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ANNUAL TECHNICAL REPORT

1 October 1979 - 30 September 1980

BODY AND SURFACE WAVE MODELING

OF OBSERVED SEISMIC EVENTS

Contract No. F49620-77-C-0022

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In Section III, a broad band amplitude study of P and SH waves from deep South American events recorded at WWSN stations in the United States provides constraints on upper mantle variations in attenuation. The prevalent feature in the short and long period bands is that the SH amplitudes show the same regional variations as the P, with the regional variations of short periods being similar, but enhanced relative to the variations at long periods. The amplitudes do not indicate a systematic regional difference between the East Coast and Rocky Mountain provinces, but both short and long periods are enhanced at the Gulf Coast and midwestern stations probably due to thick sedimentary receiver structure.

In Section IV, it is shown that Rayleigh-type Lg propagation in a laterally homogeneous continental crust can be synthesized by adding only a few overtones at periods greater than 2 sec. Under minimal assumptions it is then shown using synthetic seismograms that a wave number analyses of Lg recorded on a linear array of 10 stations covering several hundred kilometers can isolate the different overtones, providing a tool to study both crustal structures and excitation of the overtones at the source.

Section V presents short period P wave amplitude from nuclear explosions in the Soviet Union recorded by WWSN stations in the United States. A well defined amplitude pattern is obtained for each source region. The test sites at northern and southern Novaya Zemlya show a relative amplitude trend of a factor of three across the United States in their respective amplitude patterns. This is in contrast to two sites at Semipalatinsk which are in good relative agreement. A pattern of lateral variation of amplitude in the United States is obtained for a northern azimuth of approach. Stations in western U.S. do not have systematically lower amplitudes than eastern U.S. stations. Lowest amplitudes are found in Golden, Colorado and Albuquerque, New Mexico, a factor of four lower than high amplitude stations.

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1 October 1979 - 30 September 1980

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Division of Geological and Planetary Sciences
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I. Summary

The research performed under this contract during the period 1 April through 30 September 1980 can be divided into three main topics, modeling the excitation of surface waves from complex source regions, body wave amplitudes and attenuation across North America, and a technique for separation of multimode dispersion and excitation of observed vertical Lg using arrays.

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Coupling Near Source Phenomena Into Surface Wave Generation

David G. Harkrider

COUPLING NEAR SOURCE PHENOMENA INTO SURFACE WAVE GENERATION

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ABSTRACT

Theoretical explosion mechanisms for an all azimuth reversal of Rayleigh waves are presented. Linear code methods for modeling nonlinear source region phenomena and techniques for coupling near source phenomena into surface wave generation, such as vertical separation of source radiation, mixed path approximations, and the elastodynamic representation theorem are reviewed. Qualitative and quantitative results obtained using these algorithms, in particular source medium scaling, are discussed.

1. INTRODUCTION

Non-isotropic and "reversed" surface waves from underground events in Russia have emphasized the need for estimating the magnitude and mechanism of these anomalous seismic source region effects. At present the favored explanation is the superposition of a thrust-like double couple earthquake or its tectonic release equivalent.

There are a number of feasible geological and tectonic settings, where the one double couple like source plus explosion assumption is too restrictive. Volume tectonic release in a region of plane compression or tension is an example of where the source must be modeled by two right angle quadrupoles plus isotropic source. This mechanism is disturbing in that there is an isotropic component which can not be separated from the explosion by moment tensor inversion.

Because of the various parameters in the source region which can effect the amplitude and polarity of Rayleigh waves even when the signals have traversed relatively well calibrated propagation paths, this presentation will review some of these effects and present linear methods for modeling nonlinear source regions and techniques for coupling near source phenomena into surface wave generation. In the remainder of this introduction, we will discuss two possible explosion mechanisms for an all azimuth reversal of Rayleigh wave signature.

1.1 Tectonic release or triggering

A substantial amount of research has been done on the effect of superimposing a vertical pure strike slip double couple like mechanism on explosion generated surface waves. This tectonic generation model was felt to be most appropriate to NTS and the French Sahara Test site (Harkrider, 1977, and Rodi et al., 1978).

Another and potentially more troublesome tectonic mechanism is the 45° dipping pure thrust fault or tectonic release due to elastic failure in an equivalent orientation of a pre-stressed field. If strong enough the additional seismic radiation can reverse the Rayleigh wave signature which would be observed for the explosion alone.

In Figure 1, we show the synthetic fundamental Rayleigh waves for an explosion and a pure thrust 45° dipping double couple at a range of 4000 km in a western U.S. model, CIT 109. The two mechanisms are synthesized separately for enough azimuths to determine the radiation pattern for the double couple. The phase of the double couple is found to be reversed with respect to the explosion at all azimuths. The bottom synthetic is a composite seismogram of the two mechanisms. The amplitude reversed double couple (dashed) is plotted on top of the explosion. They are shown with the same initial time, i.e., no shift in time, and demonstrate the nearly perfect reversal in polarity. The peak Airy phase amplitudes are plotted in Figure 2 as a function of azimuth. The amplitudes are scaled such that the P wave moment for both the explosion and the double couple are the same, $M_0 = 10^{25}$ (dyne-cm). This does not imply that their P waves have the same amplitude since this is a strong function of source rise time and was not investigated here.

This thrust double couple mechanism can be considered the sum of two orthogonal vector dipoles of opposite sense. The vertical dipole contributes the azimuthally symmetric part of the radiation pattern and the horizontal dipole normal to the strike contributes a symmetric plus quadrupole part. The relative weights of the two dipoles to the radiation pattern is a function of frequency and source depth. The illustrated

