where it is necessary to isolate source effects from contamination by near surface effects. Several microearthquake swarms in the Socorro area have events that cover a broad range of magnitudes. It would be interesting to see if source properties can be recovered using this approach.

An alternative method for ascertaining the properties of the magma body is to employ an inversion technique which uses the seismograms as the known data. A brief (and not very successful) attempt at this technique was made during the course of this study. It was never quite clear what level of uncertainty to assign to the individual data points. A more appropriate method of inversion might be to do the inversion in the frequency domain so that the stacked spectra could be used as the known data. This allows for a clearer definition of the uncertainty in each spectral value.

How good is the one-dimensional approximation used to calculate the synthetic seismograms? The calculated Fresnel zone diameter of 1.75 to 3.5 km suggests that there may be two-dimensional effects present. The observed data on the other hand, would suggest a one-dimensional approach is adequate. A set of two dimensional synthetic seismogram calculations could help resolve this dichotomy.

SUMMARY AND CONCLUSIONS

The availability of both well recorded P and S phase reflections at one location provides a unique data set with which to investigate the existence and properties of the mid-crustal reflector. Based upon the collective behavior of the microearthquake reflections (S_zS and P_zP) from the South Canyon area near Socorro, the first major reflecting horizon of the mid-crustal reflector must be the upper surface of a non-rigid layer. Cross-correlation of the direct and reflected microearthquake S phases yields coefficients which are both positive and large in absolute value (avg=0.79). Because these two phases leave the focal region 180° out of phase, reflection must occur at a discontinuity where a phase reversal occurs. The similarity in S_zS waveforms and uniformity in the S/S_zS amplitude ratio requires that the upper surface of the magma body be very simple with little or no topography present over the area covered by the reflection points. Waveforms for the P phase reflections are not identical from event to event within the swarm. The observed changes in P_zP waveforms must then be arising from changes in thickness and or internal structure that are occurring over the area covered by the reflection points. This observation is somewhat at odds with the calculated Fresnel zone diameter of 1.75-3.5 km.

The near normal incidence geometry of both the reflected and direct P phase implies an essentially common path through the upper crust and near surface Phanerozoic rocks for these two phases. Because the earthquakes examined in this study were quite small, the initial portion of the direct P phase was used as the response function

for the path through the upper crust. This allowed the isolation of magma body effects from upper crustal effects. Synthetic seismograms, which included attenuation effects and interbed multiples, were calculated and compared to the observed waveforms.

Forward modelling of the P_zP phase supports the hypothesis that the magma body is quite thin yet has resolvable differences in internal structure over the areal extent of the reflection points covered by the swarm. The magma body in this area was best approximated by a two layer structure; a non-rigid, low-velocity layer of full melt ≈ 70 m thick underlain by an intermediate velocity layer ≈ 60 m thick. The lower layer may be crystalline mush or older partially solidified magma. In places the magma body may have a discontinuous, thin, high velocity lid. Using the stacked P_zP spectra as a reference, an average model for the magma body was calculated. This model shows total magma body thickness to be less than ≈ 160 m.

Although not clearly imaged on portions of COCORP lines 3 and 4, an accumulation of magma at mid-crustal levels is clearly present near South Canyon southwest of Socorro. Whether or not the magma in this area is "connected" to the main magma body further north and east is not well established.

Earlier workers had proposed a multilayer, cyclic model to explain observations north and east of Socorro. Although the evidence presented here is not adequate to dismiss the multilayer, cyclic model out of hand, the simple, two layer model does fit the observed data better than does the more complex, multilayer, cyclic model. The simple model is preferred on the basis of geological plausibility.

The success in modelling the P_zP phase indicates that for this type of geometry and event strengths, utilizing the direct P phase as the upper crustal response function is a valid modelling technique. This technique will also be very useful in earthquake source studies where it is also desirable to separate near-surface and source effects.

REFERENCES

- Ake, J. P. (1984). An analysis of the May and July, 1983 Socorro Mtn. microearthquake swarms, *Geophysics Open File Rep. No. 49 (M.S. Indep. Study)*, New Mexico Inst. of Mining and Technol., Socorro, N.M., 107 p.
- Ake, J., S. Jarpe, and A. Sanford (1984). Determination of source characteristics for recent microearthquake swarms in the Socorro area (abs.), *New Mexico Geol. Soc. Meeting Abstracts*.
- Aki, K., and P. G. Richards (1980). *Quantitative Seismology: Theory and Methods*, vols. 1 and 2, W. H. Freeman and Co., San Francisco, Calif.
- Anstey, N. A. (1977). *Seismic Interpretation: The Physical Aspects*, Internl. Human Resourc. Devlp. Corp., Boston, Mass.
- Balesta, S. T., and A. I. Faberov (1968). Seismic studies of Piip crater breakthrough, *Bull. Volcan.*, *32*, 395-399.
- Bjornsson, A., G. Johnson, S. Sigurdsson, G. Thorbergsson, and E. Tryggvasson (1979). Rifting of the plate boundary in north Iceland 1975-1978, *J. Geophys. Res.*, *84*, 3029-3038.
- Blackwell, D. D. (1971). The thermal structure of the continental crust, in *The Structure and Physical Properties of the Earth's Crust*, *Geophys. Monogr. Ser.*, Vol. 14, ed. J. G. Heacock, p. 169-184, Amer. Geophys. Union, Washington, D.C.
- Brocher, T. M. (1981). Geometry and physical properties of the Socorro, New Mexico magma bodies, *J. Geophys. Res.*, *86*, 9420-9432.
- Brown, L. D., P. A. Krumhansl, C. E. Chapin, A. R. Sanford, F. A. Cook, S. Kaufman, J. E. Oliver, and F. S. Schilt (1979). COCORP seismic reflection studies of the Rio Grande rift, in *Rio Grande Rift: Tectonics and Magmatism*, ed. R. Riecker, pp 169-184, Amer. Geophys. Union, Washington, D.C.
- Brown, L. D., C. E. Chapin, A. R. Sanford, S. Kaufman, and J. Oliver (1980). Deep structure of the Rio Grande rift from seismic reflection profiling, *J. Geophys. Res.*, *85*, 4773-4800.
- Bullen, K. E., and B. A. Bolt (1985). *An Introduction to the Theory of Seismology*, 4th ed., Cambridge Univ. Press, Cambridge, England.
- Carpenter, P. J. (1984) Apparent Q for upper-crustal rocks in the Rio Grande rift of central New Mexico from the analysis of

microearthquake spectra, *Ph. D. Thesis*, New Mexico Inst. of Mining and Technol., Socorro, N. M., 150 p.

- Carpenter, P. J., and A. R. Sanford (1985). Apparent Q for upper-crustal rocks of the central Rio Grande rift, *J. Geophys. Res.*, *90*, 8661-8674.
- Cather, S. M. (1983). Laramide Sierra uplift: Evidence for major pre-rift uplift in central and southern New Mexico, in C. Chapin ed., *N. Mex. Geol. Soc. Guidb.*, *34*, 99-101.
- Chapin, C. E. (1971). The Rio Grande Rift, Part I: Modifications and additions, *N. Mex. Geol. Soc. Field. Conf. Guidb.*, *21*, 191-201.
- Chapin, C. E., and W. R. Seager (1975). Evolution of the Rio Grande rift in the Socorro and Las Cruces areas, *N. Mex. Geol. Soc. Field Conf. Guidb.*, *26*, 297-322.
- Chapin, C. E., R. M. Chamberlin, G. R. Osburn, D. W. White, and A. R. Sanford (1978). Exploration framework of the Socorro Geothermal Area, New Mexico, *Spec. Publ. N. Mex. Geol. Soc.*, *7*, 114-129.
- Condie, K. C., and A. J. Budding (1979). Geology and geochemistry of Precambrian rocks, central and south-central New Mexico, *New Mex. Bur. of Mines and Miner. Resourc. Mem.* *35*.
- Christensen, N. I. (1974). Compressional and shear wave velocities at pressures to 10 kilobars for basalts from the East Pacific rise, *Geophys. J. Roy. Astron. Soc.*, *28*, 425-429.
- Christensen, N. I. (1979). Compressional wave velocities in rocks at high temperatures and pressures, critical thermal gradients, and crustal low-velocity zones, *J. Geophys. Res.*, *84*, 6849-6857.
- Cordell, L. (1978). Regional geophysical setting of the Rio Grande rift, *Geol. Soc. Am. Bull.*, *89*, 1073-1090.
- Cranswick, E., R. Wetmiller, and J. Boatwright (1985). High-frequency observations and source parameters of microearthquakes recorded at hard-rock sites, *Bull. Seism. Soc. Am.*, *75*, 1535-1567.
- Dagget, P. H., G. R. Keller, P. Morgan, and C. L. Wen (1986). Structure of the southern Rio Grande rift from gravity interpretation, *J. Geophys. Res.*, *91*, 6157-6167.
- Daniel, R. G., and D. M. Boore (1982). Anomalous shear wave delays and surface wave velocities at Yellowstone caldera, Wyoming, *J. Geophys. Res.*, *87*, 2731-2744.

- DeVoogd, B. (1986). Deep structures and magmatic processes in two continental rifts: studies using COCORP seismic section reflection profiling in Death Valley and in the Rio Grande rift, *Ph. D. Thesis*, Cornell Univ., Ithaca, N.Y., 187 p.
- de Voogd, B. L., L. Serpa, and L. D. Brown (1987). Crustal extension and magmatic processes: COCORP profiles from Death Valley and the Rio Grande rift, *Geol. Soc. Am. Bull.*, 98, in press.
- Einarsson, P. (1978). S-wave shadows in the Krafla caldera in NE-Iceland, evidence for a magma chamber in the crust, *Bull. Volcanol.*, 41, 1-9.
- Evoy, R. J. (1978). Precision gravity reobservations and simultaneous inversion of gravity and seismic data for subsurface structure of Yellowstone region, *M. S. Thesis*, Univ. of Utah, Salt Lake City, Utah.
- Frankel, A. (1982). The effects of attenuation and site response on the spectra of microearthquakes in the northeastern Caribbean, *Bull. Seism. Soc. Am.*, 72, 1379-1402.
- Frankel, A., and H. Kanamori (1983). Determination of rupture duration and stress drop for earthquakes in southern California, *Bull. Seism. Soc. Amer.*, 73, 1527-1551.
- Gorshkov, G. S. (1958). On some theoretical problems of volcanology, *Bull. Volcan.*, 19, 103-113.
- Huppert, H. E., R. S. J. Sparks, J. A. Whitehead, and M. A. Hallworth (1986). Replenishment of magma chambers by light inputs, *J. Geophys. Res.*, 91, 6113-6122.
- Hyndman, D. W. (1972). *Petrology of Igneous and Metamorphic Rocks*, McGraw-Hill, New York, N. Y.
- Iyer, H. M. (1984). Geophysical evidence for the locations, shapes and sizes, and internal structures of magma chambers beneath regions of Quaternary volcanism, *Phil. Trans. R. Soc. London A.*, 310, 473-510.
- Iyer, H. M., and R. M. Stewart (1977). Teleseismic technique to locate magma in the crust and upper mantle, in *Magma Genesis, Proceed. of the Amer. Geophys. Union Chapman Conf. on Partial Melting in the Earth's Upper Mantle*, ed. by H. J. B. Dick, 281-299, Bull. 96, State of Oregon Dept. of Geol. and Miner. Industries.
- Iyer, H. M., J. R. Evans, G. Zandt, R. M. Stewart, J. M. Coakley, and J. N. Rollof (1981). A deep low-velocity body under the Yellowstone caldera, Wyoming: Delineation using teleseismic P-wave

- residuals and tectonic interpretation, *Bull. Geol. Soc. Amer.*, 92, 1471-1646.
- Jachens, R. C., and C. W. Roberts (1985). Temporal and areal gravity investigations at Long Valley caldera, California, *J. Geophys. Res.*, 90, 11210-11218.
- Jarpe, S. (1984). Characteristics and possible cause of an earthquake swarm in the central Rio Grande rift 28 km north of Socorro, New Mexico, *Geophys. Open File Rep. No. 50 (M. S. Indep Study)*, New Mex. Inst. of Mining and Technol., Socorro, 90 p.
- Jenkins, G. M., and D. G. Watts (1968). *Spectral Analysis and its Applications*, Holden-Day, San Francisco, CA.
- Johnston, J. A. (1978). Microearthquake frequency attenuation of S phases in the Rio Grande rift near Socorro, New Mexico, *Geophys. Open File Rep. No. 24 (M. S. Indep. Study)*, New Mexico Inst. of Mining and Technol., Socorro, 84 p.
- Kanasewich, E. R. (1981). *Time Sequence Analysis in Geophysics*, 3rd ed., Univ. of Alberta Press, Edmonton, Alberta, Canada.
- Keller, G. V. (1982). Electrical properties of rocks and minerals, in *Handbook of Physical Properties of Rocks, v. I*, ed. by R. S. Carmichael, CRC Press, Boca Raton, FL.
- Kelley, V. C., and A. M. Kudo (1978). Volcanoes and related basalts of Albuquerque basin, *Circ. 156, N. Mex. Bur. of Mines and Miner. Resourc.*, Socorro.
- Larsen, S., R. Reilinger, and L. Brown (1986). Evidence of ongoing crustal deformation related to magmatic activity near Socorro, New Mexico, *J. Geophys. Res.*, 91, 6283-6292.
- Lee, W. H. K., and J. C. Lahr (1975). HYPO71 (revised): A computer program for determining hypocenter, magnitude, and first motion pattern of local earthquakes, *U. S. Geol. Surv. Open File Rep.*, 75-311, 116 p.
- Litak, R. K., and L. D. Brown (1987). A modern perspective on the Conrad discontinuity (abs), *EOS Trans. Amer. Geophys. Union*, 68, p 350.
- Machette, M. N., and R. G. McGimsey (1982). Quaternary and Pliocene faults in the Socorro and western part of the Fort Sumner 1x2 quadrangles, New Mexico, *U. S. Geol. Surv. Misc. Field Studies Map, MF-1465-A*.
- Mizoue, M., I. Nakamura, and T. Yokota (1982). Mapping of an unusual crustal discontinuity by microearthquake reflections in

- the earthquake swarm area near Ashio, northwestern part of Tochigi Prefecture, Central Japan, *Bull. of Earthquake Res. Inst.*, 57, 653-686.
- Morgan, P., D. D. Blackwell, and R. E. Spafford (1977). Heat flow measurements in Yellowstone Lake and the thermal structure of the Yellowstone caldera, *J. Geophys. Res.*, 82, 3719-3732.
- Murase, T., and A. R. McBirney (1973). Properties of some common igneous rocks and their melts at higher temperatures, *Geol. Soc. Am. Bull.*, 84, 3563-3592.
- O'Neill, M. E., and J. H. Healy (1973). Determination of source parameters of small earthquakes from P-wave rise time, *Bull. Seism. Soc. Am.*, 63, 599-614.
- Osburn, G. R., and C. E. Chapin (1983). Ash flow tuffs and cauldrons in the northeast Mogollon-Datil volcanic field: A summary, in C. E. Chapin ed., *N. Mex. Geol. Soc. Guidb.*, 34, 197-204.
- Phinney, R. A., G. S. Pan, and K. Roychowdury (1987). On the occurrence of reflective, strongly layered zones in the mid-to lower crust (abs), *EOS Trans. Amer. Geophys. Union*, 68, p 350.
- Reilinger, R. E., J. E. Oliver, L. D. Brown, A.R. Sanford, and E. I. Balazs (1980). New measurements of crustal doming over the Socorro magma body, *Geology*, 8, 291-295.
- Reiter, M., and R. Smith (1977). Subsurface temperature data in the Socorro Peak KGRA, New Mexico, *Geotherm. Energy Mag.*, 5, 37-41.
- Richter, C. F. (1958). *Elementary Seismology*, W. H. Freeman and Co., San Francisco, Calif.
- Ricker, N. (1953). The form and laws of propagation of seismic wavelets, *Geophysics*, 18, 10-40.
- Riecker, R. E. ed. (1979). *Rio Grande Rift: Tectonics and Magmatism*, Amer. Geophys. Union, Washington, D. C.
- Rinehart, E. J., and A. R. Sanford (1981). Crustal structure of the Rio Grande rift near Socorro, New Mexico from the inversion of microearthquake S-wave reflections, *Bull. Seism. Soc. Amer.*, 71, 437-450.
- Rinehart, E. J., A. R. Sanford, and R. M. Ward (1979). Geographic extent and shape of an extensive magma body at mid-crustal depths in the Rio Grande rift near Socorro, New Mexico, in *Rio Grande Rift: Tectonics and Magmatism*, ed. R. Riecker, pp 237-251, Amer. Geophys. Union, Washington, D.C.

- Robinson, E. (1970). *Multichannel Time Series Analysis with Computer Programs*, Holden-Day and Co., San Francisco, Calif.
- Ryall, F., and A. Ryall (1981). Attenuation of P and S waves in a magma chamber in Long Valley caldera California, *Geophys. Res. Lett.*, 8, 557-560.
- Sanford, A. R. (1968). A gravity survey in central Socorro County, New Mexico, *N. Mex. Bur. Mines Miner. Resourc.*, 120, 19 p.
- Sanford, A. R., and C. R. Holmes (1962). Microearthquakes near Socorro, New Mexico, *J. Geophys. Res.*, 67, 4449-4459.
- Sanford, A. R., and L. T. Long (1965). Microearthquake crustal reflections, Socorro, New Mexico, *Bull. Seism. Soc. Amer.*, 55, 579-586.
- Sanford, A. R., and J. W. Schlue (1980). Seismic exploration for shallow magma bodies in the vicinity of Socorro, New Mexico, New Mexico Energy Inst. and New Mexico State Univ., *NMEI*, 56.
- Sanford, A. R., and P. Einarsson (1982). Magma Chambers in Rifts, in *Continental and Oceanic Rifts, Geodyn. Ser.*, vol. 8, ed. by G. Palmsson, Amer. Geophys. Union, Washington, D.C., 147-168.
- Sanford, A. R., A. J. Budding, J. P. Hoffman, O. S. Alptekin, C. A. Rush, and T. R. Topozada (1972). Seismicity of the Rio Grande rift in New Mexico, *Circ. N. Mex. Bur. Mines and Miner. Resourc.*, 120, 19 p.
- Sanford, A. R., O. S. Alptekin, and T. R. Topozada (1973). Use of reflection phases on microearthquake seismograms to map an unusual discontinuity beneath the Rio Grande rift, *Bull. Seism. Soc. Amer.*, 63, 2021-2034.
- Sanford, A. R., R. P. Mott Jr., P. J. Shuleski, E. J. Rinehart, F. J. Caravella, R. M. Ward, and T. C. Wallace (1977). Geophysical evidence for a magma body in the vicinity of Socorro, New Mexico, in *The Earth's Crust*, ed. by J. G. Heacock, pp 385-403, Geophys. Mon. 20, Amer. Geophys. Union, Washington, D.C.
- Sanford, A. R., P. J. Carpenter, and E. J. Rinehart (1983a). Characteristics of a microearthquake swarm in the Rio Grande rift near Socorro, New Mexico (abs.), *Earthquake Notes*, 54, p 98.
- Sanford, A. R., J. P. Ake, S. P. Jarpe, P. J. Carpenter, and L. H. Jaksha (1983b). Source independent spectra up to M=1.2 implied from duplication of P-phase waveforms in a Rio Grande rift microearthquake swarm (abs.), *EOS Trans. Amer. Geophys. Union*, 64, p 752.

- Sanford, A. R., P. J. Carpenter, E. J. Rinehart, and D. P. Weider (1984). Anatomy of a small earthquake swarm in the central Rio Grande rift near Socorro, New Mexico, *Geophys. Open-File Rep. No. 56*, New Mex. Inst. of Mining and Technol., Socorro, 36 p.
- Savage, J. C., and M. M. Clark (1982). Magmatic resurgence in the Long Valley caldera, California: Possible cause of the 1980 Mammoth Lakes earthquakes, *Science*, *217*, 531-533.
- Sharpe, J. A. (1941). The production of elastic waves by explosion pressures, I, Theory and empirical field observations, *Geophysics*, *7*, 144-154.
- Shuleski, P. J. (1976). Seismic fault motion and SV screening by shallow magma bodies in the vicinity of Socorro, New Mexico, *Geophys. Open File Rep. 8 (M. S. Indep. Study)*, New Mex. Inst of Mining and Technol., Socorro, 94 p.
- Smith, R. B., R. T. Shuey, J. R. Pelton, and J. P. Bailey (1977). Yellowstone hot spot: Contemporary tectonics and crustal properties from earthquake and aeromagnetic data, *J. Geophys. Res.*, *82*, 3665-3676.
- Steeple, D. W., and H. M. Iyer (1976). Low-velocity zone under Long Valley as determined from teleseismic events, *J. Geophys. Res.*, *81*, 849-860.
- Thayer, R. E., A. Bjornsson, L. Alvarez, and J. F. Hermance (1981). Magma genesis and crustal spreading in the northern neovolcanic zone of Iceland: Telluric-magnetotelluric constraints, *Geophys. Jour. Royal Astron. Soc.*, *65*, 423-442.
- Topozada, T. R., and A. R. Sanford (1976). Crustal structure in central New Mexico interpreted from the Gasbuggy explosion, *Bull. Seism. Soc. Amer.*, *66*, 877-886.
- Tryggvason, E. (1980). Subsidence events in the Krafla area, *J. Geophys.*, *47*, 141-153.
- Turcotte, D. L., and G. Schubert (1982). *Geodynamics*, John Wiley and Sons, New York, N. Y.
- Tweto, O. (1951). Form and structure of sills near Pando, Colorado, *Geol. Soc. Amer. Bull.*, *62*, 507-532.
- Utnasin, V. K., A. I. Abdurakhmanov, G. I. Ansovoc, A. Yu, V. Fedorchenka, Y. K. Markhinin, and S. T. Balesta (1976). Geophysical investigations of the Kamchatka area, in *Volcanoes and Tectosphere*, ed. by H. Aoki and S. Iizuka, Tokyo Univ. Press, Tokyo, Japan, 123-127.

- Ward, R. M. (1980). Determination of three-dimensional velocity anomalies within the upper crust in the vicinity of Socorro, New Mexico using P-arrival times from local earthquakes, *Ph. D. Dissertation*, New Mex. Inst. of Mining and Technol., Socorro, 200 p.
- Ward, R. M., J. W. Schlue, and A. R. Sanford (1981). Three dimensional velocity anomalies in the upper crust near Socorro, New Mexico, *Geophys. Res. Lett.*, 8, 553-556.
- Widess, M. B. (1973). How thin is a thin bed?, *Geophysics*, 38, 1176-1180.
- Yoder, H. S. (1976). *Generation of Basaltic Magma*, National Academy of Sciences, Washington, D.C.
- Zoeppritz, K. (1919). Uber Erdbebenwellen, *Gottinger Nachrichten*, VIIb, 66-84.