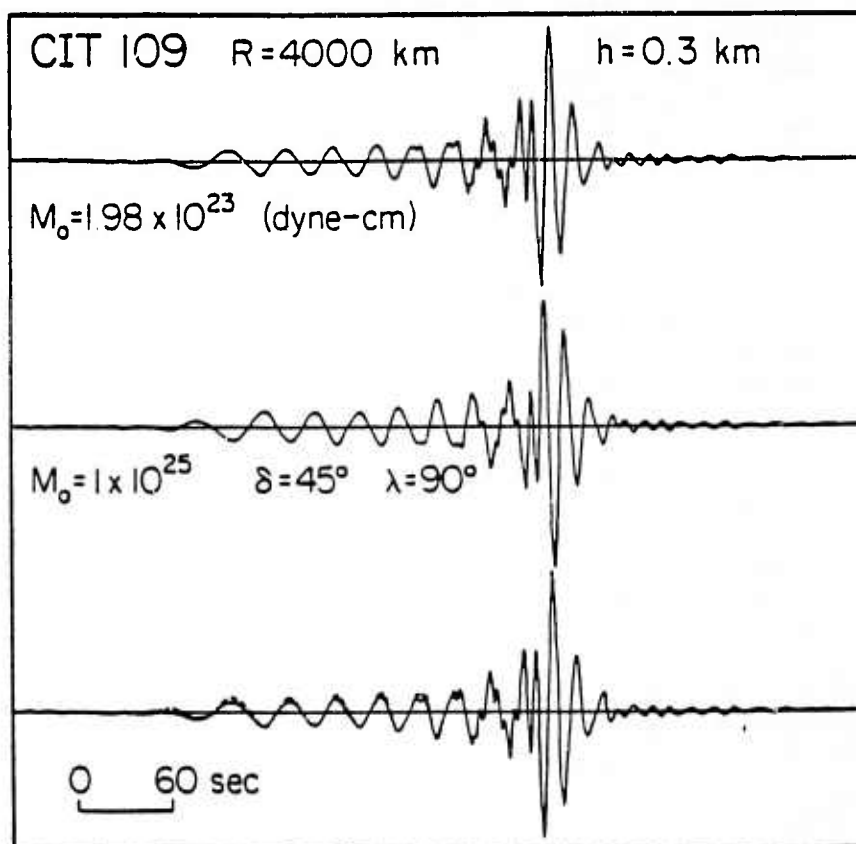


Figure 1. Vertical Rayleigh waves from an explosion and a 45° dipping pure thrust double couple fault singly and superimposed in the bottom trace. The azimuth from the fault strike is 22.5° .

synthetics and radiation pattern are for a source depth of 0.3 km. As source depth is increased from 0.01 km to 5 km in this continental model, the high frequency component of the explosion synthetic is reduced compared to the longer periods but the phase remains unchanged. At source depths less than 1 km, the double couple is out of phase with the explosion at all azimuths. On the other hand, at a source depth of 5 km, the early arriving long periods are always out of phase with the explosion but the later arriving short periods, which include the Airy phase, vary from out of phase at an azimuth of 90° to in phase at 0° which is along the fault strike.

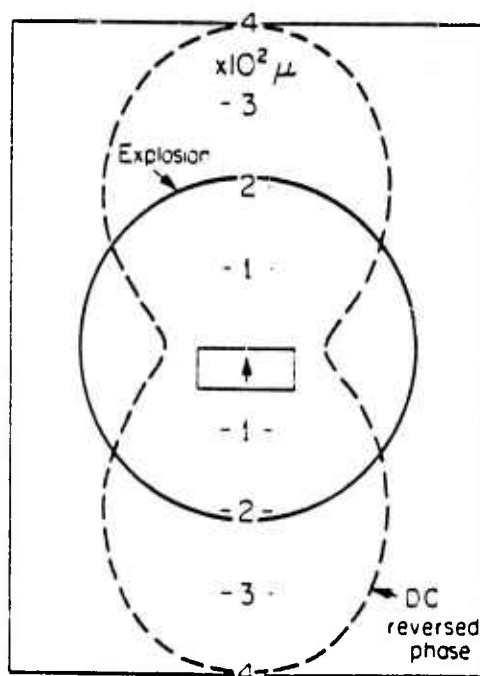


Figure 2. The peak to trough maximum Airy phase Rayleigh wave amplitudes as a function of azimuth for the explosion and double couple.

At shallow source depths, the combination of an explosion with this pure thrust 45° dipping double couple like mechanism can result in a reversal of phase at all azimuths, a quadrupole typical of pure strike slip double couple, or an ideal explosion pattern depending on the relative strengths of the two sources. Local variations away from a pure thrust or the 45° dip can also similarly modify the resultant radiation pattern.

1.2 Degradation of upgoing and downgoing source radiation

In order to investigate the effects of nonlinear regions above and below the source such as spall and cracking, Harkrider and Bache (1981) developed a wave field separation in terms of upgoing and downgoing waves from the source that can be used in most computer programs based on residue theory, steepest descent approximation or wave number integration. This technique allows one to model the resultant source wave in the linear homogeneous regime by source radiation whose history, amplitude, or travel time differs above and below the source point.

When comparing theoretical Rayleigh wave generation for two source models of NTS, Bache *et al.* (1977) found that the polarity of the Rayleigh waves generated by the downgoing radiation in the soft tuff over hard basement model were opposite from those from a source region in which the soft rock-hard rock contrast below the source is absent. If the Rayleigh waves from the upgoing source radiation or the source wave itself were degraded by surface or near surface irregularities or momentum traps then it would be possible to reverse the polarity of Rayleigh waves at all azimuths from explosions in the same general source area. This effect would be similar to that due to the superposition of the tectonic surface wave radiation from a 45° dipping pure thrust fault like mechanism mentioned previously.

In the next section we give a brief derivation of the theoretical aspects of the technique applied to explosions and the depth effect on Rayleigh wave amplitudes when the vertically suppressed explosion sources are used to model directional source effects in simple and realistic earth models.

2. VERTICAL SUPPRESSION OF SEISMIC SOURCE RADIATION

The seismic wave propagation problem is commonly formulated in terms of an integral in the wavenumber domain. Ray theory techniques for solving the problem naturally separate the contribution of waves along different takeoff angles. However, this separation is not obvious in methods where the wavenumber integral is solved directly.

Harkrider and Bache (1981) present a wavenumber formulation where the source discontinuity condition is separated into discontinuity conditions due to up and downgoing source radiation. This separated source "jump" condition can be conveniently used in most computer programs based on residue theory or wavenumber integration. With minor modification it can also be used in codes for free oscillation or normal mode synthetics. The following is an example of the formulation applied to explosion generated surface waves.

2.1 Rayleigh wave explosion source suppression theory

From Harkrider (1964) equation (43), an azimuthally symmetric compression source is given by

$$\bar{\Phi}(r, z) = \begin{cases} \int_0^{\infty} S_{01}^+ e^{-ikr_{\alpha}(z-h)} J_0(kr) dk, & z > h \\ \int_0^{\infty} S_{01}^- e^{-ikr_{\alpha}(h-z)} J_0(kr) dk, & z < h \end{cases} \quad (2.1)$$

where S_{01}^+ is the strength of the downgoing and S_{01}^- is the strength of the upgoing source radiation in wave number space. For a spherically symmetric explosion source

$$S_{01}^+ = S_{01}^- = i \frac{\bar{\Psi}(\omega)}{r_{\alpha}} \quad (2.2)$$

with $\bar{\Psi}(\omega)$ the reduced displacement potential i.e.,

$$\bar{\Psi}(r, z, t) = - \frac{\Psi(t - R/\alpha)}{R} \quad (2.3)$$

The surface vertical displacement for Rayleigh waves is given in Harkrider (1964), equation (79), as

$$\left\{ \bar{w}_0 \right\}_R = \frac{\pi}{k} \frac{N_R^{(1)}}{\left(\frac{\partial F_R}{\partial k} \right)_{\omega}} H_0^{(2)}(kr) \quad (2.4)$$

where

$$N_R^{(1)} = [GN - LH] \left[Y + \frac{K}{L} Z \right] \quad (2.5)$$

and

$$\left[Y + \frac{K}{L} Z \right] = \{ \delta \sigma \bar{y}_1^R + \delta w \bar{y}_2^R - i \delta \tau_R \bar{y}_3^R - i \delta u \bar{y}_4^R \} \quad (2.6)$$

The correspondence between the Saito (1967) and Haskell (Harkrider, 1964) notation is given by

$$\bar{y}_3^R = \left[\frac{\dot{u}^*}{\dot{w}_0} \right] \quad (2.7)$$

$$\bar{y}_1^R = \left[\frac{\ddot{w}}{\dot{w}_0} \right] \quad (2.8)$$

$$\bar{y}_2^R = k \left[\frac{\sigma^*}{\dot{w}_0/c} \right] \quad (2.9)$$

$$\bar{y}_4^R = -k \left[\frac{\tau_R}{\dot{w}_0/c} \right] \quad (2.10)$$

and

$$\dot{x}/c = ikx \quad (2.11)$$

For a dilatation source

$$\delta u = -ik (s_{01}^+ - s_{01}^-) \quad (2.12)$$

$$\delta w = -ikr_\alpha (s_{01}^+ + s_{01}^-) \quad (2.13)$$

$$\delta \sigma = k^2 c^2 \rho (\gamma - 1) (s_{01}^+ - s_{01}^-) \quad (2.14)$$

$$\delta \tau_R = -2k^2 \mu r_\alpha (s_{01}^+ + s_{01}^-) \quad (2.15)$$

where

$$\gamma = 2 \frac{\dot{s}^2}{c^2} \quad (2.16)$$

and

$$r_\alpha = (c^2/\alpha^2 - 1)^{1/2} \quad (2.17)$$

(Harkrider, 1964, equations 37)

In order to eliminate the Rayleigh wave contribution from the downgoing source wave, we set the downgoing source coefficient, s_{01}^+ , equal to zero and obtain

$$\left\{ \bar{w}_0 \right\}_R = -2\pi \mu k r_\alpha s_{01}^- [K_R + \frac{i}{2\mu r_\alpha} M_R] \frac{A_R}{H_0} H_0^{(2)}(kr) \quad (2.18)$$

where

$$\frac{A_R}{H_0} = \frac{[G^* N - L^* H]}{\left(\frac{\partial F_R}{\partial k} \right)_\omega} \quad (2.19)$$

$$K_R = \bar{y}_3^R(h) - \frac{1}{2\mu k} \bar{y}_2^R(h) \quad (2.20)$$

$$M_R = \rho c^2 (\gamma - 1) \bar{y}_1^R(h) - \frac{1}{k} \bar{y}_4^R(h) \quad (2.21)$$

or in terms of Saito (1967)

$$\underline{A}_R = \frac{1}{2cU\underline{I}_1^R} \quad (2.22)$$

and

$$\underline{I}_1^R = \int_0^\infty \rho [(\bar{Y}_1^R)^2 + (\bar{Y}_2^R)^2] dz \quad (2.23)$$

Similarly suppressing the upgoing source radiation by setting $\underline{S}_{01}$ equal to zero and combining the two results we have

$$\left\{ \bar{w}_0 \right\}_R = -2\pi k i \bar{\psi}(\omega) \underline{A}_R H_0^{(2)}(kr) \{ (2 - |\Gamma|) K_R - 1 \frac{\Gamma}{2ur_\alpha} M_R \} \quad (2.24)$$

where $\Gamma = 1$: All upgoing source wave suppressed

$\Gamma = 0$: Total source

$\Gamma = -1$: All downgoing source wave suppressed

Γ can take any value in the range $[-1, 1]$ and represents the proportion of up or downgoing source radiation suppressed. For $\Gamma = 0$ we have the result given by Harkrider, 1964, Hudson, 1969, Douglas et al., 1971 and Bache et al., 1978a, among others.

Even though this formulation has been restricted to Rayleigh waves generated by explosion sources, it is easily extended to more general seismic sources imbedded in vertically inhomogeneous media. Expressions have also been obtained in terms of the multipole source coefficient representation of finite sources (Bache and Harkrider, 1976, and Harkrider and Archambeau, 1981).

From equation (2.24), we see that separating the elastic field into that generated by the upgoing and downgoing source radiation introduces a square-root singularity, $1/r_\alpha$, into the frequency domain solution which is not found in the total source field relation $\Gamma = 0$. This is also true for the more general source type expressions. For residue or normal mode solutions, the integratable singularity occurs when the horizontal phase velocity is equal to the source depth body wave velocity corresponding to the type of source radiation. For our explosion source it is the compressional velocity at the source depth.

For the upgoing source wave formulation the jump or source condition in a layer over a half-space is exactly equivalent to satisfying the free surface with the layer homogeneous solutions plus the source term and the lower interface with only the homogeneous solutions. For the downgoing source wave formulation, the source term is incorporated only with the lower interface boundary conditions. When viewed in this light, it is not as

apparently straight forward as identifying directional source terms in the multilayered formulation. Because of this we devised a few tests as to the validity and effectiveness of this technique.

The first is analytic and demonstrates that for downgoing source waves from a source located in the half-space below the layers, no Rayleigh or Love waves are generated. On the other hand, the upgoing waves in the same example, yield the total surface wave solutions. The details are given in Harkrider and Bache (1981).