Appendix I: Data

APPENDIX I

This appendix contains a listing of the earthquake locations used in the analysis of the May-June 1977, microearthquake swarm (Table I-1). Figures I-1 through I-8 present individual, large format plots of the earthquakes displayed as a group in Figure 19.

Most spectra in this study were smoothed in the frequency domain with a three-point moving average filter (weights = 0.0, 0.5, 1.0, 0.5, 0.0). The smoothing process has certain ramifications that must be considered.

When the exact type of lag (or spectral) window is defined, the variance of the spectral estimator may be obtained. Knowing the variance, it is possible to construct a confidence interval for the true spectrum [Jenkins and Watts, 1968]. The smoothing process reduces variance but adds bias. If the truncation point of the lag window (M) is small, the variance of the smoothed spectral estimator is reduced. Reducing M increases the bias however, resulting in greater distortion of the spectrum (since the spectral window becomes broader).

As the relative position of spectral features was of interest, the spectral window was kept narrow. In other words, increased variance was tolerated to reduce spectral distortion. The 80% confidence interval for one of the P_zP spectra is shown in Figure I-9.

Table I-1. Summary of Earthquake Locations for May-June 1977
Microearthquake Swarm.

<u>DATE</u>	<u>OT</u>	<u>LAT</u>	<u>LONG</u>	<u>DEPTH</u>	<u>NO. STA.</u>
060277	064550	3400.70	10703.36	9.97	5
060277	064816	3400.70	10703.39	9.31	7
060277	065024	3400.84	10703.36	9.23	5
060277	065521	3400.66	10703.38	9.75	5
060277	081147	3401.10	10703.27	9.40	5
060277	114200	3401.12	10703.33	9.77	5
060277	120703	3401.04	10703.39	10.20	5
060277	173008	3401.01	10703.22	10.26	5

<u>NO. READ.</u>	<u>GAP</u>	<u>RMS</u>	<u>ERH</u>	<u>ERZ</u>	<u>Q</u>
10	140	0.03	0.2	0.2	B
7	140	0.02	0.3	0.5	B
10	134	0.06	0.4	0.5	B
10	141	0.06	0.4	0.4	B
9	124	0.06	0.4	0.5	B
10	123	0.04	0.2	0.3	B
10	126	0.04	0.2	0.3	B
10	127	0.06	0.5	0.5	B

Note: DEPTH, ERH AND ERZ are in km,
GAP is in degrees,
RMS is in seconds and
OT is in Universal time.

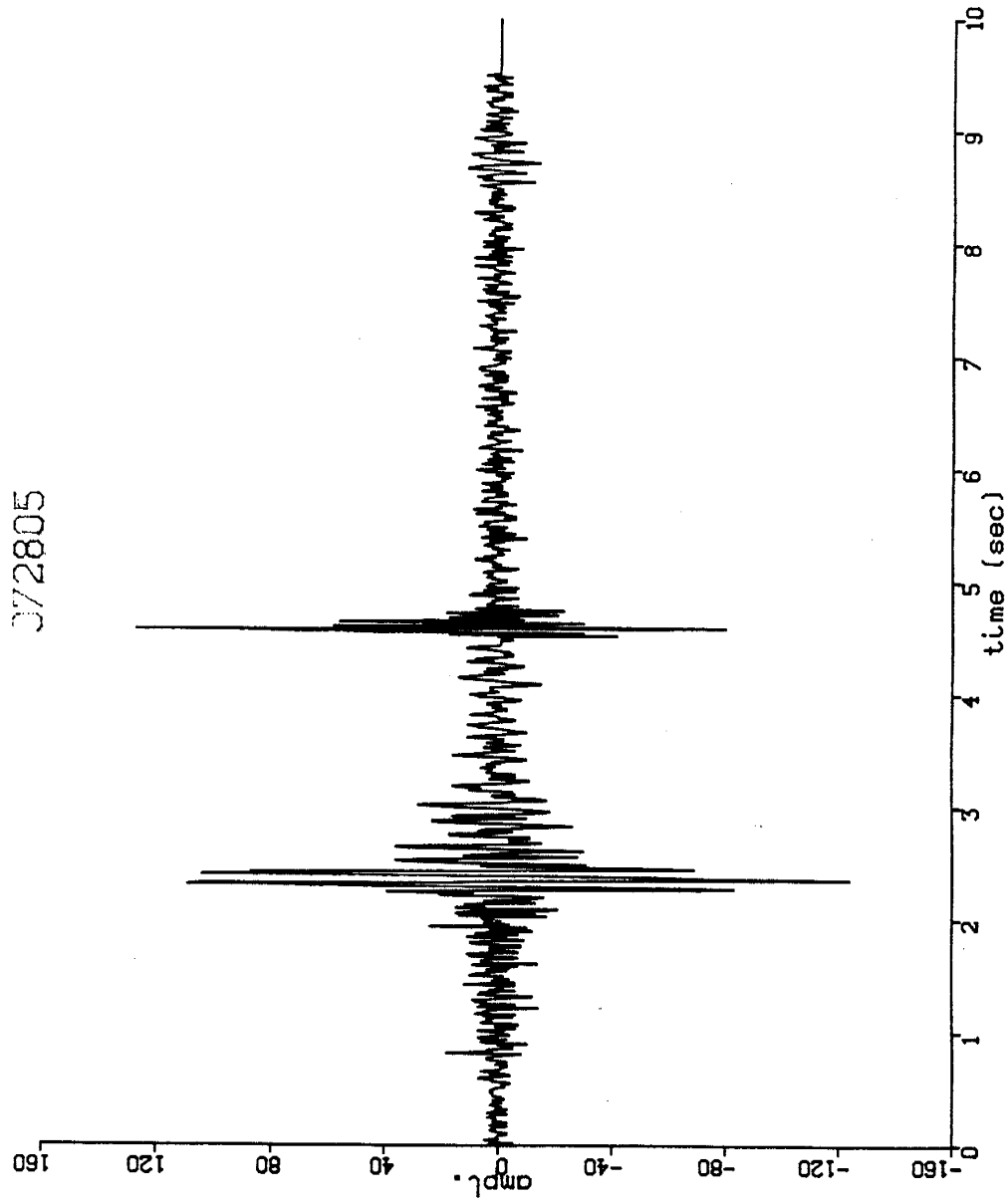


Figure I-1. May 30, 1977 072805 event.

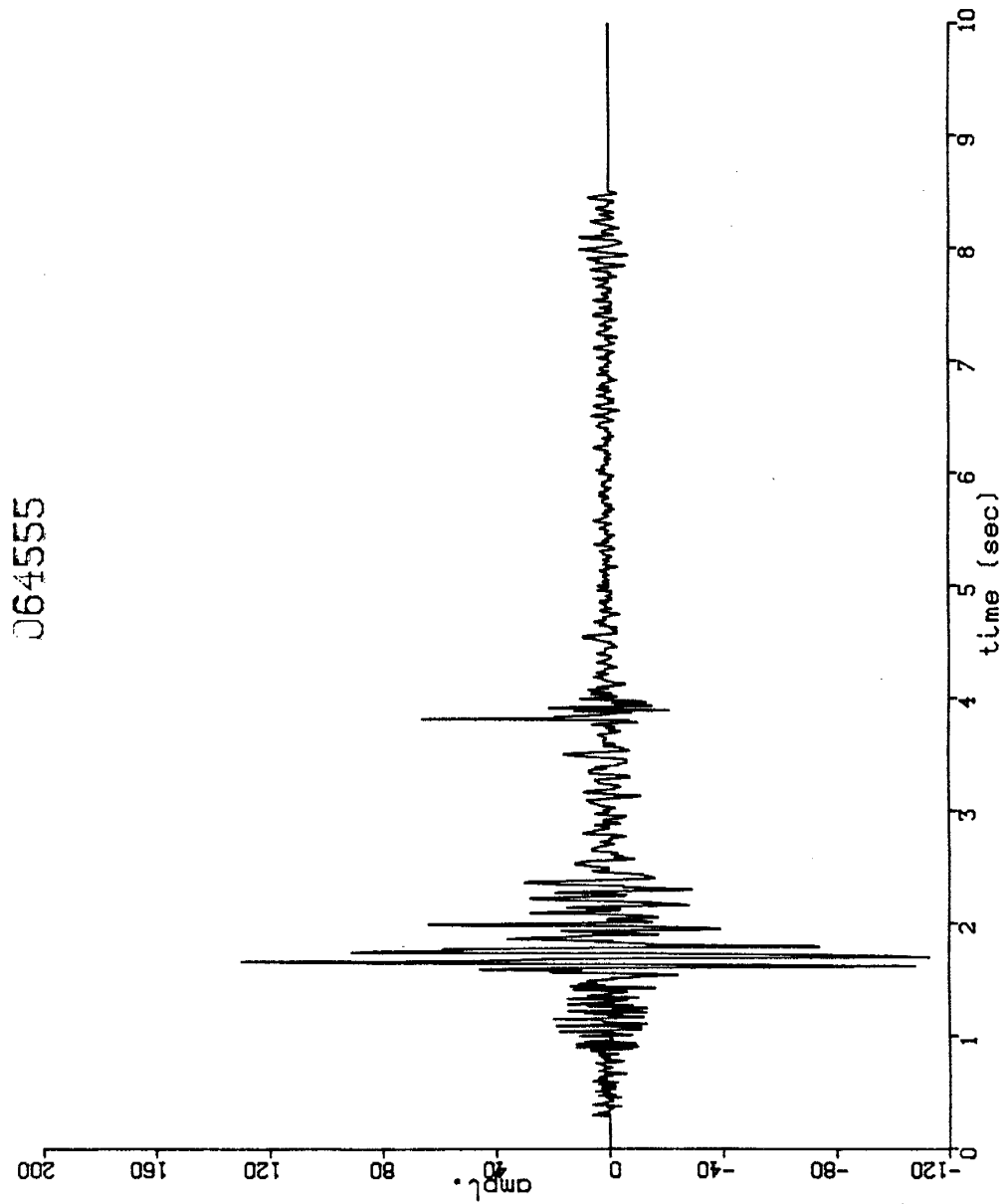


Figure I-2. June 2, 1977 064555 event.

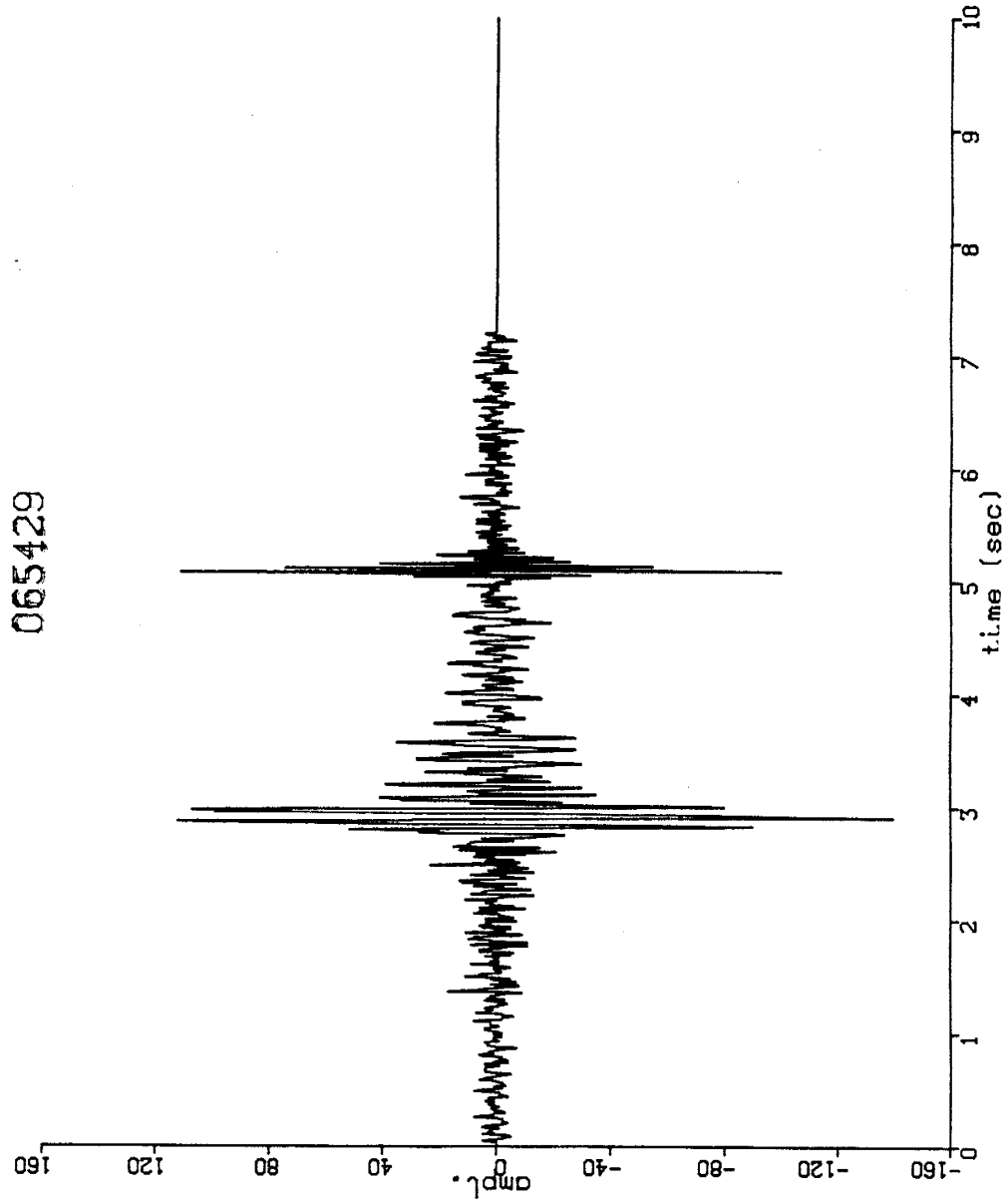


Figure I-3. May 30, 1977 065429 event.

070327

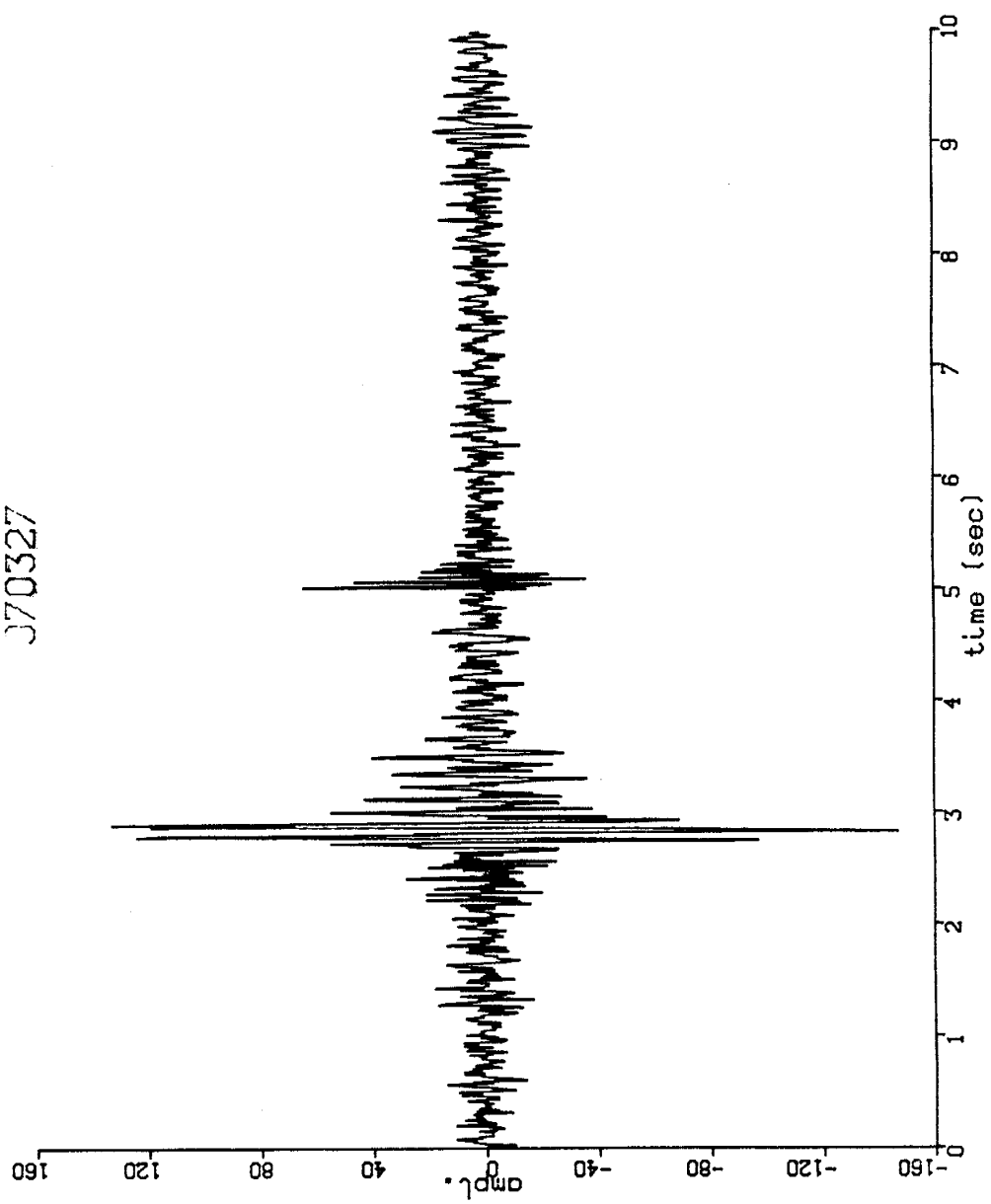


Figure I-4. May 30, 1977 070327 event.

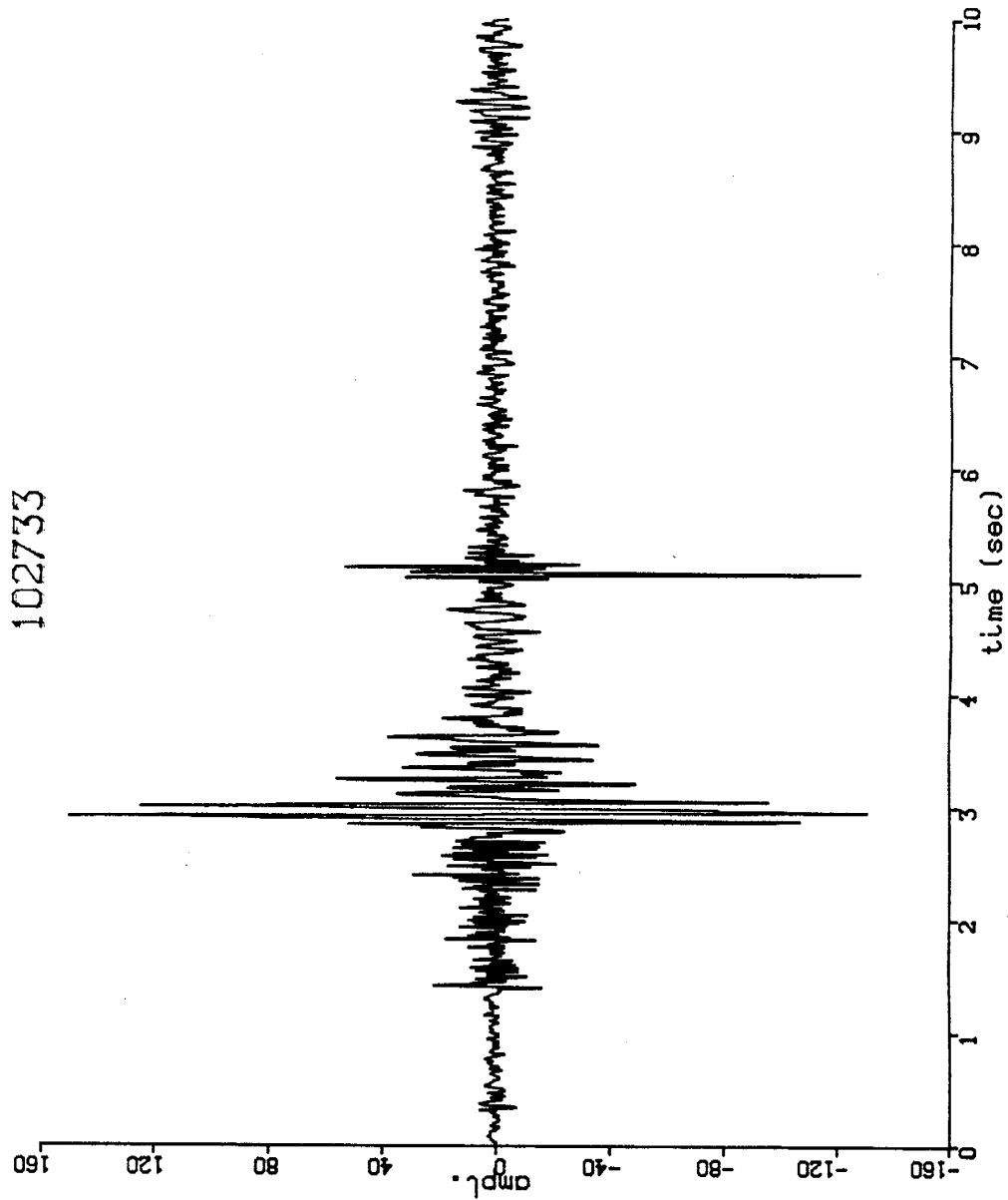


Figure I-5. May 30, 1977 102733 event.