The second is numerical. Bache and Harkrider (1976) presented expressions for computing the steeply emergent body waves exiting from the base of a model in which a generalized source is imbedded in a stack of plane elastic layers. The modified multipole source coefficient relations make it possible to use this technique to calculate the P-wave displacements at a depth of 40 km in the mantle for an explosion imbedded in a simple one layer crust-mantle model. The results are shown in Figure 3. The synthetics due to the upper, lower and total source radiation for a homogeneous half-space are shown in the top box. In the middle box, we show the synthetics for a one layer model. At the bottom we have sketched the ray paths corresponding to the arrivals labeled on the synthetics. The separation of arrivals into those initially leaving the source in upward or downward directions is excellent.

Since the steepest descent method is essentially a plane wave calculation at the dominant wave number it is not surprising that the implicit nulling of source coefficients in the wave number integrands result in good separation. It is primarily a test of the algebra and resulting numerical code. The third test is a numerical comparison of the mode theory separation using generalized ray seismograms as standards (Harkrider and Helmberger, 1981).

Isolating the contribution from upgoing and downgoing source radiation to the displacement field for SH waves trapped in a plane layered half-space is straight forward using generalized rays. SH displacement fields are especially appropriate since their total contribution is from homogeneous shear waves and not interface or surface waves. If the source and receiver are separated by distance large enough for the major contribution to be interpreted as from multiple reflected rays beyond critical angle, then there should be a close correspondence between generalized rays and summed locked mode solutions.

In Figure 4, we show comparisons between the summed modes and the generalized ray techniques for azimuthal seismograms

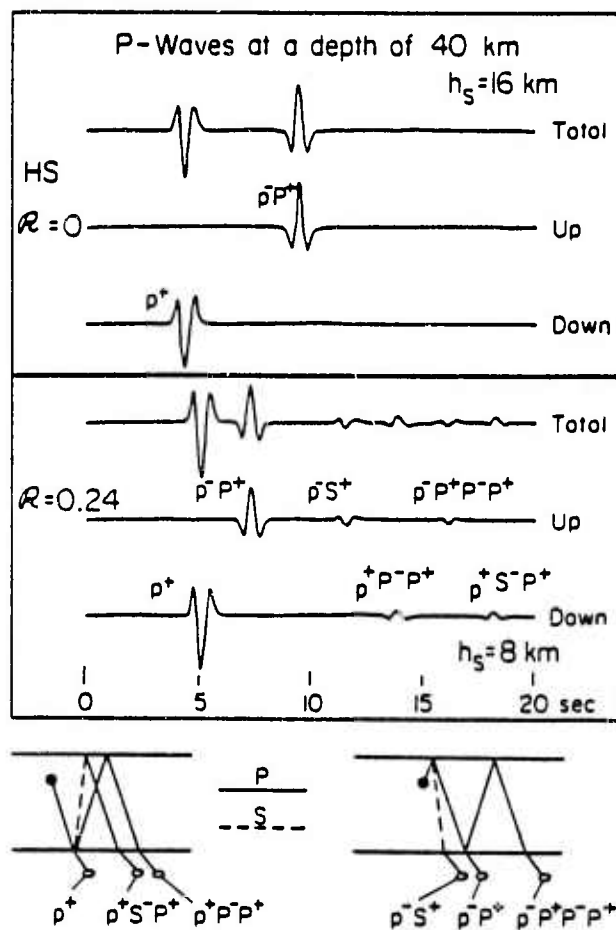


Figure 3. Steepest descent P-wave arrivals in a half-space and a one layer half-space at a depth of 40 km. Raypath diagrams at figure bottom.

excited by the upgoing, downgoing and total SH source radiation in a single layer over a half-space. The layer thickness is 32 km and the source is a horizontal pure shear dislocation at a depth of 24 km and a range of 800 km from the surface receiver. The farfield source time history is a trapezoid with $\delta t_1 = \delta t_2 = \delta t_3 = 0.5$ sec. The seismograms on the left are mode sums and on the right are generalized rays. The top, middle and bottom traces are seismograms excited by the upgoing downgoing, and total source radiation, respectively. All the seismograms in Figure 4 are on the same time and displacement scales. Considering that only six modes were used in the sums and that the time spacing was different for the two techniques, the agreement is

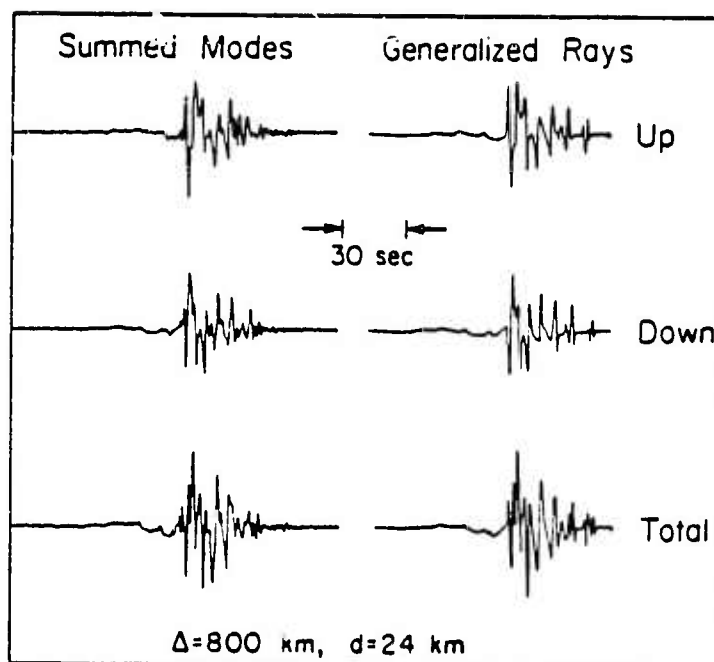


Figure 4. SH wave comparison between summed modes and generalized rays demonstrating vertical separation of source radiation in mode sum.

exceptionally good especially in phase arrival time and overall amplitude.

An interesting feature of this particular decomposition is that the SH amplitudes due to the upgoing and downgoing source radiation are essentially equal to those due to the total source even though their respective seismic energies are one half that of the total source radiation. The partition of energy is evidenced by the difference in apparent frequency in the seismograms. The effect of summing the two components is to approximately increase the number of peaks and troughs by a factor of two. Thus a better estimate of the relative seismic energies in this example is obtained by dividing the peak amplitude of each seismogram by its apparent period.

For this SH source decomposition, the branch point singularity occurs at a phase velocity equal to the shear velocity of the upper or source layer. Thus a limitation of this particular

test is that the source branch point is at infinite frequency and is not excited by the band limited source. As such, we have no indication of its effect in more general layered problems where the branch point may occur at frequencies well within the source or media response band.