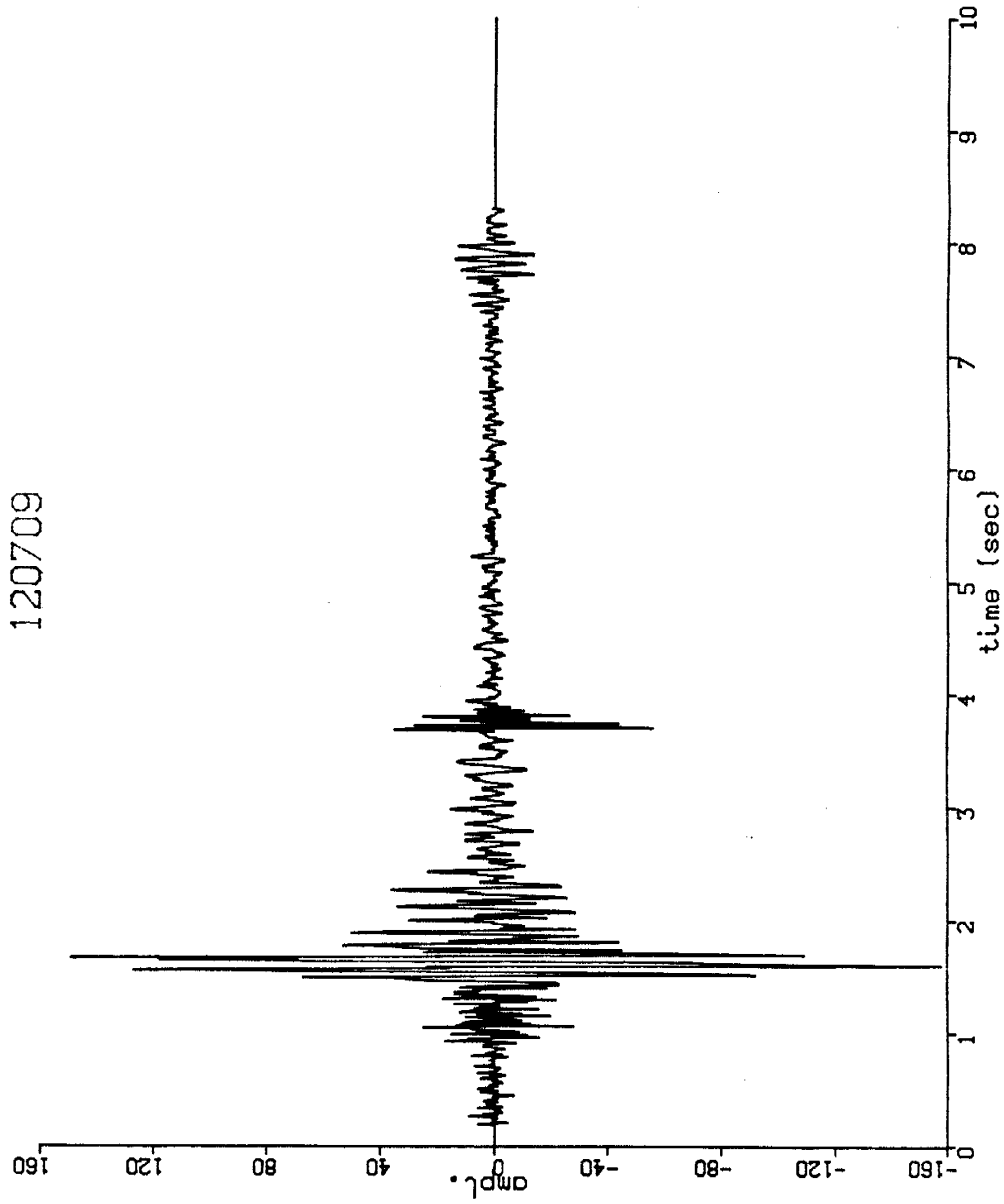


Figure I-6. June 2, 1977 120709 event.

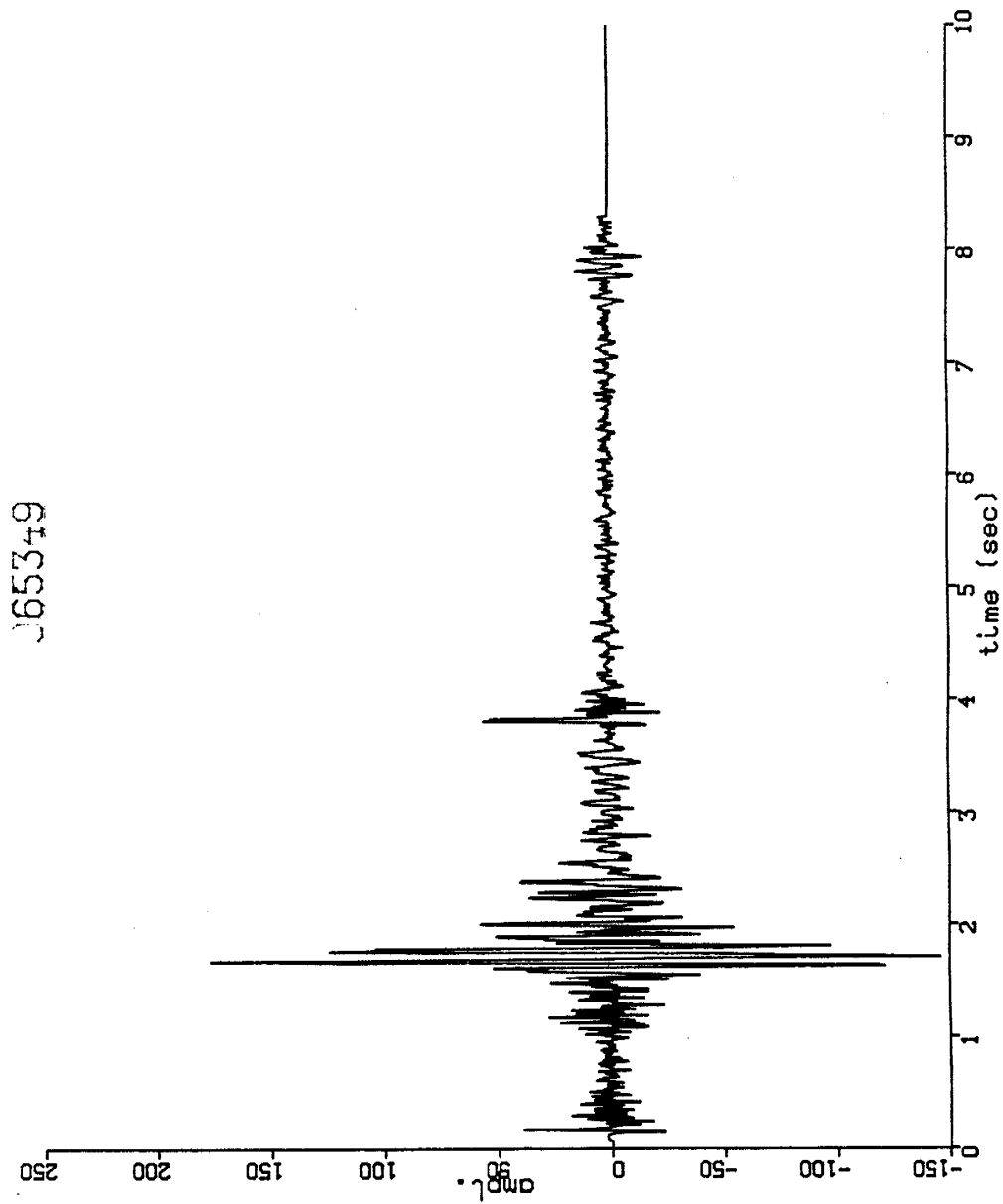


Figure I-7. June 2, 1977 065349 event.

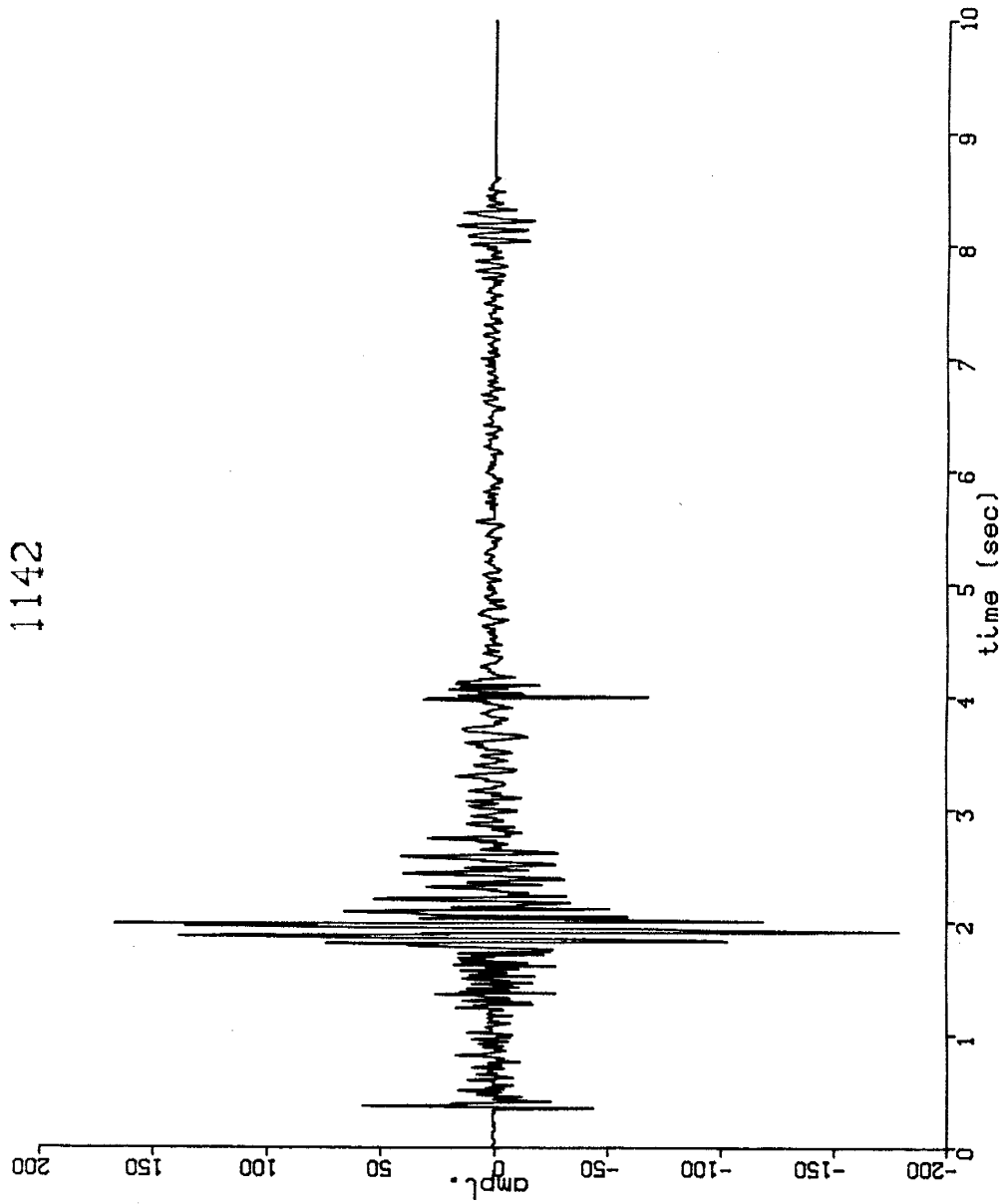


Figure I-8. June 2, 1977 114205 event.

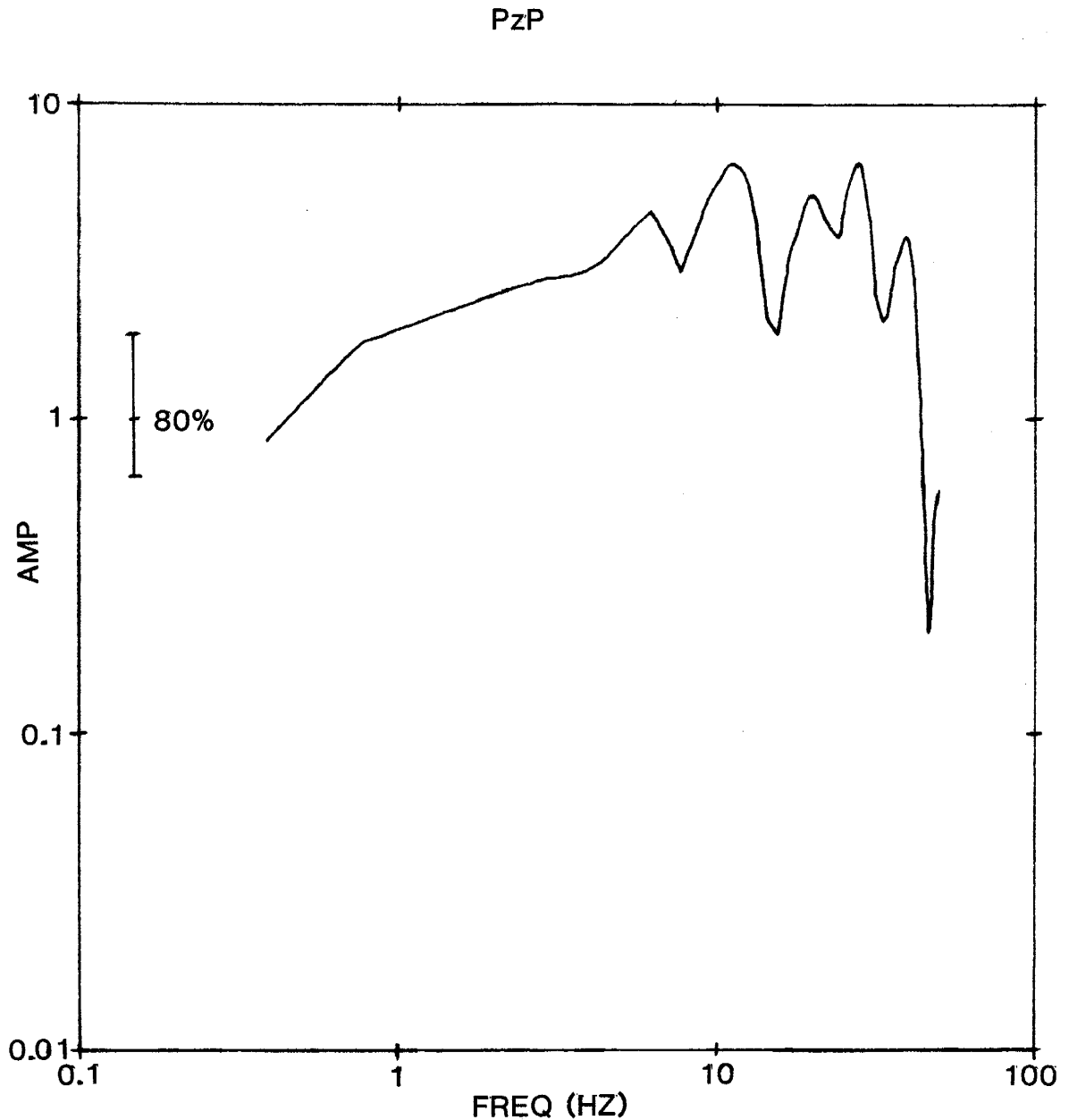


Figure I-9. Normalized spectra of P_zP phase, 114205 event. A three point moving average filter has been applied to the spectrum. Application of the spectral filter allows the calculation of confidence limits for an individual spectrum. The 80% confidence interval is indicated.

Appendix II: Instrument Response

APPENDIX II

Explicit removal of instrumental effects from the observed data was not necessary in this study. The upper crustal response function (the stacked direct P-phase) contains the instrumental response over the same bandwidth as the observed P_zP phase. Hence, convolving the upper crustal response function with the primitive synthetic insures that the synthetic P_zP phase contains the same instrumental effects as the observed P_zP . The only assumption required is that the process is linear, i.e. that the result is insensitive to the order in which convolution occurs. In addition, the process of convolution is a more stable numerical procedure than is deconvolution.

Earlier studies utilizing microearthquake data [Carpenter, 1984] also reported calibration results for DR-100 instruments. These calibration results provided the discrete amplitude response of only the DR-100 system, no seismometer response was included. To verify that the entire system (seismometer, amplifier and recording equipment) would provide adequate response across the bandwidth indicated by the modelling, a supplemental calibration was conducted. The method and results are outlined below.

The most common method of seismograph calibration is the input of a series of known, discrete frequencies into the calibration input of the seismograph. The result is the characteristic amplitude response of the system. The ideal calibration of a seismograph system however, is an impulse response test. The obvious advantage is the recovery of both amplitude and phase information over a broad spectral band. A variant on the impulse response was determined in this

study.

To minimize environmental noise the test was conducted in an abandoned mine. The seismometer (a Mark Products L4-C) was isolated from both the individual conducting the test and the rest of the test equipment. A high quality Hewlett-Packard function generator was used to generate the input pulses. An oscilloscope was used to monitor the signals at several test points during the calibration.

After a number of trials an optimal arrangement was determined. The DR-100 was operated in trigger mode as it would be during typical field deployment. It soon became evident that the input signal needed to be a step rather than a true impulse. In order to trigger on a sharp, narrow impulse, the input voltage had to be large enough that it became questionable whether or not the seismometer was being overdriven. All components were well-behaved when a step function of lower amplitude was applied. During final tests a known voltage step was applied to the calibration coil of the seismometer. Upon application of the voltage step, the unit triggered and recorded the signal. This procedure was repeated a number of times for each filter setting.

Prior to conducting the final calibration, several tests were conducted where the input voltage into the seismometer and filters was monitored to insure that the signal levels were within the specified limits of the instruments. In addition several pulses were recorded and played back to determine the final recorded signal levels and signal to noise ratio. In general, the individual traces had amplitudes of about 1500 digital counts (out of a maximum of 2047) and a good signal to noise ratio.

Because digital sampling can produce differences in the response spectra [Carpenter, 1984] a number of recorded pulses for each filter setting were stacked. Windows containing the pulses were aligned prior to stacking based upon the maximum value of the cross correlation function. This scheme is similar to stacking traces in refraction profiling and will reduce the effect of background noise on the final calibration trace. After stacking the signal to noise ratio was greater than 100 to 1 for most traces.

Since the impulse response was the desired property of the instruments, the final stage of processing was calculating the derivative of the step response. This was done using a conventional centered difference approximation.

The results of the calibration tests proved to be generally consistent with those of Carpenter [1984], indicating a small effect due to the seismometer. The present results indicate the observed P_zP phases were well within the operational bandwidth of the instrumentation. The results of the calibration tests (temporal impulse response and amplitude spectra) are presented in the remainder of this appendix.

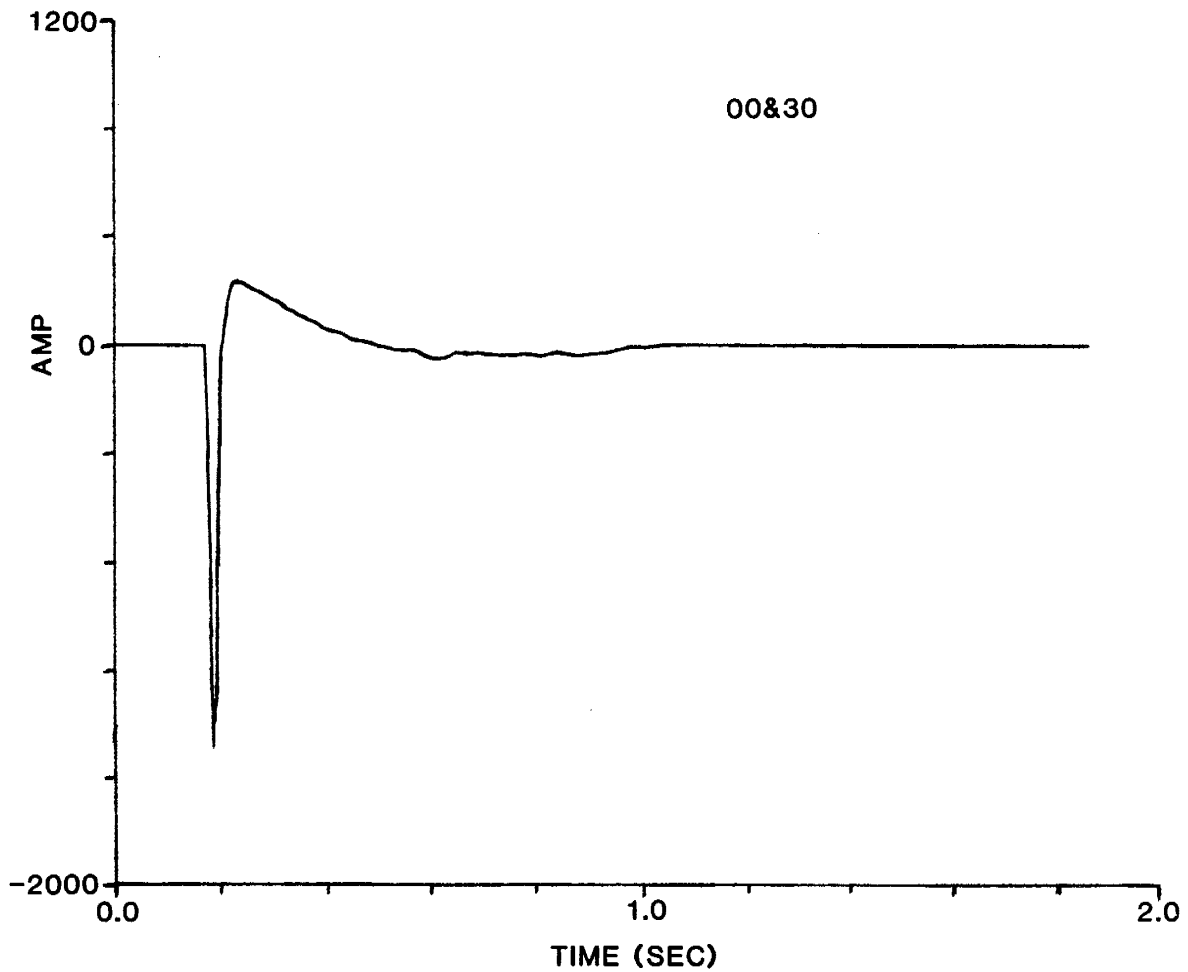


Figure II-1. Time domain impulse response of DR-100A digital recording system with L4-C seismometer. Filter setting of out and 30. A sampling rate of 200 samples/second was used for all calibration tests.