2.2 Theoretical depth effect on explosion generated Rayleigh waves

One layer half-space models. In order to gain some physical insight into the generation of fundamental Rayleigh waves by the upgoing or downgoing source radiation, we carried out a parametric model study for explosive sources in a one layer model over a half-space.

The upper medium is the same for all of the one layer model results presented here. The compressional and shear velocities of the lower medium for all four of the models were chosen so as to have a representative range of reflection coefficients at the lower interface. The ratio of compressional to shear velocities and thus the Poisson ratio was kept the same for all lower elastic media. Since this Poisson ratio differed only slightly from that of the upper medium, the normal incident shear wave reflection coefficient, R , distinguishing each model, is essentially the same as the compressional wave coefficient.

The layer thickness is 32 km in all models and the $R = 0.24$ model has been frequently used as a one layer continental crust over mantle model. The lower medium density for all but the negative reflection coefficient model was 3.4 gm/cm^3 . In order to obtain a negative reflection coefficient model without drastically changing the dispersion and thus the shape of the Rayleigh wave, we use the same velocities as the $R = 0.24$ model but reduced the lower medium density to 1.3 gm/cm^3 . The elastic model parameters are given in the Table.

For comparison purposes, we include the simple half-space results for each of the five elastic materials used in the parametric reflection coefficient models. The spectral characteristics for all nine simple propagation models are shown in Figures 5 and 6. The phase, c , and group, U , velocities as a function of period, T , for the fundamental Rayleigh mode of the four dispersive models are shown in Figure 5. Their long period asymptotes corresponding to their lower half-space Rayleigh velocities are notated as HS1, HS2 and HS3 at the right hand edge of the figure. Since only their half-space densities differ, models $R = 0.24$ and $R = -0.24$ both have the same long period asymptote HS2. For the limited period range used, the $R = -0.24$ model is still significantly different from its asymptotic value. All four layer models have the same short period asymptote, HS0, the Rayleigh velocity of their common upper layer medium.

Table. Elastic model parameters

	R	D	α km/sec	β km/sec	ρ gm/cm ³	σ	V_R km/sec
HS0		∞	6.2	3.5	2.7	.266	3.227
		32.	6.2	3.5	2.7	.266	3.227
	0.15						
HS1		∞	6.83	3.75	3.4	.285	3.47
		32.	6.2	3.5	2.7	.266	3.227
	0.24						
HS2		∞	8.2	4.5	3.4	.285	4.16
		32.	6.2	3.5	2.7	.266	3.227
	0.40						
HS3		∞	11.83	6.49	3.4	.285	6.00
		32.	6.2	3.5	2.7	.266	3.227
	-0.24						
HSR		∞	8.2	4.5	1.3	.285	4.16
		32.	6.2	3.5	2.7	.266	3.227

The amplitude responses A_R of the four dispersive models are shown as solid lines in Figure 6. The asymptotic model values are shown as dashed curves. These half-space responses are inversely proportional to period, T , and differ only by a multiplicative constant. A closed form expression for this factor as a function of density, Rayleigh and body velocities can be found in Harkrider (1970) and Harkrider *et al.* (1974).

For two materials with the same Poisson ratio, the ratio of their amplitude responses is given by

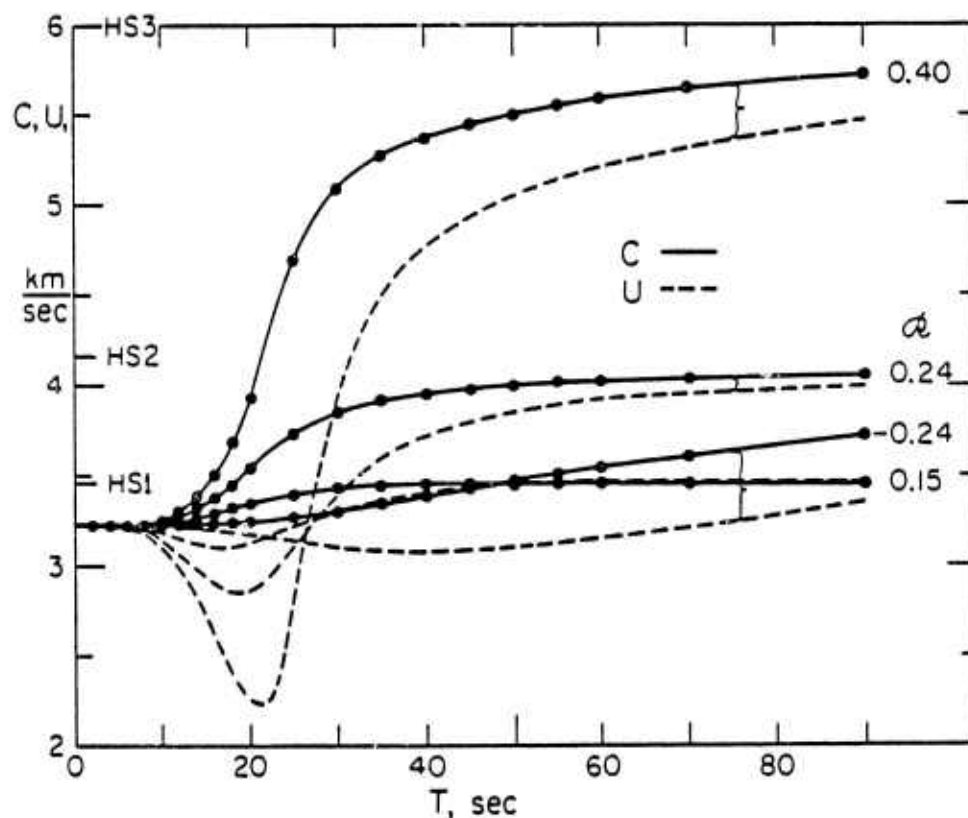


Figure 5. Phase, c , and group, U , velocity dispersion versus period, T , for the one layer half-space models. The half-space Rayleigh velocities for HS1, HS2, and HS3 are shown at left margin.

$$\left(\frac{A_{R1}}{A_{R2}}\right) = \left(\frac{\rho_2 V_2^3}{\rho_1 V_1^3}\right) \quad (2.25)$$

where V is either a body or Rayleigh velocity. Since the Poisson ratio of HSO is nearly the same as the rest of the materials, all of the dashed curves in Figure 6 are approximately related by the above. Since half-spaces HSR and HS2 have the same elastic velocities their curves differ only by the ratio of their densities.

As noted for the velocities, the $R = -0.24$ model is still appreciably different from its long period asymptote HSR at

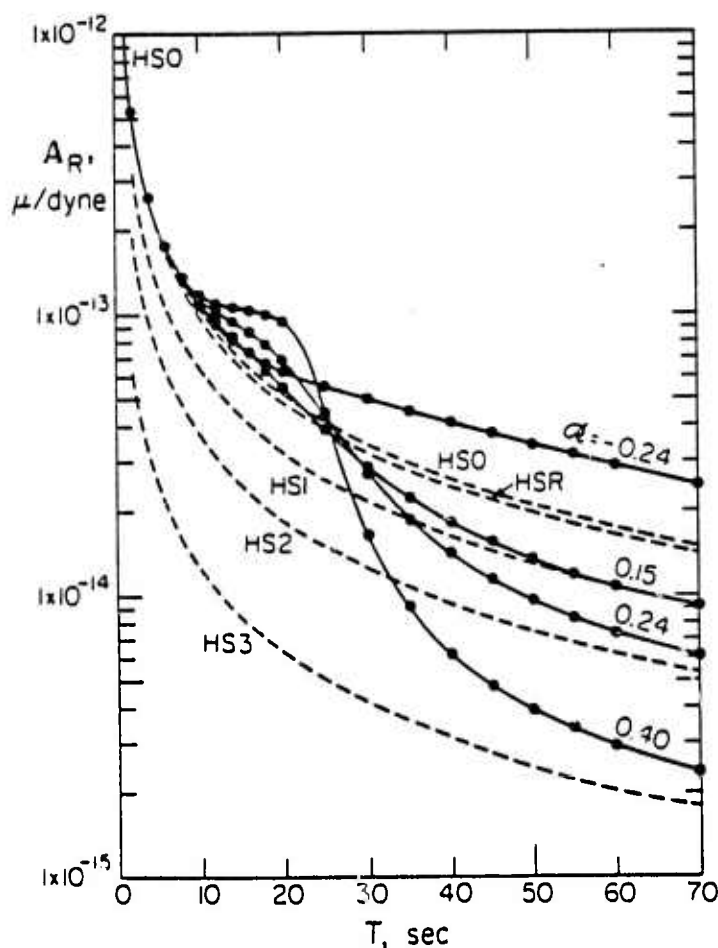


Figure 6. Amplitude response, A_R , for the half-space and one layer half-space models as a function of period, T , in microns/dyne.

70 seconds. For periods shorter than 6 seconds, all the dispersive models are essentially the same as their short period asymptotic model, HS0.

Even though the source induced branch point singularity is integratable, we used the FFT algorithm for generating our time domain synthetics and we were not sure of its applicability to integrands of this type. For these models the Rayleigh wave phase velocities are all less than the layer surface P wave

velocity and thus there is no frequency domain singularity for their fundamental Rayleigh wave mode.

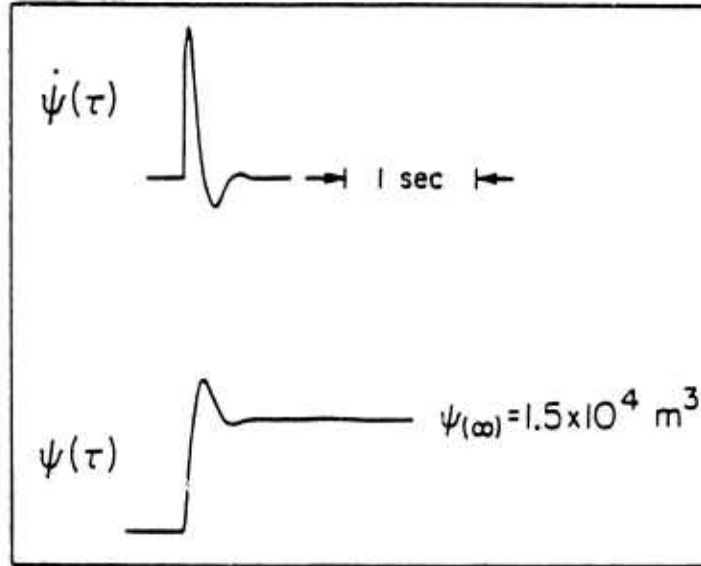


Figure 7. Reduced velocity and displacement potentials for the explosion source model.

For our parametric time domain synthetics, we used the reduced displacement potential, $\Psi(\tau)$, shown in Figure 7 as the explosion time history. The displacement, U_R , and stress σ_{RR} , fields for the direct source P wave at a spherical distance, R , are given by

$$U_R = \frac{1}{\alpha R} \dot{\Psi}(\tau) + \frac{1}{2} \frac{\Psi(\tau)}{R} \quad ; \quad \tau = t - R/\alpha \quad (2.26)$$

$$\sigma_{RR} = -\rho \frac{\ddot{\Psi}(\tau)}{R} - \frac{4\mu}{R} U_R(\tau) \quad (2.27)$$

where

$$\dot{\Psi}(\tau) = \frac{d}{d\tau} \Psi(\tau) \quad , \quad \ddot{\Psi}(\tau) = \frac{d^2}{d\tau^2} \Psi(\tau) \quad (2.28)$$

and the reduced velocity potential, $\dot{\Psi}(\tau)$, is also shown in Figure 7. This particular $\Psi(\tau)$ is one of many resulting from finite

difference code calculations for source materials appropriate for the Nevada Test Site (Bache et al., 1975). For the periods of interest, the amplitude spectra of the reduced velocity potential, $\dot{\Psi}(\omega)$, is essentially equal to a constant $\dot{\Psi}(\infty)$, equal to $1.5 \times 10^4 \text{ m}^3$. The far-field whole space displacements and stresses are given by the first terms in equations (2.26) and (2.27). The far-field displacement is then given by $\dot{\Psi}(\tau)$ in the upper layers, at a distance of 160 km, the peak amplitude is 250 microns (μ).

Since $\dot{\Psi}(\tau)$ is the same for all the parametric depth and model calculations, materials with the same velocities have the same far-field source displacement histories and materials with the same density have the same far-field source stress histories. The latter is the case for all the one layer calculations at a given depth with the one exception of a source in the half-space of the $R = -0.24$ model.