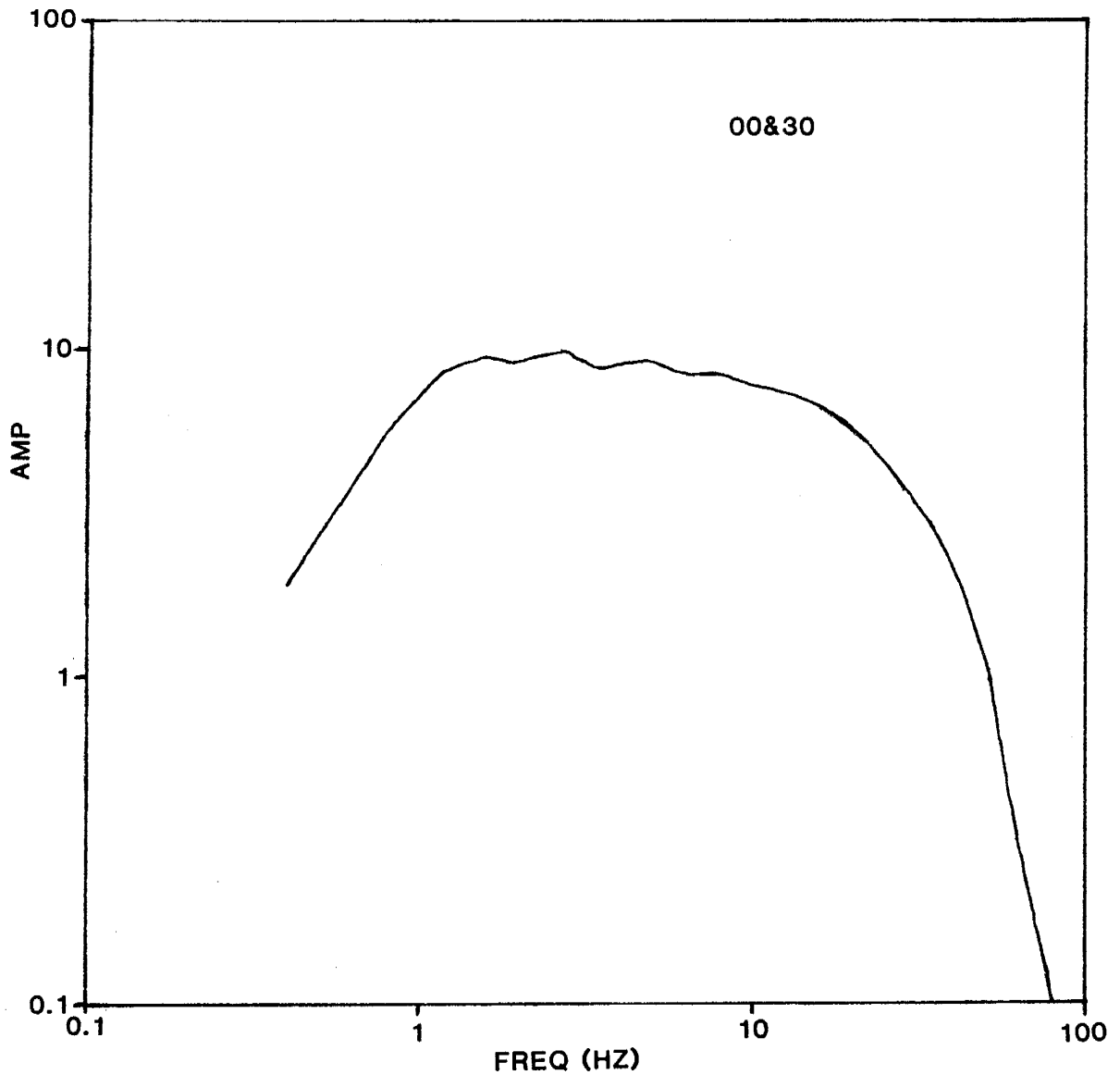


Figure II-2. Frequency domain impulse response of DR-100A digital recording system with L4-C seismometer. Filter settings of 00 and 30.

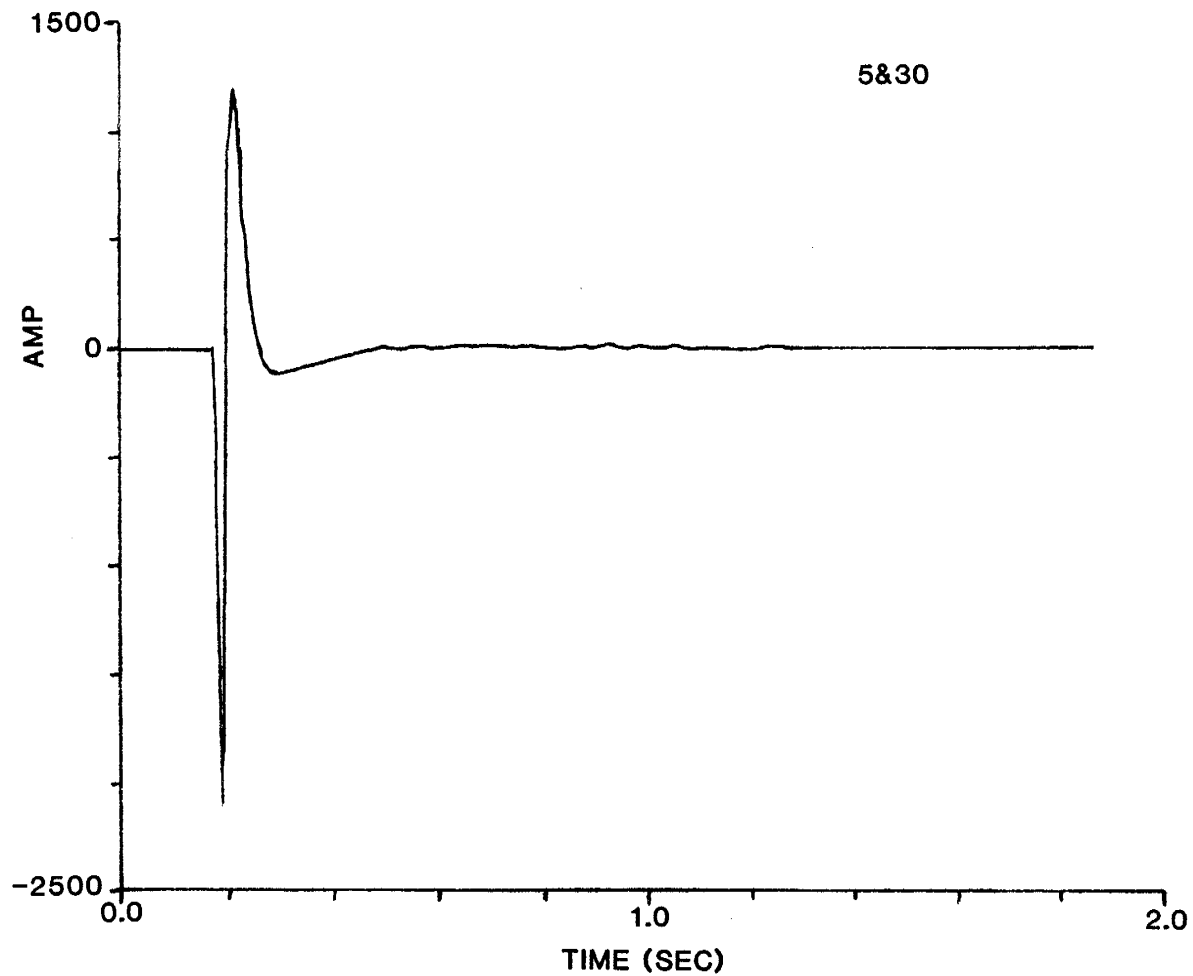


Figure II-3. Time domain impulse response of DR-100A digital recording system with L-4C seismometer. Filter settings of 5 and 30.

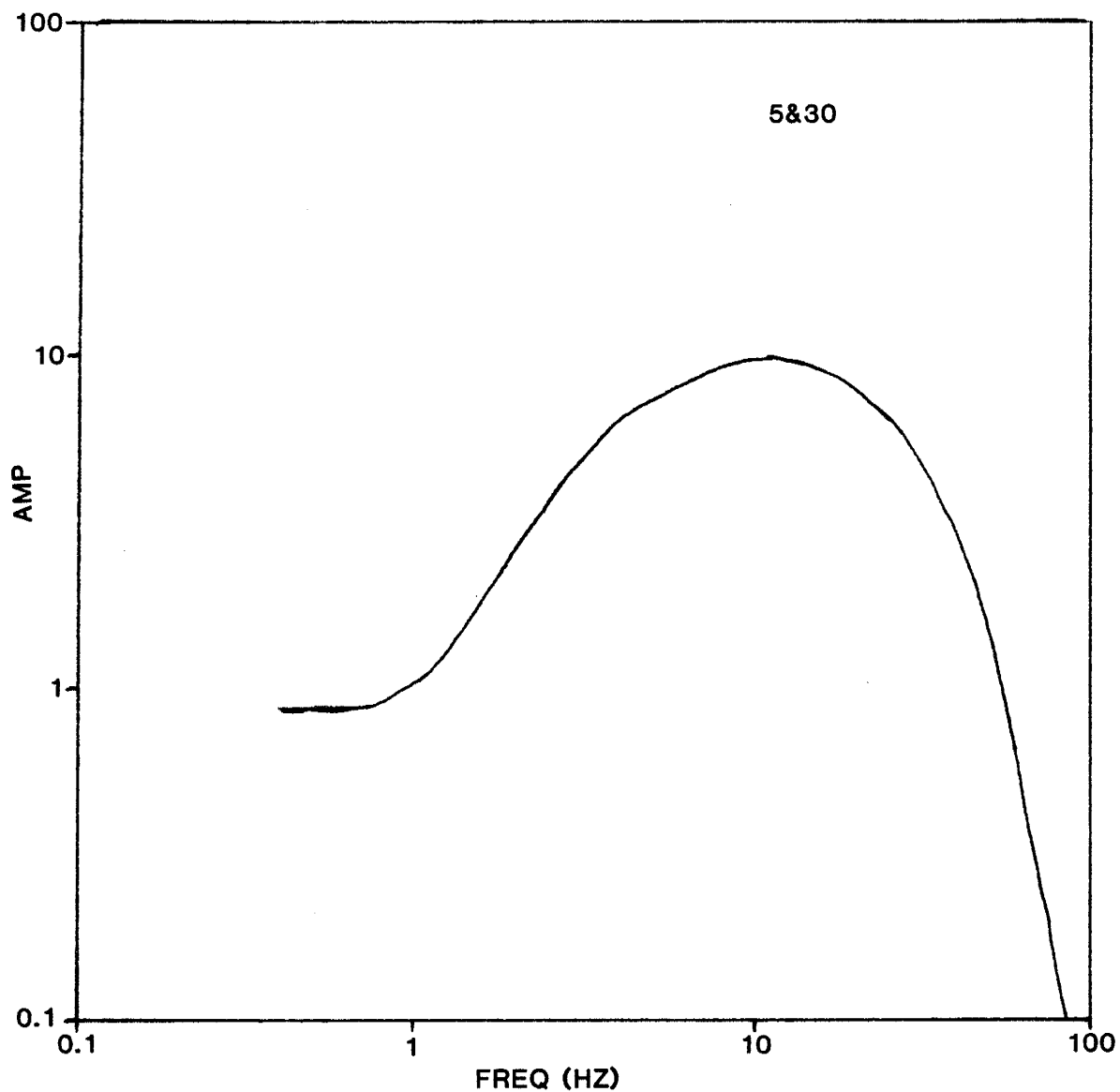


Figure II-4. Frequency domain impulse response of DR-100A digital recording system with L4-C seismometer. Filter settings of 5 and 30.

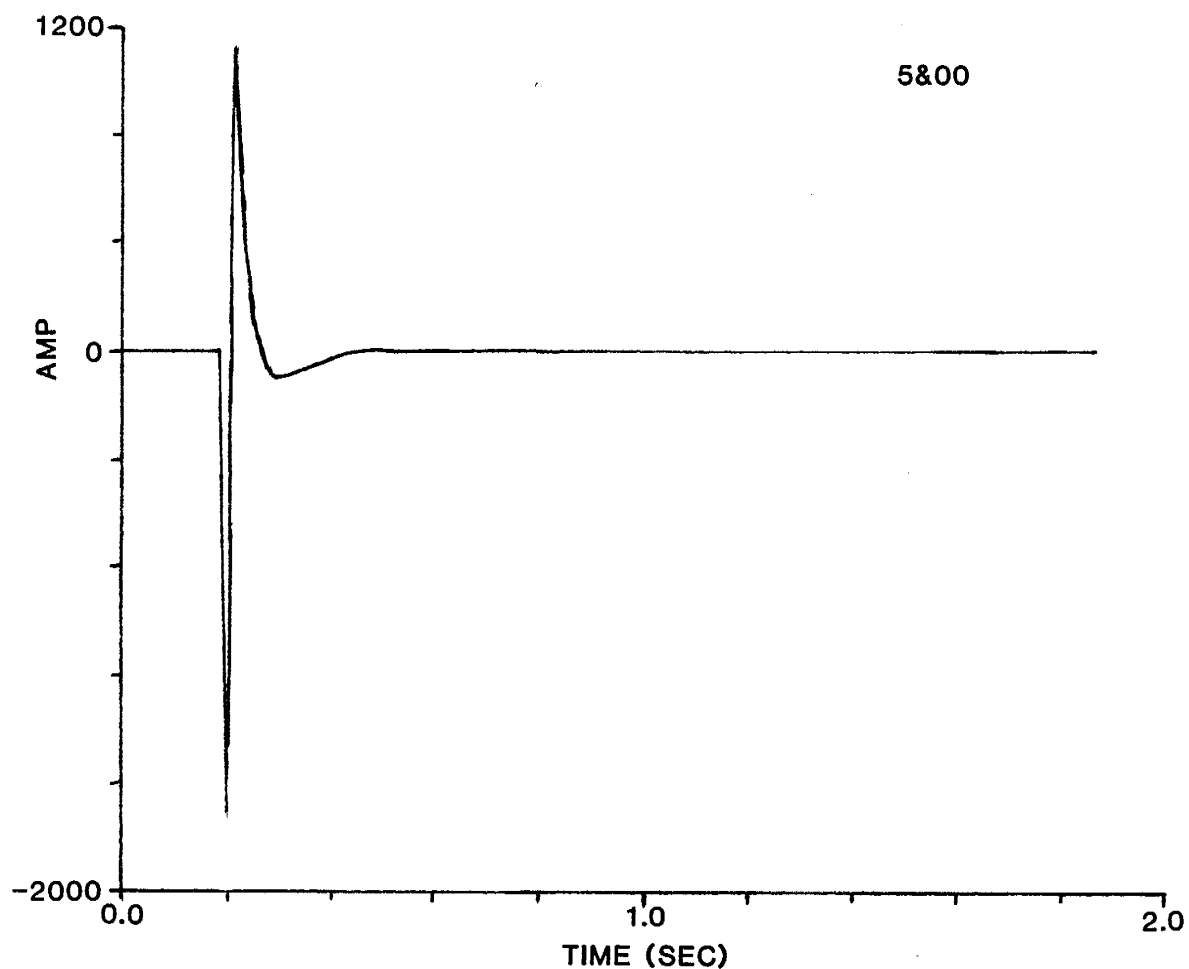


Figure II-5. Time domain impulse response of DR-100A digital recording system with L4-C seismometer. Filter settings of 5 and out.

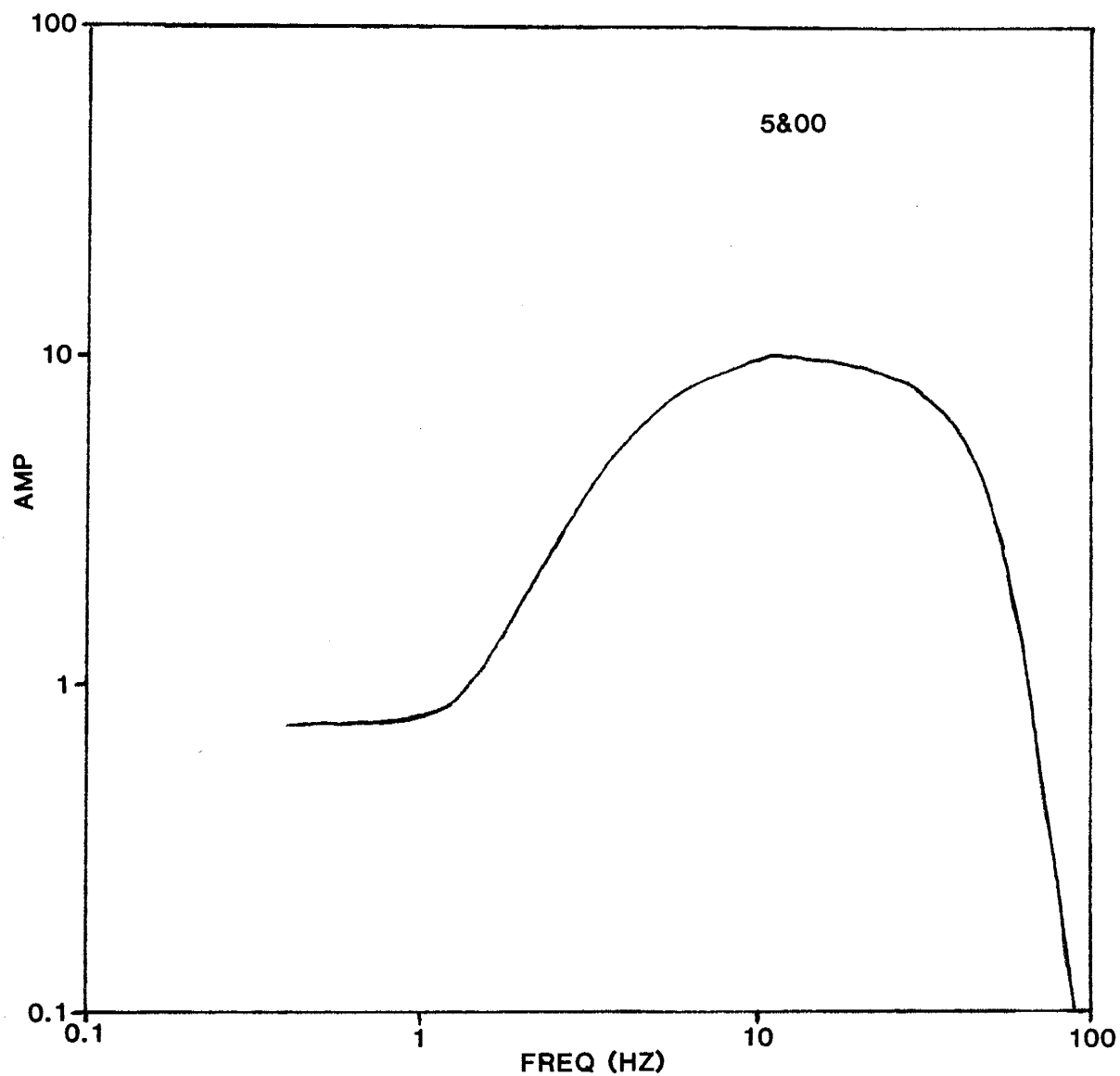


Figure II-6. Frequency domain impulse response of DR-100A digital recording system with L4-C seismometer. Filter settings of 5 and out.

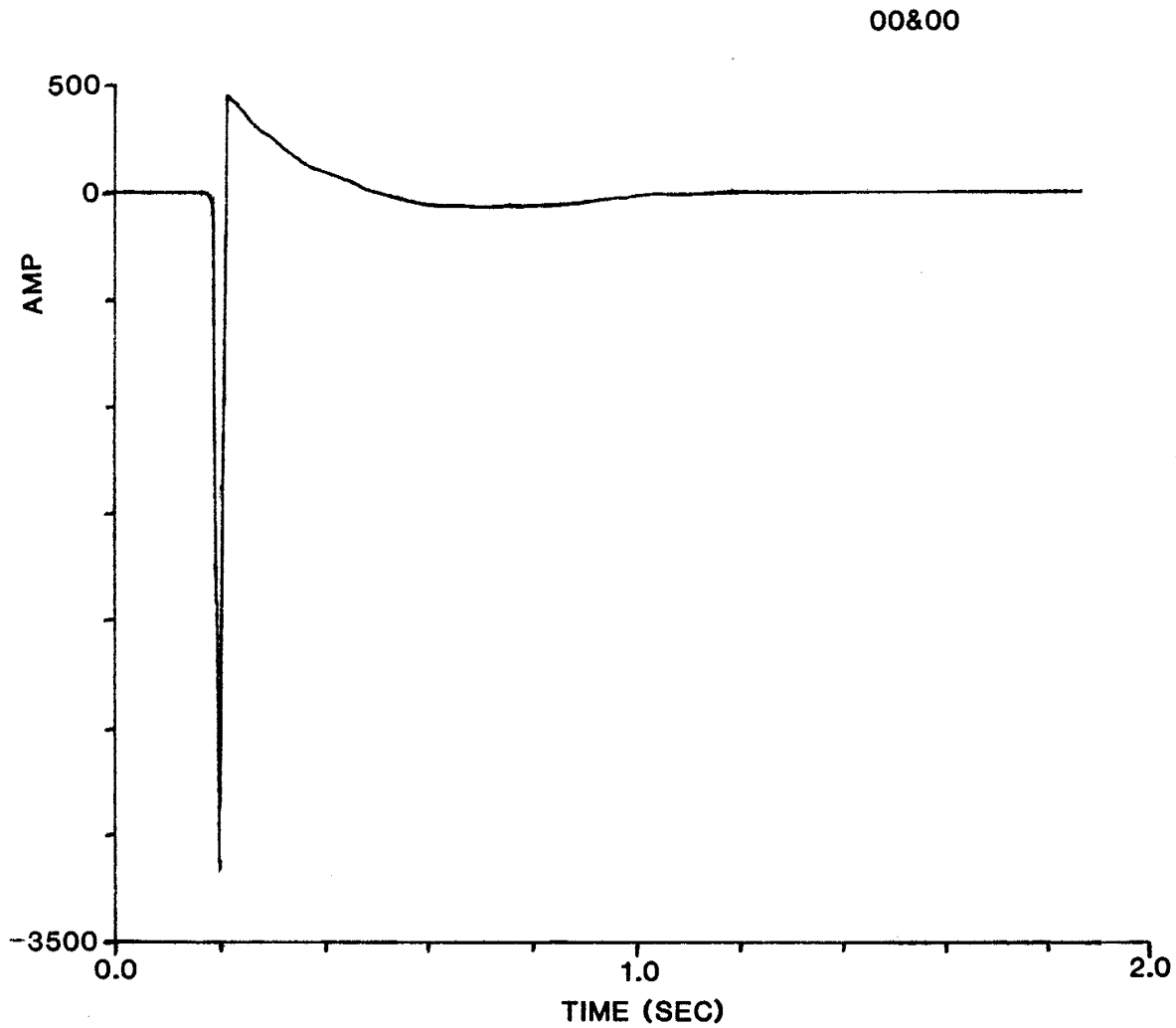


Figure II-7. Time domain impulse response of DR-100A digital recording system with L4-C seismometer. Filter settings of out and out.

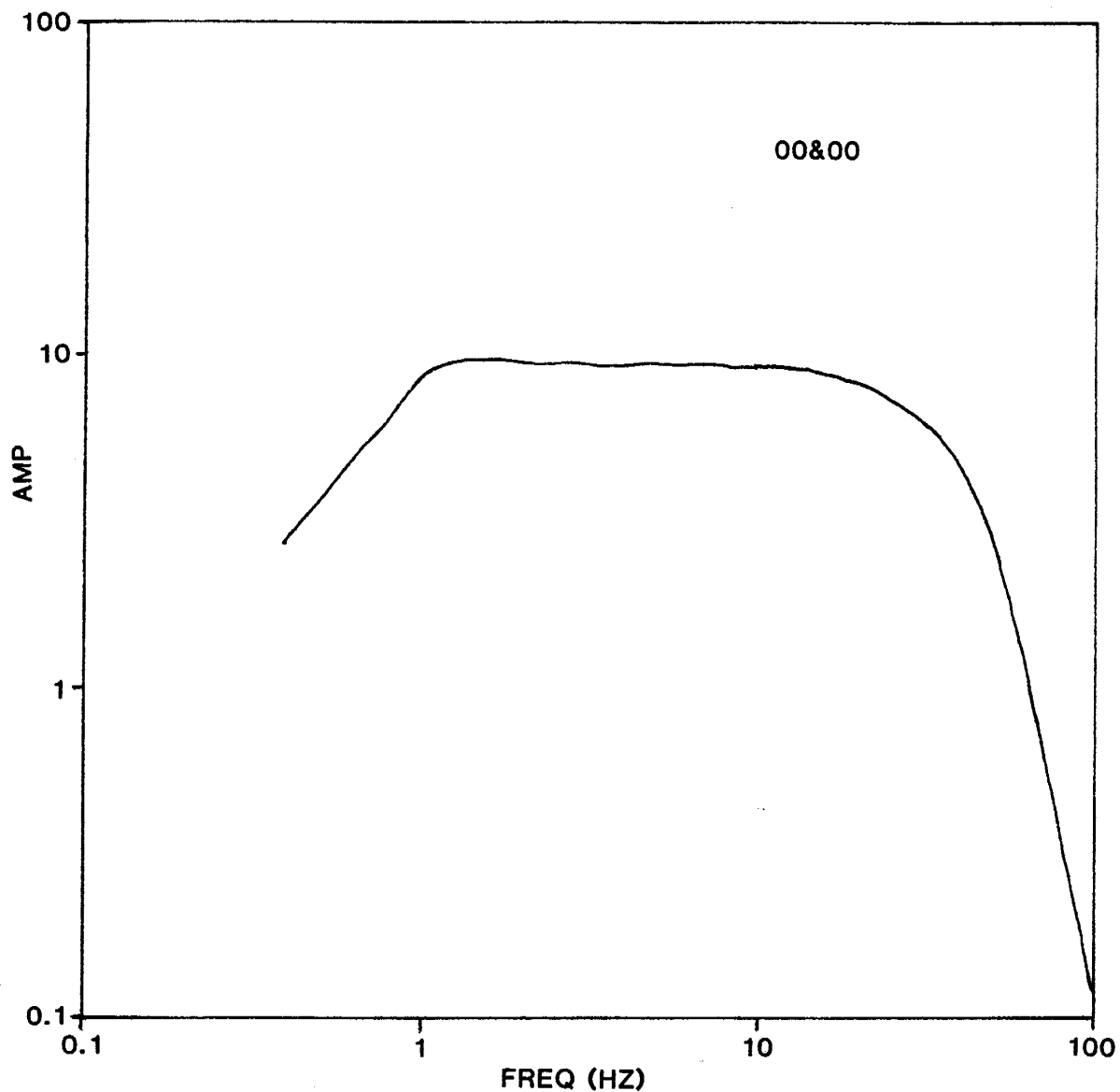


Figure II-8. Frequency domain impulse response of DR-100A digital recording system with L4-C seismometer. Filter settings of out and out.

Appendix III: S Phase Reflection Coefficient

APPENDIX III

Consider the case of an incident SV wave upon a velocity discontinuity within the earth. In the general case, the ratios of reflected, transmitted and incident waves may be found from the Zoeppritz relations [Zoeppritz, 1919]. A summary of the relations may be found in Richter [1958]. For a horizontal interface the equations simplify to:

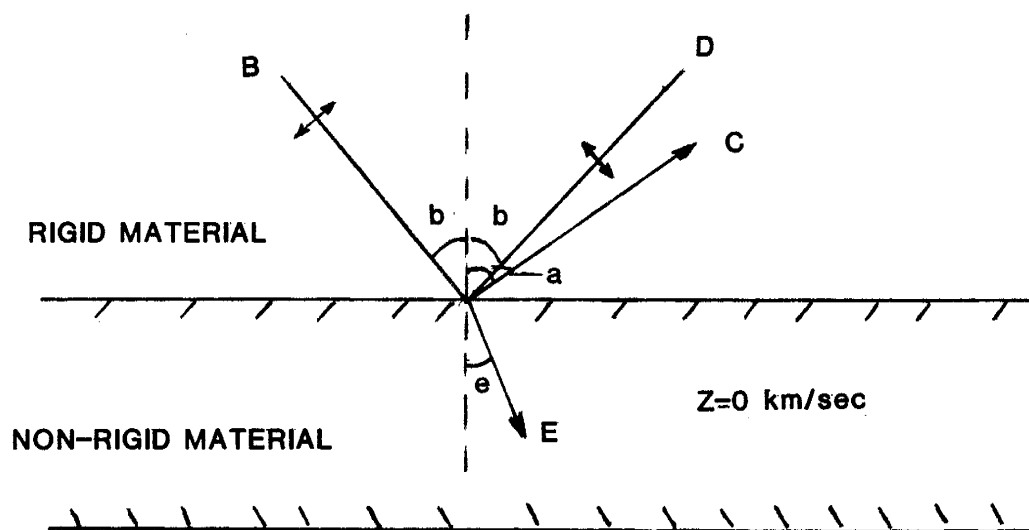
$$\begin{aligned}(B+D)\sin b + C\cos a - E\cos e - F\sin f &= 0 \\(B-D)\cos b + C\sin a + E\sin e - F\cos f &= 0 \\(B+D)\cos 2b - C\left(\frac{U}{V}\right)\sin 2a + EK\left(\frac{Z^2}{VY}\right)\sin 2e - FK\left(\frac{Z}{V}\right)\cos 2f &= 0 \\-(B-D)\sin 2b + C\left(\frac{U}{V}\right)\cos 2a + EK\left(\frac{Y}{V}\right)\cos 2f + FK\left(\frac{Z}{V}\right)\sin 2f &= 0,\end{aligned}$$

where

$$\begin{aligned}B &= \text{amplitude of incident SV} \\D &= \text{amplitude of reflected SV} \\b &= \text{angle wrt normal of incident SV} \\a &= \text{angle of reflected P wrt normal} \\e &= \text{angle of refracted P wrt normal} \\f &= \text{angle of refracted S wrt normal} \\C &= \text{amplitude of reflected P} \\E &= \text{amplitude of refracted P} \\F &= \text{amplitude of refracted SV} \\U &= \text{velocity of P in incident media} \\V &= \text{velocity of incident SV} \\Y &= \text{velocity of P in lower media} \\Z &= \text{velocity of SV in lower media} \\K &= \text{ratio of densities, } \left(\frac{\rho_2}{\rho_1}\right).\end{aligned}$$

We wish to examine a very specific case, normal incidence of SV energy upon a discontinuity separating rigid and non-rigid materials (see Figure III-1). Under these circumstances, $Z=0$ and no SV energy

Figure III-1: Definition of quantities used in the Zoeppritz relations.



will propagate into the lower material, hence F will also equal zero in the above equations. This results in considerable simplification in the governing equations. For the moment ignore any arguments relating to the angles. Then,

$$(B+D)\sin b + C\cos a - E\cos e = 0 \quad (\text{III-1})$$

$$(B-D)\cos b + C\sin a + E\sin e = 0 \quad (\text{III-2})$$

$$(B+D)\cos 2b - C\left(\frac{U}{V}\right)\sin 2a = 0 \quad (\text{III-3})$$

$$-(B-D)\sin 2b + C\left(\frac{U}{V}\right)\cos 2b = 0. \quad (\text{III-4})$$

Rewriting slightly we have

$$B\sin b + D\sin b + C\cos a - E\cos e = 0 \quad (\text{III-5})$$

$$B\cos b - D\cos b + C\sin a + E\sin e = 0 \quad (\text{III-6})$$

$$B\cos 2b + D\cos 2b - C\left(\frac{U}{V}\right)\cos 2b = 0 \quad (\text{III-7})$$

$$-B\sin 2b - D\sin 2b + C\left(\frac{U}{V}\right)\cos 2b = 0. \quad (\text{III-8})$$

From (III-5) we find:

$$E = \frac{B\sin b + D\sin b + C\cos a}{\cos e}$$

and from (III-7):

$$C = \frac{B\cos 2b + D\cos 2b}{\left(\frac{U}{V}\right)\sin 2a}. \quad (\text{III-9})$$

Substituting (III-9) into (III-8):

$$-B\sin 2b - D\sin 2b + \left(\frac{U}{V}\right)\cos 2b \left(\frac{B\cos 2b + D\cos 2b}{\left(\frac{U}{V}\right)\sin 2a} \right) = 0,$$

$$-B\sin 2b - D\sin 2b + B\frac{\cos^2 2b}{\sin 2a} + D\frac{\cos^2 2b}{\sin 2a} = 0,$$

$$B\frac{\cos^2 2b}{\sin 2a} - \sin 2b = D\sin 2b - \frac{\cos^2 2b}{\sin 2a}.$$

Hence,

$$B = D \left(\frac{\left(\frac{-\cos^2 2b}{\sin 2a} \right) + \sin 2b}{\frac{\cos^2 2b}{\sin 2a} - \sin 2b} \right). \quad (\text{III-10})$$

Since the angles and velocities conform to Snell's Law,

$$\frac{\sin a}{U} = \frac{\sin b}{V},$$

if we have normal incidence (and U and $V \neq 0$) then $b=0, \sin b = 0$, and so $\sin a$ must equal zero. Hence the right side of (III-10) becomes undefined (the indeterminate form $\frac{\infty}{\infty}$). Hence to evaluate the behaviour of B in terms of D for b near zero, we apply L'Hopital's rule to (III-10). First, since

$$\frac{\sin a}{U} = \frac{\sin b}{V}, \quad \sin 2a = \left(\frac{U}{V}\right)\sin 2b$$

for small values of b , then

$$B = D \left(\frac{\left(\frac{-\cos^2 2b}{U}\right) + \sin 2b}{\left(\frac{U}{V}\right)\sin 2b} \right). \quad (\text{III-11})$$

Define:

$$H = \left(\frac{-\cos^2 2b}{U}\right) + \sin 2b$$

and

$$J = \left(\frac{\cos^2 2b}{U}\right) - \sin 2b.$$

Then,

$$\begin{aligned} \left(\frac{dH}{d2b}\right) &= \left(\frac{U}{V}\right)\sin 2b(-2\cos 2b \sin 2b) + (\cos^2 2b)\left(\frac{U}{V}\right)\cos 2b + \cos 2b, \\ \left(\frac{dH}{d2b}\right) &= \cos 2b \left(\left(\frac{U}{V}\right)(-2\sin^2 2b + \cos^2 2b) + 1\right) \end{aligned}$$

and,

$$\left(\frac{dJ}{d2b}\right) = \frac{U}{V}\sin 2b(2\cos 2b \sin 2b) - \cos^2 2b \frac{U}{V}\cos 2b - \cos 2b,$$

$$\left(\frac{dJ}{d2b}\right) = \cos 2b \left(\frac{U}{V}(2\sin^2 2b - \cos^2 2b) - 1\right).$$

It is now possible to evaluate the behavior of (III-11) when $2b$ is near zero. For normal incidence:

$$B = \lim_{2b \rightarrow 0} D \left(\frac{\frac{U}{V}(-2\sin^2 2b + \cos^2 2b) + 1}{\frac{U}{V}(2\sin^2 2b - \cos^2 2b) - 1} \right). \quad (\text{III-12})$$

Since

$$\lim_{2b \rightarrow 0} \sin^2 2b = 0$$

and

$$\lim_{2b \rightarrow 0} \cos^2 2b = 1,$$

equation (III-12) becomes

$$B = D \left(\frac{\frac{U}{V} + 1}{-\frac{U}{V} - 1} \right) \quad (\text{III-13})$$

so

$$B = -D.$$

or

$$\frac{B}{D} = -1 \quad (\text{III-14})$$

So, for normal incidence of SV energy upon a rigid-non-rigid boundary, the reflection coefficient is shown to be negative unity.

Appendix IV: Thermal Lifetime of Magma Body

APPENDIX IV

One interesting question that arises from the inferred thin nature of the mid-crustal magma body is thermal lifetime. In other words, how long will a body as thin as postulated by this study retain the unique seismic characteristics that enable it to be easily identified?

Following Turcotte and Schubert [1980], the time for a sill of thickness $2b$ to solidify can be shown to be

$$t_s = \frac{b^2}{4\kappa\lambda_2^2} \quad (\text{IV-1})$$

where

$$\begin{aligned} t_s &= \text{solidification time,} \\ \kappa &= \text{thermal diffusivity,} \\ 2b &= \text{sill thickness,} \end{aligned}$$

and

$$\lambda_2 = \text{a decay parameter.}$$

The decay parameter, λ_2 , is evaluated from

$$L \frac{\sqrt{\pi}}{c} (T_m - T_o) = \frac{e^{\lambda_2^2}}{\lambda_2 (1 + \operatorname{erf} \lambda_2)} \quad (\text{IV-2})$$

where

$$\begin{aligned} L &= \text{latent heat,} \\ T_m &= \text{melt temperature,} \\ T_o &= \text{wall rock temperature, and} \\ c &= \text{specific heat.} \end{aligned}$$

Since the purpose of this exercise is merely to estimate the ther-

mal lifetime of the sill, the value of λ_2 was estimated graphically from Turcotte and Schubert (their figure 4-21). To evaluate equation (IV-2), define

$$\begin{aligned} L &= 320\text{kJ/kg}, \\ T_m - T_o &= 600^\circ \text{K}, \\ c &= 1.2\text{kJ/kg}^\circ\text{K}. \end{aligned}$$

Using these values, λ_2 is found to be 0.5. Next, define

$$\kappa = 0.7 \frac{\text{mm}^2}{\text{sec}},$$

and,

$$b = 37.5\text{m},$$

(i.e. a sill thickness of 75 m), then from equation (IV-1) the solidification time is found to be 63.7 years. If the sill thickness is increased to 150 m then the solidification time becomes 255 years.

Although this may be viewed only as an estimate of the thermal life of the body, it does point out that the body will possess the unique seismic characteristics for only a short period of time after emplacement. This suggests that resupply of the sill may be a frequent and ongoing phenomena.

Appendix V: Computer Programs

APPENDIX V

This appendix presents listings of the programs used to analyze and plot digital seismograms and construct the synthetic seismograms used in this study. Descriptions of the programs are contained in the comments of each program. Most of the programs were written by the author, a few are from Robinson [1980].

```

c cccorrl.for
c simple driver program to compute cross-correlation of two
c time series. an equal sampling interval is assumed. note-
c the first series lags the second, i.e. it is progressively
c shifted towards the left. the input files are input by the
c user and are of standard format:
c      (i,f)
c      npts,delt
c      x(i)
c      .
c      .
c      x(npts)
c      eof
c the output is normalized and output to cros.dat, format
c is the same as the input files, compatible with plot.for etc.
c printed output is also provided.
c symbols used: x(i)=first input wavelet
c               nptsx=length of x
c               y(i)=second input wavelet
c               nptsy=length of y
c               delt=sampling interval
c               nlags=length of c
c               c(i)=normalized cross correlation values
c
c written by jon ake, 7/86
c modified to be sun compatible, 7/87 jpa
c
c dimension x(2049),y(2049),xy(4097),c(4097)
c character*10 vfnninx,vfnniny
c
c format (' ')
c format (' specify first input file: ',5)
c format (' specify second input file: ',5)
c format (a10)
c format (i7,f13.5)
c format (f13.4)
c format (' input files: (x,y) '/')
c format (' n=',i5/)
c format (5x,f13.4,5x,f13.4)
c format (/ ' normalized cross-correlation function' /)
c format (4x,i5,3x,f13.4)
c format (5x, ' cor-max:',f9.5,3x, ' cor-min',f9.5/)
c format (/5x, ' normalization factor: ',f13.4/)
c format (/3x, ' cross correlation of ',a10,' and ',a10/)
c
c specify input files
c
c write (6,2)
c read (5,4) vfnninx
c write (6,3)
c read (5,4) vfnniny
c
c open (unit=3,access='sequential',file='prfil.d')
c open (unit=21,access='sequential',file=vfnninx)
c open (unit=22,access='sequential',file=vfnniny)
c open (unit=23,access='sequential',file='cros.dat')
c
c input data

```

```

20 read (21,5) nptsx,delt
   do 20 i=1,nptsx
     read (21,6) x(i)
     read (22,5) nptsy,delt
     do 22 i=1,nptsy
       read (22,6) y(i)
       set up parameters for correlation
       c
       if series are of unequal length, add zeroes to
       the rear of the shorter series
       c
       zero=0.0
       if (nptsx.eq.nptsy) go to 94
       if (nptsx.gt.nptsy) go to 92
       do 91 i=nptsx+1,nptsy
         x(i)=zero
         n=nptsy
         go to 95
       do 93 i=nptsy+1,nptsx
         y(i)=zero
         n=nptsx
         go to 95
       n=nptsx
       continue
       nlags=(2*n)-1
       compute cross-correlation
       c
       compute negative lag values
       c
       do 50 j=1,n
         term=zero
         it=n-j+1
         do 52 i=1,j
           term1=x(i)*y(it)
           term=term+term1
           it=it+1
           c(j)=term
         compute positive lag values
         c
         do 54 j=2,n
           term=zero
           it=1
           do 55 i=j,n
             term1=x(i)*y(it)
             term=term+term1
             it=it+1
             c(n+j-l)=term
           compute normalization factor for cross correlation
           c
           smul=zero
           smu2=zero
           do 36 i=1,n
             sum1=x(i)*x(i)
             sum2=y(i)*y(i)
             smul=smul+sum1
             smu2=smu2+sum2
           36

```



```

c
dimension tr(40),r(40),srce(4097),amp(4097),z(40),rho(40)
dimension f(40),t(40),it(40),temp(4097),v(40),q(40),d(40,40)
dimension temp(4097),cal(257),zcon(4097),send(4097)
dimension qbar(40),rbar(40),vbar(40),coeff(513)
complex source(4097)
character*10 vfnin,calin,resp,headr,respp,pfn

format (' ')
format (3x,'imax=',i5)
format (7x,'primitive')
format (17,i7,i7)
format (17,f13.5)
format (/15x,'synthetic analysis'//)
format (i5)
format (i5,i5)
format (f13.4,f13.4,f13.4,f13.4)
format (f13.4)
format (f13.4,f13.4,f13.4,f13.4/)
format (/3x,'scaling factors'//)
format (/3x,'time shifting factors'//)
format (3x,'i=',i5,'r(i)='f9.5',tr(i)='f9.5)
format (/3x,'reflection and transmission coefficients'//)
format (/5x,'integer time shifting factors'//)
format (f13.4,f13.4)
format (/3x,'filtered amplitudes'//)
format (/5x,'amplitude spectra'//)
format (' calculate spectra of synthetic? y or n')
format (a2)
format (3x,'i=',i5,'it=',i5,'t(i)='f10.5)
format (/5x,'synthetic amplitudes'//)
format (3x,i5,2x,f10.5)
format (/3x,'delta-f='f7.3//)
format (5x,'input instrument response data file:',$)
format (2x,'input data'//)
format (2x,'velocity',5x,'density',2x,'thickness',6x,'q'//)
format (3x,f8.2,4x,f7.2,4x,f9.2,4x,f7.2)
format (' input header designation (10 chars max): ', $)
format (3x,'model: ',a10//)
format (' print primitive synthetic? (y or n): ', $)
format (' designate input data file: ', $)
format (a10)

specify data file
write (6,1063)
read (5,1064) vfnin

open (unit=21,access='sequential',file=vfnin)
open (unit=22,access='sequential',file='plot.dat')
open (unit=23,access='sequential',file='fit.dat')
open (unit=24,access='sequential',file='ploff.dat')
open (unit=26,access='sequential',file='scale.dat')

set up line printer file

open (unit=3,access='sequential',file='prfil.d')

input data

```

```

read (21,4) n,ns,m
do 30 i=1,n
  read (21,10) deltat
do 32 i=1,ns
  read (21,9) v(i),rho(i),z(i),q(i)
  read (21,10) srce(i)
provide documentation
write (3,6)
write (6,110)
read (5,1064) headr
write (3,111) headr
write (3,106)
write (3,107)
do 108 i=1,n
  write (3,109) v(i),rho(i),z(i),q(i)
  write (3,15)
calculate reflection and transmission coefficients
ni=n-1
do 34 i=1,ni
  r(i)=(v(i+1)*rho(i+1)-v(i)*rho(i))/(v(i+1)*rho(i+1)+v(i)*rho(i))
  tr(i)=1.-r(i)
  write (3,14) i,r(i),tr(i)
input number of phases to be considered
if (n.eq.3) jk=3
if (n.eq.4) jk=7
if (n.eq.5) jk=13
if (n.eq.6) jk=21
compute scaling factors
s=tr(1)*(1+r(1))
f(1)=r(1)
f(2)=s*r(2)
f(3)=s*r(2)*(-1.*r(1)*r(2))
if (n.eq.3) go to 24
sl=tr(2)*r(3)
f(4)=s*sl*(1.+r(2))
f(5)=s*sl*(-1.*r(2)*r(3)*(-1.*r(1))*r(2))
f(6)=s*sl*r(2)*(-1.*r(1))*r(2))
f(7)=f(6)
if (n.eq.4) go to 24
s2=tr(2)*tr(3)*r(4)
s3=(1.+r(3))*(1.+r(2))
f(8)=s*s2*s3
f(9)=s*s2*s3*(-1.*r(3))*r(4)
f(10)=s*s2*s3*r(2)*(-1.*r(1))
f(11)=f(10)
f(12)=s*s2*s3*(1.+r(2))*(-1.*r(1))*tr(2)*tr(3)
f(13)=f(12)
if (n.eq.5) go to 24
s4=tr(2)*tr(3)*tr(4)*r(5)*(1.+r(4))
f(14)=s*s3*s4
f(15)=s*s3*s4*(-1.*r(4))*r(5)
f(16)=s*s3*s4*r(2)*(-1.*r(1))