Since we were interested in determining the amplitude as well as signal character of Rayleigh waves generated by upgoing and downgoing source radiation as a function of source depth, we wanted the shape of our Rayleigh signals to remain reasonably unchanged over a wide range of depths. For meaningful absolute amplitude comparisons between our parametric models, we decided to use the amplitudes measured at ranges near enough to reduce the effects of the different degrees of geometric dispersion in the layer models. The distance used for the depth effect figures is 160 km, which is well beyond critical distance for a critically reflected P wave.

Since the dominant periods for the surface Rayleigh displacement are also a strong function of depth, it was decided to include the effect of an instrument in order to realistically band limit the Rayleigh waves. For our amplitude-depth figures the amplitudes are the maximum peak to trough values (PP) of Rayleigh waves measured through a long period LRSM seismograph response. Unless stated otherwise, the following figures in this section are for synthetics which include this instrument.

The depth effect on the dominant period of the surface vertical Rayleigh wave in a homogeneous half-space is shown in Figure 8 with and without a long period WWSN instrument response. The effect with the LRSM-LP response is shown at the left of Figure 10. The half-space, HSO, is of the same material as the upper layer medium in the one layer model. In some figures this half-space is labeled $R = 0$. For all illustrated Rayleigh synthetics, upward displacement is toward the bottom of the figure with the exception of those which include a WWSN-LP response.

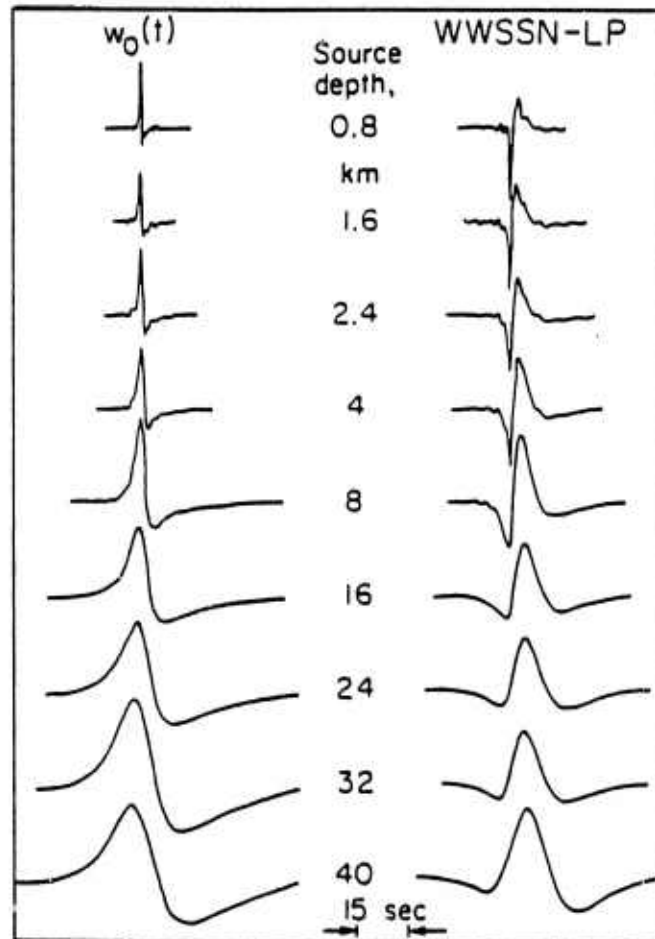


Figure 8. Vertical Rayleigh wave displacements on a half-space with and without WWSSN-LP instrument for selected source depths.

The explosion spectral source depth, h , dependence for a half-space is given by

$$K_R(h) = \frac{(\gamma - 1)}{\gamma r_\alpha^*} \exp[\omega r_\alpha^* h / V_R] \quad (2.29)$$

where

$$r_\alpha^* = - (1 - V_R^2 / \alpha^2)^{1/2} \quad (2.30)$$

which acts as a low pass exponential filter. The attenuation of high frequencies increases with increasing source depth. For a surface source, the filter is flat for all frequencies. The change in shape with increasing depth is most discernable for the Rayleigh wave synthetics without an instrument, e.g., the left-hand side (LHS) of Figure 8, and is most rapid between the near surface, 0.8 and 8.0 km, depths.

The change of dominant period and shape at the near surface depths is slightly less rapid as seen through the WWSSN-LP instrument and is negligible for the $R = 0$, LRSM-LP synthetics in Figure 10. The characteristics of the LRSM-LP instrument were deemed desirable in that the maximum peak to trough value occurred at or near the same phase as the source depth was varied. This led to less scatter in amplitude versus depth plots for all the models including the one layer models.

Rayleigh waves excited by step RDP explosions in half-spaces of the same Poisson ratio are invariant when scaled in time as Vt , where V is one of the three intrinsic velocities. Their shapes are identical and differ only in absolute amplitude. Except for the spectral depth factor $K_R(h)$, this would be true for all half-spaces when scaled by $V_R t$. The inclusion of the instrumental responses changes this, but it was observed for all our half-space synthetics that the dominant period decreased as the model Rayleigh velocity V_R was increased. The maximum Rayleigh amplitudes for the LRSM-LP half-space synthetics as a function of source depth are plotted in Figure 9.

For half-spaces with the same Poisson ratio, the explosion spectra is proportional, $(\sim)$, to the elastic constants as

$$\left\{ \bar{w}_0 \right\}_R \sim \frac{\mu}{\rho} \frac{\Psi(\infty)}{V_R^3} K_R(h) (\omega/V_R)^{1/2} \quad (2.31)$$

Thus for near surface sources and by keeping $\Psi(\infty)$ the same in each source media

$$\left\{ \bar{w}_0 \right\}_R \sim (1/\beta)^{3/2} \quad (2.32)$$

This dependence is essentially the same as that deduced by Marshall (1970) and is the approximate relation between time domain amplitudes shown in Figure 9 at a source depth of 0.8 km. This also predicts the shallow depth values of half-space HS0 which has only a slightly different Poisson ratio. Since half-spaces HS2 and HSR have the same Poisson ratio and body velocities, their explosion seismograms are identical at all ranges and