```

```

f(17)=f(16)
f(18)=s*s1*s3*s4*(1.+r(2))*(-1.*r(1))
f(19)=f(18)
f(20)=s*s2*s3*s4*(1.+r(3))*(1.+r(2))*(-1.*r(1))
f(21)=f(20)
write (3,12)
do 22 i=1,jk
  write (3,10) f(1)
22
c
c
c
c
c
compute time shifting factors
shift relative to primary reflection
s4=2.*z(2)/v(2)
t(1)=0.0
t(2)=s4
t(3)=2.*s4
if (n.eq.3) go to 35
s5=2.*z(3)/v(3)
t(4)=s4+s5
t(5)=s4+2.*s5
t(6)=2.*s4+s5
t(7)=t(6)
if (n.eq.4) go to 35
s6=2.*z(4)/v(4)
t(8)=s4+s5+s6
t(9)=s4+s5+2.*s6
t(10)=2.*s4+s5+s6
t(11)=t(10)
t(12)=2.*s4+2.*s5+s6
t(13)=t(12)
if (n.eq.5) go to 35
s7=2.*z(5)/v(5)
t(14)=s4+s5+s6+s7
t(15)=t(14)+s7
t(16)=t(14)+s4
t(17)=t(16)
t(18)=t(14)+s4+s5
t(19)=t(18)
t(20)=t(18)+s6
t(21)=t(20)
write (3,13)
do 37 i=1,jk
  write (3,10) t(i)
37
c
c
c
convert to integers
write (3,16)
do 46 i=1,jk
  tt=t(i)/deltat
  write (3,10) tt
  it(i)=int(tt)
  ft=float(it(i))
  if (tt-ft.gt.0.5000) it(i)=it(i)+1
  write (3,10) i, it(i), t(i)
46
c
no=256
imax=1
do 47 i=1,jk
  if (it(i).gt.imax) imax=it(i)
  write (3,1)
47

```

```

write (3,2) imax
imax2=imax+10
imax=imax+no+10
write (3,2) imax
write (3,1)
leng=2049
c
c
c
calculate weighted coefficients for attenuation
distance
nhalf=64
do 69 i=us+1,nhalf
  srce(i)=0.0
  ns=nhalf
  rbar(1)=2.*(z(1))
  rbar(2)=rbar(1)+2.*z(2)
  rbar(3)=rbar(2)+2.*z(2)
  if (n.eq.3) go to 28
  rbar(4)=rbar(2)+2.*z(3)
  rbar(5)=rbar(4)+2.*z(3)
  rbar(6)=rbar(4)+2.*z(2)
  rbar(7)=rbar(6)
  if (n.eq.4) go to 28
  rbar(8)=rbar(4)+2.*z(4)
  rbar(9)=rbar(8)+2.*z(4)
  rbar(10)=rbar(8)+2.*z(2)
  rbar(11)=rbar(10)
  rbar(12)=rbar(10)+2.*z(3)
  rbar(13)=rbar(12)
  if (n.eq.5) go to 28
  rbar(14)=rbar(8)+2.*z(5)
  rbar(15)=rbar(14)+2.*z(5)
  rbar(16)=rbar(14)+2.*z(2)
  rbar(17)=rbar(16)
  rbar(18)=rbar(14)+2.*z(2)+2.*z(3)
  rbar(19)=rbar(18)
  rbar(20)=rbar(14)+2.*z(2)+2.*z(3)+2.*z(4)
  rbar(21)=rbar(20)
28
continue
c
c
c
velocity
sl=2.*(v(1)*z(1))
s2=2.*(v(1)*z(1)+v(2)*z(2))
vbar(1)=sl/rbar(1)
vbar(2)=s2/rbar(2)
vbar(3)=(s2+2.*v(2)*z(2))/rbar(3)
if (n.eq.3) go to 29
s3=2.*(v(1)*z(1)+v(2)*z(2)+v(3)*z(3))
vbar(4)=s3/rbar(4)
vbar(5)=(s3+2.*v(3)*z(3))/rbar(5)
vbar(6)=(s3+2.*v(2)*z(2))/rbar(6)
vbar(7)=vbar(6)
if (n.eq.4) go to 29
s4=2.*(v(1)*z(1)+v(2)*z(2)+v(3)*z(3)+v(4)*z(4))
vbar(8)=s4/rbar(8)
vbar(9)=(s4+2.*v(4)*z(4))/rbar(9)
vbar(10)=(s4+2.*v(2)*z(2))/rbar(10)
vbar(11)=vbar(10)
vbar(12)=(s4+2.*v(2)*z(2)+2.*v(3)*z(3))/rbar(12)

```

```

vbar(l3)=vbar(l2)
if (n.eq.5) go to 29
s5=2.*(v(l1)*z(l1)+v(2)*z(2)+v(3)*z(3)+v(4)*z(4)+v(5)*z(5))
vbar(l4)=s5/rbar(l4)
vbar(l5)=(s5+2.*v(5)*z(5))/rbar(l5)
vbar(l6)=(s5+2.*v(2)*z(2))/rbar(l6)
vbar(l7)=vbar(l6)
vbar(l8)=(s5+2.*v(2)*z(2)+2.*v(3)*z(3))/rbar(l8)
vbar(l9)=vbar(l8)
vbar(20)=(s5+2.*v(2)*z(2)+2.*v(3)*z(3)+2.*v(4)*z(4))/rbar(20)
vbar(21)=vbar(20)
continue

29
c
c
c
quality factor
sl=2.*(q(l1)*z(l1))
s2=2.*(q(l1)*z(l1)+q(2)*z(2))
qbar(l1)=sl/rbar(l1)
qbar(l2)=s2/rbar(l2)
qbar(3)=(s2+2.*q(2)*z(2))/rbar(3)
if (n.eq.3) go to 25
s3=2.*(q(l1)*z(l1)+q(2)*z(2)+q(3)*z(3))
qbar(4)=s3/rbar(4)
qbar(5)=(s3+2.*q(3)*z(3))/rbar(5)
qbar(6)=(s3+2.*q(2)*z(2))/rbar(6)
qbar(7)=qbar(6)
if (n.eq.4) go to 25
s4=2.*(q(l1)*z(l1)+q(2)*z(2)+q(3)*z(3)+q(4)*z(4))
qbar(8)=s4/rbar(8)
qbar(9)=(s4+2.*q(4)*z(4))/rbar(9)
qbar(10)=(s4+2.*q(2)*z(2))/rbar(10)
qbar(11)=qbar(10)
qbar(12)=(s4+2.*q(2)*z(2)+2.*q(3)*z(3))/rbar(12)
qbar(13)=qbar(12)
if (n.eq.5) go to 25
s5=2.*(q(l1)*z(l1)+q(2)*z(2)+q(3)*z(3)+q(4)*z(4)+q(5)*z(5))
qbar(14)=s5/rbar(14)
qbar(15)=(s5+2.*q(5)*z(5))/rbar(15)
qbar(16)=(s5+2.*q(2)*z(2))/rbar(16)
qbar(17)=qbar(16)
qbar(18)=(s5+2.*q(2)*z(2)+q(3)*z(3))/rbar(18)
qbar(19)=qbar(18)
qbar(20)=(s5+2.*q(2)*z(2)+2.*q(3)*z(3)+2.*q(4)*z(4))/rbar(20)
qbar(21)=qbar(20)
continue

25
c
c
c
apply q-filter to each pulse and sum waveforms
call zzero (amp, leng)
do 60 j=1,jk
  jv=j
  que=qbar(jv)
  dist=rbar(jv)
  vel=vbar(jv)
  call qfilter (srce,no,ns,deltat,que,dist,vel,n,jv,tamp)
do 50 i=1,no
  temp(i)=tamp(i)*f(j)
do 54 k=1,no
  ik=k-1+it(j)
  amp(ik)=amp(ik)+temp(k)

```

```

continue
60
c
c
print out primitive synthetic
write (6,112)
read (5,72) respp
if (respp.eq.'n'.or.respp.eq.'n') go to 114
write (3,102)
write (3,3)
write (3,103) (i,amp(i),i=1,imax)
n=200
write (23,5) n,deltat
write (23,10) (amp(i),i=1,n)
continue
114
c
c
continue
99
c
c
input instrument and path impulse response to be convolved
with primitive synthetic to produce final seismogram
write (6,104)
read (5,1064) calin
open (unit=25,access='sequential',file=calin)
read (25,5) nc,delt
do 77 i=1,nc
  read (25,10) cal(i)
  write (3,10) cal(i)
77
c
c
c
convolve instrument response with primitive synthetic
do 59 i=1,2000
  send(i)=0.0
  zcon(i)=0.0
  lz=imax+nc-1
  call fold (amp,imax,cal,nc,zcon,lz)
  time shift and normalize amplitudes
  for plotting purposes
go to 93
lz=lz+20
call zzero (amp,lz)
k=20
do 23 i=1,lz
  amp(k)=zcon(i)
  k=k+1
23
c
c
call norml0 (amp,lz)
continue
93
c
c
transform real synthetic to complex vector form
and calculate spectra, normalize amplitude
spectra for plotting purposes
nspec=512
signi=+1.
call czero (source,nspec)
do 70 i=1,lz
  source(i)=cmplx(amp(i),0.0)
  write (6,71)
  read (5,72) respp
70

```

```

c
c using this amplitude spectra a phase spectra is computed using
c a hilbert transform, minimum phase behaviour is assumed.
c each individual frequency component is then scaled, the
c resultant source wavelet is then inverse transformed and
c returned to the main program.
c
c dimension srce(4097),q(40),d(40,40),v(40),z(40),phi(4097)
c complex amspc(4097),qfilt(4097)
c dimension ampr(4097),amod(4097),hmod(4097),tamp(4097)
c character*10 resp2,resp3
c
c format (' ',)
c format (/8x,' freq',10x,' atten',10x,' mod',9x,' logmod'/)
c format (/8x,' amspc',23x,' qfilter'/)
c format (5x,15.5x,f13.4,5x,f13.4)
c format (' phase=',i5)
c format (' print out q-filter? (Y or N): ', $)
c format (a2)
c format (' print attenuation coeff. etc.? (Y or N): ', $)
c format (f13.4,f13.5,f13.5,f13.5)
c format (/5x,' phase no:',i5/)
c format (' jv=',i3)
c
c query user about print options
c
c write (6,108) jv
c if (jv.gt.1) go to 20
c write (6,8)
c read (5,7) resp3
c write (6,6)
c read (5,7) resp2
c continue
c
c zero dummy array used in calculation
c
c zero=0.0
c do 88 i=1,4097
c   amspc(i)=cmplx(zero,zero)
c
c pi=3.141593
c
c load time shifting array (increments of delt)
c
c js(1)=2
c js(2)=3
c js(3)=3
c js(4)=3
c js(5)=3
c js(6)=3
c js(7)=3
c js(8)=4
c js(9)=4
c js(10)=4
c js(11)=4
c js(12)=4
c js(13)=4
c js(14)=5
c js(15)=5
c js(16)=5

```

using this amplitude spectra a phase spectra is computed using a hilbert transform, minimum phase behaviour is assumed. each individual frequency component is then scaled, the resultant source wavelet is then inverse transformed and returned to the main program.

dimension srce(4097),q(40),d(40,40),v(40),z(40),phi(4097)
 complex amspc(4097),qfilt(4097)
 dimension ampr(4097),amod(4097),hmod(4097),tamp(4097)
 character*10 resp2,resp3

format (' ',)
 format (/8x,' freq',10x,' atten',10x,' mod',9x,' logmod'/)
 format (/8x,' amspc',23x,' qfilter'/)
 format (5x,15.5x,f13.4,5x,f13.4)
 format (' phase=',i5)
 format (' print out q-filter? (Y or N): ', \$)
 format (a2)
 format (' print attenuation coeff. etc.? (Y or N): ', \$)
 format (f13.4,f13.5,f13.5,f13.5)
 format (/5x,' phase no:',i5/)
 format (' jv=',i3)

query user about print options

write (6,108) jv
 if (jv.gt.1) go to 20
 write (6,8)
 read (5,7) resp3
 write (6,6)
 read (5,7) resp2
 continue

zero dummy array used in calculation

zero=0.0
 do 88 i=1,4097
 amspc(i)=cmplx(zero,zero)

pi=3.141593

load time shifting array (increments of delt)

js(1)=2
 js(2)=3
 js(3)=3
 js(4)=3
 js(5)=3
 js(6)=3
 js(7)=3
 js(8)=4
 js(9)=4
 js(10)=4
 js(11)=4
 js(12)=4
 js(13)=4
 js(14)=5
 js(15)=5
 js(16)=5

```

c
c if (resp.eq.'n'.or.resp.eq.'n') go to 113
c call iftc (nspec,source,signl)
c call zzero (send,nspec)
c call spcamp (nspec,source,send)
c nfi=(nspec/2)+1
c delf=1./(nspec*deltat)
c call normi0 (send,nfi)
c
c continue
c
c write results to plot file and line printer
c
c write (3,5) lz,deltat
c write (22,5) lz,deltat
c write (3,18)
c do 201 i=1,lz
c   write (3,10) amp(i)
c   write (22,10) amp(i)
c
c write (24,5) nfi,delf
c do 202 i=1,nfi
c   write (24,10) send(i)
c   write (3,1)
c   write (3,105) delf
c   write (3,19)
c   write (3,103) (i,send(i),i=1,nfi)
c
c continue
c
c calculate weighted reflection coefficient
c series and send to output file
c
c ncof=imax2+22
c call zzero (coeff,ncof)
c do 441 i=1,jk
c   jon=it(i)+21
c   scfa=f(i)
c   coeff(jon)=scfa
c   write (3,1)
c   write (26,5) ncof,deltat
c   do 442 i=1,ncof
c     write (26,10) coeff(i)
c
c   close (unit=21)
c   close (unit=22)
c   close (unit=23)
c   close (unit=25)
c   close (unit=24)
c   close (unit=26)
c
c stop
c end
c
cccccccccccccccccccccccccccccccccccccccccccccccccccccccccccc
c subroutine qfiltr (srce,no,ns,deltat,que,dist,vel,n,jv,tamp)
c qfilt applies a q-filter to an incident source pulse,
c one pulse for each contribution to the p-phase. the
c source pulse is transformed to the frequency domain and
c the number of cycles/layer is computed, this result is
c multiplied times the loss/cycle (calculated from q).

```

if (resp.eq.'n'.or.resp.eq.'n') go to 113
 call iftc (nspec,source,signl)
 call zzero (send,nspec)
 call spcamp (nspec,source,send)
 nfi=(nspec/2)+1
 delf=1./(nspec*deltat)
 call normi0 (send,nfi)

continue

write results to plot file and line printer

write (3,5) lz,deltat
 write (22,5) lz,deltat
 write (3,18)
 do 201 i=1,lz
 write (3,10) amp(i)
 write (22,10) amp(i)

write (24,5) nfi,delf
 do 202 i=1,nfi
 write (24,10) send(i)
 write (3,1)
 write (3,105) delf
 write (3,19)
 write (3,103) (i,send(i),i=1,nfi)

continue

calculate weighted reflection coefficient
 series and send to output file

ncof=imax2+22
 call zzero (coeff,ncof)
 do 441 i=1,jk
 jon=it(i)+21
 scfa=f(i)
 coeff(jon)=scfa
 write (3,1)
 write (26,5) ncof,deltat
 do 442 i=1,ncof
 write (26,10) coeff(i)

close (unit=21)
 close (unit=22)
 close (unit=23)
 close (unit=25)
 close (unit=24)
 close (unit=26)

stop
 end

cccccccccccccccccccccccccccccccccccccccccccccccccccccccccccc
 subroutine qfiltr (srce,no,ns,deltat,que,dist,vel,n,jv,tamp)
 qfilt applies a q-filter to an incident source pulse,
 one pulse for each contribution to the p-phase. the
 source pulse is transformed to the frequency domain and
 the number of cycles/layer is computed, this result is
 multiplied times the loss/cycle (calculated from q).

```

c
c
c
js(17)=5
js(18)=5
js(19)=5
js(20)=5
js(20)=5
transform source vector
c
c
c
lit=1
nfi=1024
nfd2pl=(nfi/2)+1
call czero (qfilt,nfi)
call czero (ampspc,nfi)
do 17 i=ns+1,nfi
  srce(i)=zero
  signi=1.0
  j=1
  do 18 i=10,nfi
    amspc(i)=cmplx(srce(j),zero)
    j=j+1
  call fftc (nfi,ampspc,signi)
  nfi=float(nfi)
  delf=1./(nfi*deltat)
loop over number of layers and apply q-filter
c
c
c
if (resp3.eq.'n'.or.resp3.eq.'n') go to 24
write (3,2)
continue
nm2=n-2
a0=1.
amod(1)=1.0
hmod(1)=0.5*log(amod(1))
fr=0.0
write (3,5) jv
do 40 if=2,nfd2pl
  fr=fr+delf
  at=(pi*dlist)/(que*vel)
  amod(if)=a0*exp(-1.*at*fr)
  hmod(if)=0.5*log(amod(if))
  if (resp3.eq.'n'.or.resp3.eq.'n') go to 40
  write (3,9) fr,at,amod(if),hmod(if)
  continue
  if=nfd2pl-1
  do 69 i=nfd2pl+1,nfi
    amod(i)=amod(if)
    hmod(i)=0.5*log(amod(i))
    if=if-1
69
c
c
c
use hilbert transform to calculate phase spectra
from amplitude spectra calculated above
c
c
c
call hlbrt (nfi,hmod,phi)
call reimag (nfi,amod,phi,qfilt)
jts=js(jv)
write (3,1)
write (3,*) jts
call shift (qfilt,deltat,delf,nfi,nfd2pl,jts)
c
c
c
apply q-filter

```

```

c
c
c
do 35 iff=1,nfi
  write (3,*) phi(iff),qfilt(iff)
  amspc(iff)=qfilt(iff)*ampspc(iff)
  continue
  if=nfd2pl-1
  do 32 i=nfd2pl+1,nfi
    x=real(ampspc(iff))
    y=(-1.0)*aimag(ampspc(iff))
    amspc(i)=cmplx(x,y)
    if=if-1
  do 98 i=1,nfi
    write (3,4) i,ampspc(i)
  inverse transform q-filtered pulse
c
c
c
signi=-1.0
call fftc (nfi,ampspc,signi)
call fftc (nfi,qfilt,signi)
write (3,1)
return real part of series to main program
c
c
c
k=1
do 83 ji=1,iit+1
  amspc(ji)=cmplx(zero,zero)
c
c
c
do 81 i=1,no
  tamp(i)=real(ampspc(i))
  write out test cases
c
c
c
if (resp2.eq.'n'.or.resp2.eq.'n') go to 93
write (3,3)
do 99 i=1,nfi
  write (3,*) amspc(i),qfilt(i)
  continue
  write (3,1)
c
c
c
return
end
c
c
c
subroutine fftc (lx,cx,signi)
  fftc is a cooley-tukey type fast fourier transform that
  utilizes complex input vectors (one-d algorithm)
  lx
  cx(k)=sqrt(1/lx) sum (cx(j)*exp(2*pi*signi*i*(j-1)*(k-1)/lx))
  j=1
  for k=1,2,...,(lx=2**integer)
  lx=length of x vector (input)
  cx=complex x vector (input)
  signi=-1.= forward transform
  =+1. inverse transform
  complex cx(2049),carg,cexp,cv,ctemp,ceye
  pi=3.14159265
  ceye=(0.0,1.0)
  j=1

```



```

22 read (22,6) y(i)
23 do 23 i=np2+1,np2+l0
24   y(i)=zero
25   do 24 i=1,npl+l0
26     write (3,6) x(i)
27     do 26 i=1,np2+l0
28       write (3,6) y(i)
29   c
30   c
31   resample data
32   ni=np2/npl
33   ns=npl+l
34   nc=ns+npl-l
35   do 70 j=1,ni
36     k=j
37     do 25 i=1,npl
38       xs(i)=y(k)
39     k=k+ni
40     xs(ns)=zero
41     do 27 i=1,ns
42       write (3,6) xs(i)
43   c
44   c
45   calculate correlation
46   call xcov (x,npl,xs,ns,c)
47   c
48   normalize cross correlation function
49   smul=zero
50   smu2=zero
51   do 36 i=1,ns
52     sum1=x(i)*x(i)
53     sum2=xs(i)*xs(i)
54     smul=smul+sum1
55     smu2=smu2+sum2
56   tnorm=sqrt(smul*smu2)
57   do 38 i=1,nc
58     c(i)=c(i)/tnorm
59   c
60   c
61   find max and min of correlation
62   ymax=zero
63   ymin=zero
64   do 40 i=1,nc
65     if (c(i).lt.ymin) ymin=c(i)
66     if (c(i).gt.ymax) ymax=c(i)
67   c
68   c
69   print results
70   write (3,1)
71   write (3,7) j
72   write (3,8) ymax,ymin
73   c
74   c
75   reset variables
76   cmax(j)=ymax
77   cmin(j)=ymin
78   call zzero (c,nc)

```

```

70 call zzero (xs,ns)
71 continue
72 determine best fit
73 do 72 i=1,ni
74   write (6,11) i,cmax(i),cmin(i)
75   write (6,12)
76   read (5,13) la
77   c
78   write (3,1)
79   write (3,9) la,li
80   resample data
81   k=la
82   do 54 i=1,npl
83     xs(i)=y(k)
84     k=k+ni
85     xs(ns)=zero
86   c
87   write to output file
88   write (23,5) ns,dellt
89   do 55 i=1,ns
90     write (23,6) xs(i)
91   c
92   close (unit=21)
93   close (unit=22)
94   close (unit=23)
95   stop
96   end

```

```

c      quickt.for, a quick fourier transform routine that
c      provides amplitude and phase spectra, input is from
c      a sample input file:(i,f)-npts,delt
c      (f)-amp(i)
c      to eof
c
c      the program uses fftc which is in progs.for, the cmd
c      file for quickt loads progs automatically. the input file
c      is specified by the user. the output files are phi.dat
c      and ptof.dat for the phase and amplitude spectra
c      respectively. input data (npts) does not have to be a
c      power of two. line printer output is also provided.
c      written by jon ake 5/86
c
c      program modified: 11/86, user now has the option of
c      normalizing output prior to writing to output file, in
c      addition the user may also apply a 3 point moving average
c      filter to the amplitude spectra if desired.
c
c      modified to run on the sun system, 7/87  jpa
c
c      dimension amps(2049),phas(2049),ftr(2049),fti(2049)
c      dimension fns(2049),fr(2049),amp3(2049)
c      complex camp(2049)
c      character*10 vfmin
c      character*3 resp,resp2
c
c      format (' ')
c      format (' specify input file: ', $)
c      format (a10)
c      format (i,f,1)
c      format (i7,f13.5)
c      format (f13.4)
c      format (5x,f13.4,5x,f13.4)
c      format (2x, ' fourier analysis of ',a10/)
c      format (' input desired length of time series: ', $)
c      format (i7)
c      format (/12x,' freq',15x,' amp',16x,' phase'//)
c      format (5x,f13.4,5x,f13.4,5x,f13.4)
c      format (' apply 3 pt. maf to spectra? (y or n) ', $)
c      format (' normalize output? (y or n) ', $)
c      format (a3)
c      format (f13.4,f13.4)
c
c      specify input file
c
c      write (6,2)
c      read (5,3) vfmin
c
c      open (unit=3,access='sequential',file='prfil.d')
c      open (unit=21,access='sequential',file='vfmin')
c      open (unit=40,access='sequential',file='ptof.dat')
c      open (unit=23,access='sequential',file='phi.dat')
c      open (unit=24,access='sequential',file='comp.dat')
c
c      provide documentation
c
c      write (3,1)
c      write (3,9) vfmin
c      write (3,1)

```

```

c      read in data, put in complex form
c
c      write (6,10)
c      read (5,11) np
c      do 66 i=1,np+1
c      camp(i)=(0.0,0.0)
c      read (21,6) npts,delt
c      write (6,11) np
c      write (6,6) npts,delt
c      do 22 i=1,npts
c      read (21,7) amps(i)
c      write (6,7) amps(i)
c      camp(i)=cmplx(amps(i),0.0)
c      do 23 i=1,np
c      amps(i)=0.0
c
c      calculate complex spectra
c
c      signi=-1.0
c      scale=1.0
c      n2=(np/2)+1
c      delf=1./ (np*delt)
c
c      call fftc (np,camp,signi)
c
c      write (3,1)
c      write (3,6) np,delt
c      write (24,6) np,delf
c      write (3,1)
c      do 25 i=1,np
c      write (24,67) camp(i)
c      write (3,67) camp(i)
c
c      calculate amplitude and phase spectra
c
c      do 28 i=1,np
c      ftr(i)=real(camp(i))
c      fti(i)=aimag(camp(i))
c
c      call compol (ftr,fti,np,scale,amps,phas)
c
c      normalize amplitude spectra for plotting purposes
c
c      write (6,4)
c      read (5,12) resp
c      if (resp.eq.'n'.or.resp.eq.'n') go to 43
c
c      amax=0.0
c      do 40 i=1,np
c      if (amps(i).gt.amax) amax=amps(i)
c      do 42 i=1,np
c      amps(i)=9.9*amps(i)/amax
c
c      apply 3 pt. moving average filter if desired
c
c      write (6,15)
c      read (5,12) resp2
c      if (resp2.eq.'n'.or.resp2.eq.'n') go to 43
c      amp3(1)=(0.0+amps(1)+0.5*amps(2))/3.

```

```

44      amp3(np)=(0.5*amps(np-1)+amps(np)+0.0)/3.
45      amp3(i)=(0.5*amps(i-1)+amps(i)+0.5*amps(i+1))/3.
46      amp3(i)=amp3(i)
47      calculate frequency vector
48      continue
49      do 16 i=1,n2
50      fi=i-1
51      fr(i)=fi*delf
52      send output to line printer and plot files
53
54      format (' output files:')
55      write (6,99)
56      write (3,1)
57      write (3,6) n2,delf
58      write (3,13)
59      do 30 i=1,n2
60      write (3,14) fr(i),amps(i),phas(i)
61      write (40,6) n2,delf
62      write (6,6) n2,delf
63      write (23,6) n2,delf
64      do 32 i=1,n2
65      write (40,7) amps(i)
66      write (6,7) amps(i)
67      write (23,7) phas(i)
68      close (unit=3)
69      close (unit=21)
70      close (unit=40)
71      close (unit=23)
72      close (unit=24)
73      stop
74      end
75
76      30
77
78      32
79
80      43
81
82      16
83
84      99
85
86      30
87
88      32
89
90      43
91
92      16
93
94      99
95
96      30
97
98      32
99

```

```

PROGRAM:STATISTICS AND LINEAR
REGRESSION OF DATA
MAY 1983, JON AKE
(updated 11/84, jpa)
symbols used:
xsum      sum of x values (independent)
ysum      sum of y values (dependent)
x2sum     sum of squares of x values
y2sum     sum of squares of y values
xysum     sum of products of x and y values
xbar      1st moment (mean) of x values
ybar      1st moment (mean) of y values
xvar      variance of x values
yvar      variance of y values
xycov     covariance between x and y values
slope     slope of least squares line
b          intercept of least squares line
corr       correlation coefficient of least squares line
stdev     standard deviation of slope values
n          number of data points
stdvx     standard deviation in x values
stdvy     standard deviation in y values
ssq       sum of squares of errors
p(i)      dummy end points used in plotting
pt1       min x value for plot
pt2       max x value for plot
input file format:
n          (i)
pt1,pt2   (f,f)
x(i),y(i) (f,f)
DIMENSION X(100),Y(100)
character vfin*10
format (' designate input data file: ', $)
3 format (//5X, ' OUTPUT:STATISTICS & LINEAR REGRESSION PROG. '//)
4 format (//10X, ' DEGREE OF CORRELATION:',F8.4/)
5 format (//10X, ' XBAR=',F8.4, ' YBAR=',F8.4/)
6 format (//10X, ' STD. DEV. IN X=',F8.4, ' STD. DEV. IN Y=',F8.4/)
7 format (//10X, ' SLOPE OF LEAST SQUARES LINE:',F8.4/)
8 format (//10X, ' Y-INTERCEPT OF LST. SQRS. LINE:',F8.4/)
9 format (//10X, ' CORRELATION COEFFICIENT OF LINE:',F8.4/)
10 format (//10X, ' STD. DEV. OF SLOPE VALUE,+/-:',F8.4/)
11 format (//10X, ' STD. DEV. OF SLOPE VALUE,+/-:',F8.4/)
12 format (//10X, ' STD. DEV. OF SLOPE VALUE,+/-:',F8.4/)
13 format (//10X, ' STD. DEV. OF SLOPE VALUE,+/-:',F8.4/)
14 format (//10X, ' STD. DEV. OF SLOPE VALUE,+/-:',F8.4/)
15 format (//10X, ' STD. DEV. OF SLOPE VALUE,+/-:',F8.4/)
16 format (//10X, ' STD. DEV. OF SLOPE VALUE,+/-:',F8.4/)
17 format (//10X, ' STD. DEV. OF SLOPE VALUE,+/-:',F8.4/)
input and open data file
write (5,3)
read (5,16) vfin
OPEN (UNIT=1,DEVICE='DSK',FILE=vfin,ACCESS='SEQU')

```

```

C documentation
C WRITE (3,4)
C INPUT DATA
C READ (1,12) N
C READ (1,13) PTL,PT2
C DO 50 I=1,N
C   READ (1,13) X(I),Y(I)
C   INITIALIZATION
C   XSUM=0.
C   YSUM=0.
C   FN=N
C   X2SUM=0.
C   Y2SUM=0.
C   FN2=FN-2.
C   XYSUM=0.
C
C   COMPUTE SUMS BY LOOPING OVER X VALUES
C   DO 100 I=1,N
C     XSUM=XSUM+X(I)
C     YSUM=YSUM+Y(I)
C     X2SUM=X2SUM+X(I)**2
C     Y2SUM=Y2SUM+Y(I)**2
C     XYSUM=XYSUM+(X(I)*Y(I))
C     CONTINUE
C   100
C   COMPUTE FIRST MOMENTS
C   XBAR=XSUM/FN
C   YBAR=YSUM/FN
C   X2BAR=X2SUM/FN
C   Y2BAR=Y2SUM/FN
C   X1BAR=XYSUM/FN
C
C   COMPUTE VARIANCES AND PARAMETERS
C   XVAR=X2BAR-(XBAR**2)
C   YVAR=Y2BAR-(YBAR**2)
C   XYCOV=XYBAR-XBAR*YBAR
C   SLOPE=XYCOV/XVAR
C   B=YBAR-SLOPE*XBAR
C   CORR=-XYCOV/SQRT(XVAR*YVAR)
C   STDVX=SQRT(XVAR)
C   STDVY=SQRT(YVAR)
C
C   WRITE RESULTS
C   WRITE (3,17) VENIN
C   WRITE (3,15) N
C   WRITE (3,6) XBAR,YBAR
C   WRITE (3,7) STDVX,STDVY
C   WRITE (3,8) SLOPE
C   WRITE (3,9) B
C   WRITE (3,10) CORR
C
C
C   INITIALIZE AND DEFINE SUMS FOR
C   ESTIMATING ERROR IN SLOPE
C   XLSUM=0.
C   YLSUM=0.
C   DO 300 I=1,N
C     YLSUM=(Y(I)-YBAR)**2+YLSUM
C     XLSUM=(X(I)-XBAR)**2+XLSUM
C     CONTINUE
C   300
C   P1=B-SLOPE
C   P2=B+PT2*SLOPE
C   WRITE (3,14) P1,P2
C
C   COMPUTE ERROR IN SLOPE
C   SSQ=(YLSUM-SLOPE**2*XLSUM)/FN2
C   STDEV=SQRT(SSQ)/SQRT(XLSUM)
C   WRITE (3,11) STDEV
C
C   SL1=SLOPE+STDEV
C   SL2=SLOPE-STDEV
C   P3=SL1*PT1-SL1*XBAR+YBAR
C   P4=SL1*PT2-SL1*XBAR+YBAR
C   P5=SL2*PT1-SL2*XBAR+YBAR
C   P6=SL2*PT2-SL2*XBAR+YBAR
C
C   CALL SUBPLT (X,Y,N,N1,N2,XBAR,YBAR,P1,P2,PT1,PT2,
C   * P3,P4,P5,P6)
C
C   CLOSE (UNIT=1,DEVICE='DSK')
C   CLOSE (UNIT=3,DEVICE='LPT')
C
C   STOP
C   END
C
C   PLOT SUBROUTINE
C
C   SUBROUTINE SUBPLT (X,Y,N,N1,N2,XBAR,YBAR,P1,P2,PT1,PT2,
C   * P3,P4,P5,P6)
C
C   DIMENSION X(100),Y(100)
C
C   CALL INITIAL (1)
C   XAX='NAG.'
C   YAX='LOGT.'
C   FVX=-1.
C   DVX=1.
C   FVY=0.
C   DVY=1.
C   CALL PLOT (2,1,1,-3)
C   CALL AXIS (-1,0,0,XAX,-4,6,0,0,FVX,DVX)
C   CALL AXIS (-1,0,0,YAX,4,4,90,0,FVY,DVY)
C   DO 50 I=1,N1
C     CALL SYMBOL (X(I),Y(I),.05,14,0,0,-1)
C   DO 52 I=N1+1,N
C     CALL SYMBOL (X(I),Y(I),.06,0,0,0,-1)
C   CALL SYMBOL (XBAR,YBAR,.10,4,0,0,-1)
C   CALL PLOT (PT1,PT1,13)
C   CALL PLOT (PT2,PT2,12)

```



```

1059 READ (5,1059) ANSWER
      FORMAT (A5)
      IF (ANSWER.EQ. 'Y'.OR. ANSWER.EQ. 'y') GO TO 1070
735   write (5,735)
      format(' ***NO EVENTS PROCESSED***')
      go to 931
c
16   write (5,731)
731   format(' input events sequentially from file?', $)
      read (5,1059) reply
c
c
c
      READ IN MASTER EVENT DATA
      DO 30 I=1,1991,10
      READ(1,40) (Y1(J),J=I,I+9)
      FORMAT(10F5.0)
      IF(Y1(I).EQ. 99999.) GO TO 45
      GO TO 30
      NPTS1=I-1
      GO TO 56
      CONTINUE
      CONTINUE
56   CONTINUE
      WRITE (3,3) CMPL
      FORMAT (///3X,' MASTER EVENT: ',LA25)
c
c
c
      rewind?
      if (reply.eq.'Y'.or.reply.eq.'y') go to 733
      REWIND (1)
      continue
c
c
c
      read in second event header
c
c
1171  READ (1,1065,END=1200) CMP2
1189  CONTINUE
      GO TO 21
1200  WRITE (5,1058)
1058  FORMAT (3X,' END OF FILE ENCOUNTERED')
c
      WRITE (5,1057)
      FORMAT (' INPUT NEW MASTER EVENT?<Y OR N>')
      READ (5,1059) RESP
      IF (RESP.EQ.'Y') GO TO 1070
      CONTINUE
c
67
c
      write (5,734)
      format(' ***END OF RUN***', $)
      go to 931
c
c
c
      READ IN second event data
c
c
21   DO 50 I=1,1991,10
      READ(1,40,END=55) (Y2(J),J=I,I+9)
      IF(Y2(I).EQ. 99999.) GO TO 55
      GO TO 50
      NPTS2=I-1
      GO TO 57
      CONTINUE
55   CONTINUE
50   CONTINUE
57   CONTINUE

```

```

      WRITE (3,2) CMP2
      FORMAT (10X,' SECOND EVENT: ',LA25/)
      call calc (npts1,npts2,y1,y2,c,ymin,ymax,j1,j2)
      print out results
      write (3,180) ymin,j1
      write (3,182) ymax,j2
      format (10x,' min corr=',f6.3,' at j=',i5)
      format (10x,' max corr=',f6.3,' at j=',i5////)
      go to 1171
c
c
c
      close (unit=1)
      stop
      end
c
c
c
      subroutine calc (npts1,npts2,y1,y2,c,ymin,ymax,j1,j2)
      dimension y1(1500),y2(1500),c(3000)
      SET UP PARAMETERS FOR CROSS-CORRELATION
      IF (NPTS1.EQ.NPTS2) GO TO 94
      IF (NPTS1.GT.NPTS2) GO TO 92
      DO 91 I=NPTS1+1,NPTS2
      Y1(I)=0.
      N=NPTS2
      GO TO 95
      DO 93 I=NPTS2+1,NPTS1
      Y2(I)=0.
      N=NPTS1
      GO TO 95
      N=NPTS1
      CONTINUE
      NLAGS=2*N-1
      LMAX=N
      LMIN=-1*(N-1)
      COMPUTE CROSS-CORRELATION
      COMPUTE NEGATIVE LAG VALUES
      DO 210 J=1,N
      TERM=0.
      IT=N-J+1
      DO 200 I=1,J
      TERM=Y1(I)*Y2(IT)
      TERM=TERM+TERM1
      IT=IT+1
      C(J)=TERM
      200
      210
      COMPUTE POSITIVE LAG VALUES
      DO 220 J=2,N
      TERM=0.
      IT=1
      DO 215 I=J,N

```

```

C      Shaper. for
C      Simple program to calculate the shaping filter (f)
C      given the input wavelet to the filter (b) and the
C      recorded trace (d). The program uses subroutines
C      contained in progs.for, the .cmd file loads progs
C      for the user. The input files are specified by the user
C      and are of standard format:
C      recorded trace
C      nptsl,dt      (i,f)
C      d(i)          (f)
C      .
C      .
C      d(nptsl)
C      source wavelet
C      npts2,dt      (i,f)
C      b(i)          (f)
C      .
C      .
C      b(npts2)      eof
C
C      The output file 'shape.dat', contains the filter coefficients.
C      Format is identical to the input file (one vector), format is
C      compatible with plot.for etc. Printed output is also provided.
C      The output file 'output.dat' contains the convolution of the
C      filter coefficients with the source wavelet.
C
C      B=INPUT WAVELET TO SHAPING FILTER
C      D=RECORDED SEISMIC TRACE
C      F=ESTIMATE OF SHAPING FILTER
C      Z=FILTER OUTPUT
C
C      WRITTEN BY JON AKE 6/86
C      REVISED TO USE TWO SEPARATE INPUT FILES, 7/86
C
C      DIMENSION B(1025),D(1025),F(1025),C(2049),Z(2049)
C      CHARACTER*10 VFMIN,resp,VFNIN2
C
C      FORMAT (' ')
C      FORMAT (5X,F,F,F,F,F,F,F,F,F,F)
C      FORMAT (5X,I,I,I,I,I,I)
C      format (i,f)
C      format (f)
C      format (' specify SEISMIC TRACE input file: ', $)
C      format (a10)
C      format (' specify length of filter: ', $)
C      format (i)
C      FORMAT (/10X, ' FILTER COEFFICIENTS'//)
C      FORMAT (/10X, ' INPUT WAVELET'//)
C      FORMAT (/10X, ' RECORDED SEISMIC TRACE'//)
C      FORMAT (/10X, ' FILTER OUTPUT'//)
C      format (' write filter output to file? (y or n): ', $)
C      format (a2)
C      FORMAT (' SPECIFY SOURCE WAVELET INPUT FILE: ', $)
C
C      specify input files
C
C      write (5,7)
C      read (5,8) vfnin
C      WRITE (5,17)
C      READ (5,8) VFNIN2

```

```

C      TERM1=Y1(I)*Y2(IT)
C      TERM=TERM+TERM1
C      IT=IT+1
C      C(N+J-1)=TERM
C
C      COMPUTE NORMALIZATION FACTOR FOR
C      CROSS-CORRELATION
C
C      SMU1=0.
C      SMU2=0.
C      DO 760 I=1,N
C      SUM1=Y1(I)*Y1(I)
C      SUM2=Y2(I)*Y2(I)
C      SMU1=SMU1+SUM1
C      SMU2=SMU2+SUM2
C      TNORM=SQRT(SMU1*SMU2)
C
C      NORMALIZE CROSS-CORRELATIONS
C
C      DO 790 I=1,NLAGS
C      C(I)=C(I)/TNORM
C
C      NLAG1=NLAGS+1
C
C      FIND MAX AND MIN OF CROSS-
C      CORRELATION FUNCTION
C
C      YMAX=0.
C      YMIN=0.
C      DO 190 J=1,NLAG1
C      IF (C(J).LT.YMIN) GO TO 186
C      IF (C(J).GT.YMAX) GO TO 188
C      GO TO 190
C      J1=J
C      YMIN=C(J)
C      GO TO 190
C      YMAX=C(J)
C      J2=J
C      CONTINUE
C
C      return
C      end

```

```

C      open (unit=21,device='dsk',access='seqin',file=vfnin)
C      open (unit=22,device='dsk',access='seqout',file='shape.dat')
C      open (unit=23,device='dsk',access='seqout',file='output.dat')
C      OPEN (UNIT=24,DEVICE='DSK',ACCESS='SEQIN',FILE=VFNN2)
C
C      read in data
C
C      read (21,5) ld,dt
C      do 20 i=1,ld
C        read (21,6) D(i)
C        read (24,5) lb,dt
C        do 22 i=1,lb
C          read (24,6) B(i)
C          write (5,9)
C          read (5,10) N
C          LZ=N+LB-1
C
C      normalize input data
C      bmax=0.0
C      do 60 i=1,lb
C        if (b(i).gt.bmax) bmax=b(i)
C        do 61 i=1,lb
C          b(i)=b(i)/bmax
C          dmax=0.0
C        do 62 i=1,ld
C          if (d(i).gt.dmax) dmax=d(i)
C          do 63 i=1,ld
C            d(i)=d(i)/dmax
C            do 64 i=ld+1,lb
C              d(i)=0.0
C            C
C          C
C        C
C      calculate shaping filter and actual filter output
C      CALL SHAPE (B,LB,D,LD,N,F,C,LC)
C      CALL FOLD (B,LB,F,N,Z,LZ)
C
C      send results to line printer
C
C      WRITE (3,2)
C      WRITE (3,5) lb,dt
C      write (3,5) ld,dt
C      WRITE (3,2)
C      WRITE (3,4) LC
C      WRITE (3,2)
C      WRITE (3,11)
C      do 24 i=1,n
C        write (3,*) f(i)
C      WRITE (3,2)
C      WRITE (3,12)
C      do 25 i=1,lb
C        WRITE (3,*) B(i)
C      WRITE (3,2)
C      WRITE (3,13)
C      do 26 i=1,ld
C        WRITE (3,*) D(i)
C      WRITE (3,2)
C      WRITE (3,14)
C      DO 27 I=1,LZ

```

```

27      WRITE (3,*) Z(I)
C      write estimate of filter to output file
C
C      write (22,5) N,dt
C      do 28 i=1,N
C        write (22,6) F(i)
C        write (5,15)
C        read (5,16) resp
C        if (resp.eq.'n'.or.resp.eq.'N') go to 30
C        write (23,5) lz,dt
C        do 29 i=1,lz
C          write (23,6) z(i)
C        continue
C      close (unit=21,device='dsk')
C      close (unit=22,device='dsk')
C      close (unit=23,device='dsk')
C      CLOSE (UNIT=24,DEVICE='DSK')
C      STOP
C      END

```

```

C      PLOT FOR
C      PLOT.FOR IS A SIMPLE PROGRAM TO PLOT A TIME SERIES.
C      THE PROGRAM IS USED WITH THE SYNTHETIC SEISMOGRAM
C      PROGRAMS SYNTH.FOR AND CONV.FOR, IT READS AN INPUT
C      FILE. *** DAT, STIPULATED BY THE USER. FORMAT IS
C      IMAX,DELTAT (I,F)
C      AMP(I)
C      .
C      .
C      .
C      EOF
C
C      THE PROGRAM UTILIZES GRELIB SUBROUTINES, THESE ARE
C      LOADED BY THE PLOT.CMD FILE. A LIMIT OF NPTS<1000 IS
C      (UNFORTUNATELY) REQUIRED BY THE GRELIB ROUTINES.
C      WRITTEN BY JON AKE JUNE 1985.
C      TO PLOT ON THE TEKTRONIX PLOTTER USE:
C      OFFSETS=(0.25,0.30); SCALE FACTORS=(1.3,1.3)
C
C      Modified for compatibility with SUN machines-6/87
C
C      DIMENSION AMP(4097),T(4097),amp1(4097)
C      CHARACTER*10 head,VFNIN
C
C      3  FORMAT (' DESIGNATE INPUT FILE: ', $)
C      4  FORMAT (' INPUT HEADER FOR PLOT, 10 CHARS. MAX', $)
C      5  FORMAT (I7,F13.5)
C      6  FORMAT (F13.4)
C      7  FORMAT (A10)
C      8  FORMAT (' INPUT DESIRED NUMBER OF POINTS FOR PLOT: ', $)
C      9  FORMAT (I7)
C
C      DESIGNATE INPUT FILE
C
C      WRITE (6,3)
C      READ (5,7) VFNIN
C
C      OPEN (UNIT=40,ACCESS='Sequential',FILE=VFNIN)
C
C      READ IN DATA
C
C      WRITE (6,8)
C      READ (5,9) IMAX
C      READ (40,5) NPTS,DELTAT
C      DO 81 I=1,1000
C      81  AMP(I)=0.0
C      DO 26 I=1,NPTS
C      26  READ (40,6) AMP(I)
C      DO 10 I=1,IMAX
C      10  FI=I
C      T(I)=(FI-1.)*DELTAT
C
C      DO 73 I=1,npts
C      73  write (6,74) amp(i)
C      74  format (3x,f20)
C
C      DO 93 I=1,npts
C      93  amp(i)=-1.*amp(i)
C
C      CC
C      CC

```

```

C      INITIALIZE FOR PLOTTING
C
C      call initial(-1,'output.plt')
C
C      RESCALE PLOT FIELD AND DATA
C
C      CALL FACTOR (0.80)
C      CALL SCALE (T,10.0,IMAX,1)
C      CALL SCALE (AMP,8.0,IMAX,1)
C
C      PLOT AND LABEL AXES
C
C      I1=IMAX-1
C      I2=IMAX-2
C      CALL AXIS (1.0,1.0,'TIME (SEC)',-10,10.0,0.0,T(I1),T(I2))
C      CALL AXIS (1.0,1.0,'AMPL.',5,8.0,90.0,AMP(I1),AMP(I2))
C
C      PLOT TIME SERIES
C
C      CALL PLOT (1.0,1.0,-3)
C      CALL LINE (T,AMP,IMAX,1,0,0)
C
C      ADD TITLE
C
C      WRITE (6,4)
C      READ (5,7) head
C      CALL NEWPEN (3)
C      CALL SYMBOL (4.5,8.0,0.2,head,0.0,10)
C
C      call rstr(-1)
C
C      CLOSE (UNIT=40)
C
C      STOP
C      END
C

```

```

c      plot2.for
c      plot2.for is a simple program to superimpose and plot
c      a pair of time series. the program is used with the
c      output of the synthetic seismogram programs synths.for
c      and conv.for, it reads input files, ***.dat, stipulated
c      by the user. format is 'standard':
c      imax,deltat (i,f)
c      amp(i)      (f)
c      .
c      .
c      eof
c      (for both files)
c      the program utilizes grflib subroutines, these are
c      loaded by the plot.cmd file. a limit of npts<1000 is
c      (unfortunately) required by the grflib routines.
c      written by jon ake sept. 1986
c      to plot on the tektronix plotter use:
c      offsets=(0.25,0.30); scale factors=(1.3,1.3)
c      modified to run on the sun system, 7/87 jpa
c
c      dimension amp(4097),t(4097),amp1(4097),x1(10),x2(10)
c      dimension y1(10),y2(10)
c      character*10 head,vfnin,file2,head1,head2
c
c      format (' ')
c      format (' o.k. ')
c      format (' designate first input file(obs.): ',%)
c      format (' input header for plot, 10 chars. max',%)
c      format (i7,f13.5)
c      format (f13.4)
c      format (a10)
c      format (' input desired number of points for plot: ',%)
c      format (i6)
c      format (' designate second input file(synth.): ',%)
c      format (5x,f13.4,2x,f13.4)
c      format (/, scale factors: '/')
c      format (' input desired scale factors: ',%)
c      format (f13.4,f13.4)
c
c      designate input files
c
c      write (6,3)
c      read (5,7) vfnin
c      write (6,10)
c      read (5,7) file2
c
c      open (unit=22,access='sequential',file=vfnin)
c      open (unit=40,access='sequential',file=file2)
c
c      read in data
c
c      write (6,8)
c      read (5,9) imax
c      read (22,5) npts1,deltat
c      do 81 i=1,1000
c      amp(i)=0.0
c      amp(i)=0.0

```

```

26  do 26 i=1,npts1
    read (22,6) amp(i)
    read (40,5) npts2,deltat
    do 28 i=1,npts2
28  read (40,6) amp1(i)
    do 12 i=1,imax
    fi=1
12  t(i)=(fi-1.)*deltat
    normalize data for plotting
    c
    c
    ymax=0.0
    ymax1=0.0
    n=npts1
    if (npts2.gt.npts1) n=npts2
    do 30 i=1,n
    if (abs(amp(i)).gt.ymax) ymax=abs(amp(i))
30  if (abs(amp1(i)).gt.ymax1) ymax1=abs(amp1(i))
    do 32 i=1,n
    amp(i)=7.8*amp(i)/ymax
32  amp1(i)=7.8*amp1(i)/ymax1
    initialize for plotting
    call initial(-1,'output.plt')
    c
    c
    input header parameters
    head1='data'
    head2='synth'
    x1(1)=4.5
    x1(2)=5.5
    y1(1)=7.6
    y1(2)=7.6
    y2(1)=7.1
    y2(2)=7.1
    c
    c
    rescale plot field and data
    call factor (0.80)
    call scale (t,10.0,imax,1)
    call scale (amp,8.0,imax,1)
    call scale (amp1,8.0,imax,1)
    c
    c
    plot and label axes
    il=imax+1
    i2=imax+2
    write (6,13)
    write (6,11) amp(il),amp(i2)
    write (6,1)
    write (6,11) amp1(il),amp1(i2)
    write (6,14)
    read (5,15) amp(il),amp(i2)
    amp1(il)=amp(il)
    amp1(i2)=amp(i2)
    call axis (1.0,1.0,'time (sec)',-10.10,0.0,0,t(il),t(i2))
    call axis (1.0,1.0,'velocity',8.8,0.90,0,amp(il),amp(i2))
    c

```



```

12  t(i)=(fi-l.)*delf
c
c  normalize for plotting
c  ymax=0.0
c  ymax1=0.0
c  do 30 i=1,imax
c    if (abs(amp(i)).gt.ymax) ymax=abs(amp(i))
c    if (abs(amp1(i)).gt.ymax1) ymax1=abs(amp1(i))
c  do 32 i=1,imax
c    amp(i)=9.8*amp(i)/ymax
c    amp1(i)=9.8*amp1(i)/ymax1
c 32
c
c  initialize for plotting
c
c  call initl(-1,'output.plt')
c
c  head1='data'
c  head2='synth'
c
c  rescale plot field and data
c
c  call factor (0.80)
c  call scale (t,10.0,imax,1)
c  call scale (amp,8.0,imax,1)
c  call scale (amp1,8.0,imax,1)
c
c  plot and label axes
c
c  il=imax+1
c  i2=imax+2
c  write (6,13)
c  write (6,11) amp(il),amp(i2)
c  write (6,1)
c  write (6,11) amp1(il),amp1(i2)
c  write (6,14)
c  read (5,15) amp(il),amp(i2)
c  amp(il)=amp(il)
c  amp(i2)=amp(i2)
c  call axis (1.0,1.0,'freq (hz)',-9,10.0,0.0,t(il),t(i2))
c  call axis (1.0,1.0,'amp1.',5,8.0,90.0,amp1(il),amp1(i2))
c
c  plot time series
c
c  call newpen (6)
c  call plot (1.0,1.0,-3)
c  call line (t,amp,imax,1,0,0)
c  call plot (0.0,0.0,3)
c  call newpen (7)
c  call line (t,amp1,imax,1,2,2)
c
c  add title and header
c
c  write (6,4)
c  read (5,7) head
c  call newpen (1)
c  call symbol (4.5,8.0,0.2,head,0.0,10)
c  call symbol (5.6,7.7,0.2,head1,0.0,10)
c  call symbol (5.6,7.2,0.2,head2,0.0,10)
c  call plot (4.5,7.8,3)

```

```

c  call newpen (6)
c  call plot (5.5,7.8,2)
c  call plot (4.5,7.3,3)
c  call newpen (7)
c  call plot (5.5,7.3,2)
c  call symbol (4.5,7.3,0.1,2,0.0,-1)
c  call symbol (5.5,7.3,0.1,2,0.0,-1)
c
c  call rstr (-1)
c
c  close (unit=40)
c  close (unit=22)
c
c  stop
c  end
c

```

```

C      DIMENSION AMP(NSAM)
C
C-----NORMALIZE BY MAX VALUE
C
C      CALL WYNORM(AMP,NSAM)
C
C-----CONVERT TO DECIBELS
C
C      DO 100 I=1,NSAM
C      100  AMP(I) = 20.0 * ALOG10(AMP(I))
C
C      RETURN
C      END

```

```

C CCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCC
SUBROUTINE DECON(X,LX,N,ALPHA,WHITN,LUN,W,Y)
C-----GENERAL DECON SUBROUTINE TO DO SPIKING AND PREDICTIVE
DECONVOLUTION
C-----INPUT PARAMETERS
C X = INPUT TRACE TO BE DECONVOLVED
C LX = LENGTH OF X
C N = LENGTH OF DECON OPERATOR MUST BE > ALPHA
C ALPHA = PREDICTION DISTANCE IN NUMBER OF SAMPLES
C SET ALPHA = 1 FOR SPIKING DECON
C SET ALPHA > 1 FOR PREDICTION DECON
C WHITN = AMOUNT OF PREWHITENING IN UNITS OF PERCENT
C LUN = DEVICE NUMBER FOR ERROR MESSAGES
C-----OUTPUT PRODUCTS:
C W = DECON OPERATOR FOR SPIKING DECON (ALPHA=1)
C W WILL HAVE LENGTH N
C - PRED-ERROR OPERATOR FOR ALPHA > 1
C W WILL HAVE LENGTH N+ALPHA
C Y = DECONVOLVED OUTPUT WILL HAVE LENGTH LX+N+ALPHA-1
C ALL VALUES BEYOND INDEX LX WILL BE ZEROED
C
C DIMENSION X(1),W(1),Y(1),SPC1(300),SPC2(250),SPC3(250)
C IALF = IFIX(ALPHA - 0.5) + 1
C IF(IALF.LE.O) IALF = 1
C LW = IALF + N
C LT1 = LW + 1
C LT2 = LT1 + N
C-----DO SOME CHECKING OF POTENTIAL USER ERRORS
C IF(WHITN.LT.O .OR. WHITN.GT.100.)
$ WRITE(LUN,6005)WHITN
6005 FORMAT('X','*****ERROR IN DECON*****VALUE OF PREWHITENING IS ',F8.2,
$ , '---MUST BE BETWEEN 0-100')
C IF(N.LT.IALF) WRITE(LUN,6010)IALF,N
6010 FORMAT('X','*****ERROR IN DECON*****VALUE OF PRED-DIST IS ',I5,
$ , 'AND OPERATOR LENGTH IS ',I5,//
$ IX,'----PRED-DIST MUST BE < OPER LENGTH')
C IF(LW.GT.350) WRITE(LUN,6020)IALF,N
6020 FORMAT('X','*****ERROR IN DECON*****VALUE OF PRED-DIST IS ',I5,
$ , 'AND OPERATOR LENGTH IS ',I5,//
$ IX,'----SUM OF TWO CANNOT EXCEED 300')
C-----CALCULATE AUTO CORRELATION MATRIX OF INPUT TRACE
C CALL CROSS(X,LX,X,LX,LW,SPC1)
C-----CALCULATE RHS OF WIENER-HOPF EQUATION
C IF(IALF.EQ.1) THEN
C-----CALCULATE A SPIKE FOR SPIKING DECON
NOTE THAT SCALE FACTOR ON CROSS-COR IS UNKNOWN
CALL IMPULS(SPC2,N,1)
ELSE
C-----CALCULATE SHIFTED AUTO COR FOR PREDICTIVE DECON

```

```

C-----SUBROUTINE BNDFS(LFILT,SR,FL,FH,FILT)
C
C---BANDPASS FILTER SUBROUTINE
C   PRODUCES A ZERO PHASE FILTER OPERATOR THAT PASSES FREQUENCIES
C   BETWEEN FL AND FH
C
C-----INPUT PARAMETERS
C   LFILT = NUMBER OF SAMPLES IN FULL SYMMETRIC FILTER-----MUST BE ODD
C   SR = SAMPLE SPACING BETWEEN FILTER POINTS IN MS
C   FL = LOW PASS FREQ IN HZ
C   FH = HI PASS FREQ IN HZ
C
C-----OUTPUT
C   FILT = FILTER OPERATOR OF LENGTH LFILT
C
C+++++NOTE: IF (FH-FL) LESS THAN 1000/(SR*LFILT)
C            THEN FILTER GENERATED WILL BE NARROWEST POSSIBLE
C            CENTERED AT (FL+FH)/2
C
C            DIMENSION FILT(1)
C            DATA PI /3.14159265/
C
C            DT = SR/1000.
C            IF (FH .GT. 0.5/DT) FH = 0.75*0.5/DT
C
C            FN = LFILT
C            IF ((FH-FL) .LT. 1.0/(DT*FN) ) THEN
C              FC = (FH+FL)/2.0
C              WL = 2.0*PI*(FC*DT - 0.5/FN)
C              WH = 2.0*PI*(FC*DT + 0.5/FN)
C            ELSE
C              WL = FL*DT*2.0*PI
C              WH = FH*DT*2.0*PI
C
C            ENDF
C            CALL WCNTR(LFILT,M)
C            FILT(M) = (WH - WL)/(DT*PI)
C            M1 = M+1
C            DO 50 I=M1,LFILT
C              FI = I-M
C              FILT(I) = (SIN(WH*FI) - SIN(WL*FI))/(PI*FI*DT)
C            50 CONTINUE
C
C-----FILL IN SYMMETRIC PART OF FILTER
C            FILT WILL BE CENTERED AT (LFILT+1)/2
C
C            MM = M+M
C            DO 100 I=1,M
C              J = MM-I
C              FILT(I) = FILT(J)
C            100 RETURN
C            END

```

```

C-----
SUBROUTINE EUREKA(LR,R,G,F,A)
C---EUREKA SOLVES THE LEAST SQUARES NORMAL EQUATIONS
C    FOR F(I) AND THE PREDICTION ERROR OPERATOR A(I)
C---
C    INPUT PARAMETERS
C    LR = LENGTH OF ALL ARRAYS---NUMBER OF POINTS IN INVERSE FILTER
C    R = AUTOCOR OF INPUT SIGNAL FROM ZERO-LAG TO LAG LR-1
C    G = CROSSCOR BETWEEN SIGNAL AND DESIRED WAVESHAPE
C---
C    OUTPUT PRODUCTS
C    F = ARRAY CONTAINING FILTER COEFFS
C    A = COEFFS OF UNIT STEP PRED-ERROR OPERATOR
C    DIMENSION R(LR),G(LR),F(LR),A(LR)
C    V = R(1)
C    A(1) = 1.0
C    F(1) = G(1)/V
C    IF(LR.LE.1) RETURN
C
DO 60 L=2,LR
D = 0.0
Q = 0.0
L3 = L - 1
DO 10 J=1,L3
K = L - J + 1
D = D + A(J)*R(K)
Q = Q + F(J)*R(K)
10 CONTINUE
C = D/V
IF(L.EQ.2) GOTO 40
L1 = (L-2)/2
L2 = L1 + 1
IF(L2.LT.2) GOTO 30
DO 20 J=2,L2
HOLD = A(J)
K = L - J + 1
A(J) = A(J) - C*A(K)
A(K) = A(K) - C*HOLD
20 CONTINUE
IF(2*L1.EQ.L-2) GOTO 40
L3 = L2 + 1
A(LT3) = A(LT3) - C*A(LT3)
40 A(L) = -C
V = V - C*D
S = (Q-G(L))/V
DO 50 J=1,L3
K = L - J + 1
F(J) = F(J) - S*A(K)
50 CONTINUE
F(L) = -S
60 CONTINUE
RETURN
END

```

```
C CCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCC
C SUBROUTINE FOLD(A,LX,LA,B,LB,Z,LZ)
C -----CONVOLUTION BY FOLDING OF VECTOR A WITH VECTOR B
C INPUT-----
C   A      = ARRAY TO BE CONVOLVED
C   LA     = LENGTH OF A
C   B      = OTHER ARRAY TO BE CONVOLVED
C   LB     = LENGTH OF ARRAY B
C   Z      = ARRAY OF A CONVOLVED WITH B
C   LZ     = LENGTH OF Z = LA+LB-1
C DIMENSION A(2),B(2),Z(101)
C LZ = LA + LB - 1
C CALL ZZERO(Z,LZ)
C DO 10 I=1,LA
C   DO 10 J=1,LB
C     K = I + J - 1
C     Z(K) = A(I)*B(J) + Z(K)
C 10 CONTINUE
C RETURN
C END
C CCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCCC
C SUBROUTINE IMPULS(X,LX,KSPIKE)
C -----MAKE A UNIT IMPULSE IN ARRAY X OF LENGTH LX
C PUT SPIKE AT INDEX KSPIKE
C DIMENSION X(LX)
C CALL ZZERO (X,LX)
C X(KSPIKE) = 1.0
C RETURN
C END
```

```

C-----
SUBROUTINE FFT(X,Y,TABLE,M,LL,ISN)
C-----FFT IS IN-PLACE DISCRETE FOURIER TRANSFORM USING SANDE
C ALGORITHM AND MARKEL FOURING MODIFICATION.
C NOTE THAT THIS SUBROUTINE REQUIRES THAT THE COSINE TABLE
C BE PREVIOUSLY GENERATED. THUS A CALL TO SUBROUTINE COSQT
C IS REQUIRED BEFORE THE CALL TO FFT
C-----
C-----INPUT PARAMETERS
C
X   = ARRAY OF LENGTH 2**M TO HOLD REAL PART OF COMPLEX INPUT
Y   = ARRAY OF LENGTH 2**M TO HOLD IMAGINARY PART OF INPUT
TABLE = ARRAY OF LENGTH N/4+1 WHERE N=2**M TO HOLD THE QUARTER-
        LENGTH COSINE TABLE
M   = INTEGER POWER OF 2 FOR SIZE OF FFT TO PERFORM
        (I.E. N=2**M) MAXIMUM VALUE IS 12 (FOR N OF 4096)
LL  = INTEGER POWER OF 2 OF ACTUAL NUMBER OF NON-ZERO DATA
        POINTS OF INPUT DATA. LL MUST BE .LE. M
        2**M - 2**LL ZEROS ARE APPENDED TO THE FIRST
        2**L DATA POINTS FOR THE DFT. PRUNING ELIMINATES
        THE CALCULATION COSTS FOR THE ADDITIONAL ZEROS
        APPENDING ZEROS IS USEFUL FOR PERFORMING INTERPOLATION
        OF THE DFT IN THE FREQ OR TIME DOMAIN.
ISN  = -1 FOR FORWARD FFT
      = +1 FOR INVERSE FFT
C
C----- OUTPUT PRODUCTS
C
X   = REAL PART OF TRANSFORMED ARRAY
Y   = IMAGINARY PART OF TRANSFORMED ARRAY
C-----NOTE FFT DESTROYS THE INPUT ARRAY BY WRITING THE
      TRANSFORMED ARRAY OVER TOP
C
      DIMENSION X(1096),Y(1096),TABLE(1025),L(12)
      EQUIVALENCE (L12,L(1)),(L11,L(2)),(L10,L(3)),(L9,L(4)),(L8,L(5)),
      $           (L7,L(6)),(L6,L(7)),(L5,L(8)),(L4,L(9)),(L3,L(10)),
      $           (L2,L(11)),(L1,L(12))
C
      N       = 2**M
      ND4     = N/4
      ND4P1  = ND4 + 1
      ND4P2  = ND4P1 + 1
      ND2P2  = ND4 + ND4P2
      LLL    = 2**LL
C
C-----CHECK TO BE SURE THAT COSINE TABLE WAS GENERATED
C
      IF (.NOT.(TABLE(1).EQ. 1.0.AND. TABLE(ND4P1).EQ. 0.0))
      $
      CALL COSQT(M, TABLE)
C
      DO 10 LO=1,M
      LMX = 2** (M-LO)
      LMM = LMX
      LIX = 2*LMX
      ISCL = N/LIX
C
C-----TEST FOR PRUNING

```

```

C
1 IF(LO-M+LL) 1,2,2
2 IMM = LIL
3 DO 9 LM=1,IMM
4   IARG = (IM-1)*ISCL + 1
5   IF(IARG.LE.ND4P1) GOTO 4
6   K1 = ND2P2 - IARG
7   C = -TABLE(K1)
8   K3 = IARG - ND4
9   S = ISN*TABLE(K3)
10  GOTO 6
11  C = TABLE(IARG)
12  K2 = ND4P2 - IARG
13  S = ISN*TABLE(K2)
14  CONTINUE
15  DO 8 LI=LIX,N,LIX
16    J1 = LI - LIX + IM
17    J2 = J1 + LMX
18    T1 = X(J1) - X(J2)
19    T2 = Y(J1) - Y(J2)
20    X(J1) = X(J1) + X(J2)
21    Y(J1) = Y(J1) + Y(J2)
22    X(J2) = C*T1 - S*T2
23    Y(J2) = C*T2 + S*T1
24  CONTINUE
25 9 CONTINUE
26 10 CONTINUE
27 C -----PERFORM BIT REVERSAL
28 C
29 DO 40 J=1,L2
30   L(J) = 1
31   IF(J - M) 31,31,40
32   L(J) = 2*(M+1-J)
33 CONTINUE
34 JN = 1
35 DO 84 J1=1,L1
36   DO 82 J2=J1,L2,L1
37     DO 80 J3=J2,L3,L2
38       DO 78 J4=J3,L4,L3
39         DO 76 J5=J4,L5,L4
40           DO 74 J6=J5,L6,L5
41             DO 72 J7=J6,L7,L6
42               DO 70 J8=J7,L8,L7
43                 DO 68 J9=J8,L9,L8
44                   DO 66 J10=J9,L10,L9
45                     DO 64 J11=J10,L11,L10
46                       DO 62 J12=J11,L12,L11
47                         IF(JN - J12) 51,51,54
48                           R = X(JN)
49                           X(JN) = X(J12)
50                           X(J12) = R
51                           FI = Y(JN)
52                           Y(JN) = Y(J12)
53                           Y(J12) = FI
54                         IF(ISN) 56,56,55
55                         X(J12) = X(J12)/FLOAT(N)
56                         Y(J12) = Y(J12)/FLOAT(N)
57                         JN = JN + 1
58                       CONTINUE
59                     CONTINUE
60                   CONTINUE
61                 CONTINUE
62               CONTINUE
63             CONTINUE
64           CONTINUE
65         CONTINUE
66       CONTINUE
67     CONTINUE
68   CONTINUE
69   CONTINUE
70   CONTINUE
71   CONTINUE
72   CONTINUE
73   CONTINUE
74   CONTINUE
75   CONTINUE
76   CONTINUE
77   CONTINUE
78   CONTINUE
79   CONTINUE
80   CONTINUE
81   CONTINUE
82   CONTINUE
83   CONTINUE
84   CONTINUE
85   RETURN
86 END

```

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