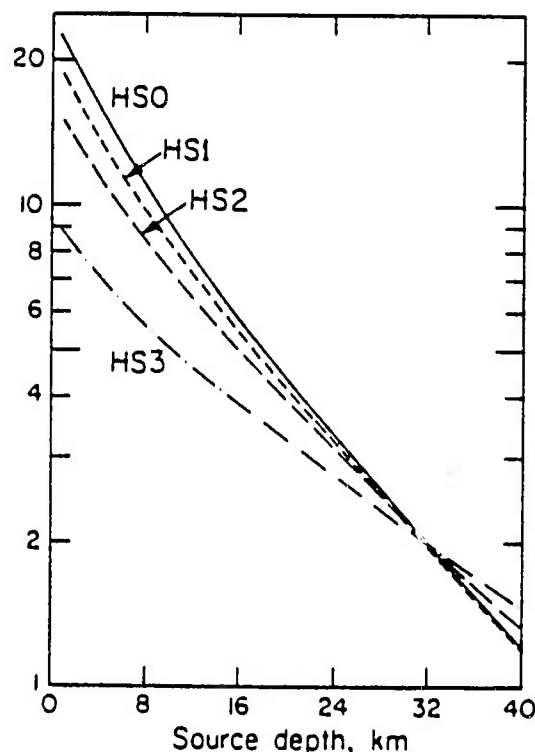


Figure 9. Maximum peak to trough, -PP, Rayleigh wave amplitude on the half-space models in microns as a function of source depth.

depths. All the depth curves in Figure 9 show the expected almost exponential relation predicted by spectral equation (2.29). The curves deviate from straight lines because of the changing dominant period with changing depth. The use of the narrow band LRSM-LP instrument reduces this deviation.

Rayleigh wave LRSM-LP synthetics for the standard one layer crust-mantle model, $R = 0.24$ at the two distances 160 and 1000 km are shown in Figure 10 for a variety of source depths. The corresponding synthetics for the extreme dispersion model $R = 0.40$ can be seen on the LHS of Figures 14 and 15. The synthetics for both models at the 160 km distance are quite similar. However, at the distance of 1000 km, their signatures are very dissimilar as one would expect because of the difference in dispersion. As in the half-space models, the effect of increasing source depth is to progressively filter out the higher frequencies.

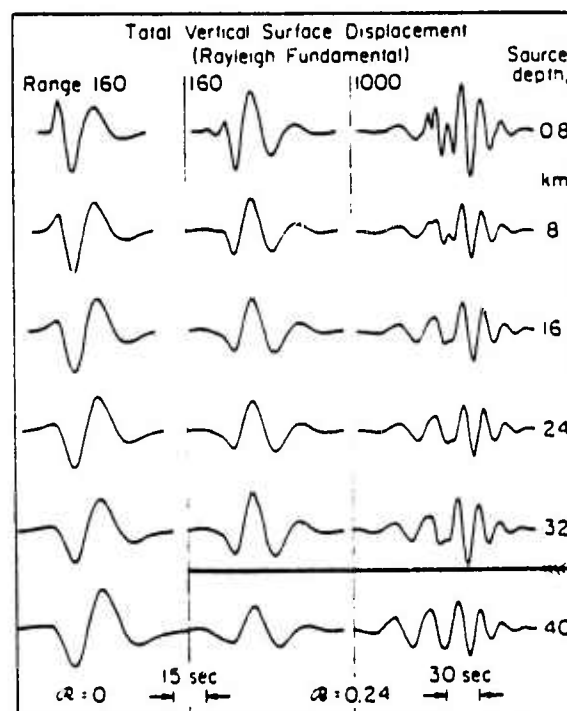


Figure 10. Rayleigh wave LRSM-LP seismograms on the one layer half-space, $R = 0.24$, and the half-space of the same surface material at two distances and six source depths.

The LRSM-LP maximum PP values for the one layer models as a function of depth are shown in Figure 11. For comparison purposes we show the half-space values of model $R = 0$ or HSO. As previously mentioned this is a half-space of the same material as the upper layer in all the one layer models. Referring to Figure 9, HSO has the largest amplitudes of all the half-space models for source depths from the surface down to 32 km, which is the layer thickness of the one layer models. Thus all the one layer models have larger explosion generated LRSM-LP amplitudes than any individual half-space model. This is also generally true for WWSSN synthetics if we ignore the scatter at source depths less than 16 km.

The increase in amplitude with source depth in the lower part of the layer for the positive reflection coefficient models is due to the contribution of the downgoing source radiation to the generation of the fundamental Rayleigh wave and, of course,

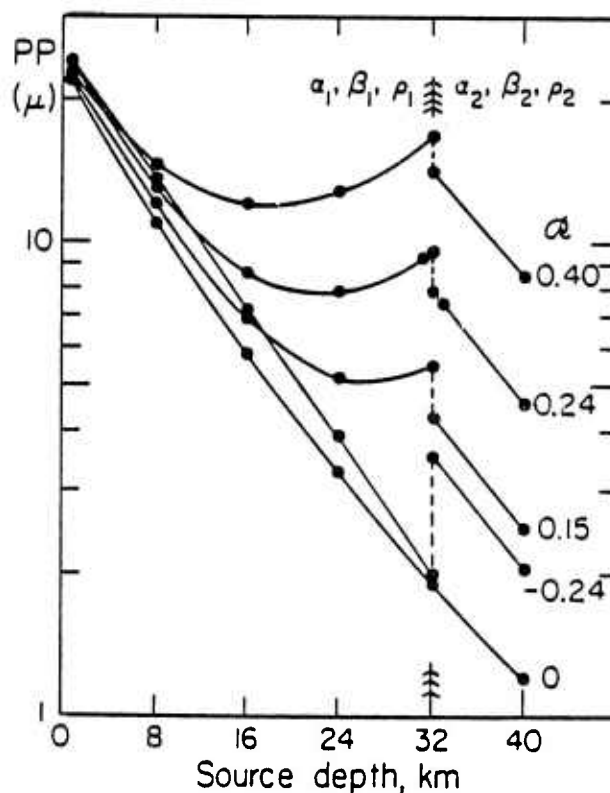


Figure 11. Maximum LRSM-LP Rayleigh amplitudes, PP , for the one layer half-space models as a function of source depth.

is not present in the half-space model. This effect will be discussed in more detail later. Below the layer interface the amplitudes show the expected exponential decay of a half-space as source depth is increased. This decay rate is essentially the same for all one layer models and is practically indistinguishable from that of HSO or HSI at those depths.

The discontinuity in source depth effect as the source crosses the bottom interface is due to the spectral discontinuity in $\mu K_R(h)$ at this interface. The magnitude and sign of this jump is frequency dependent and in the synthetics depends on the dominant spectral frequencies. The dashed discontinuities in Figure 11 roughly correspond to spectral values between 20 and 30 sec.

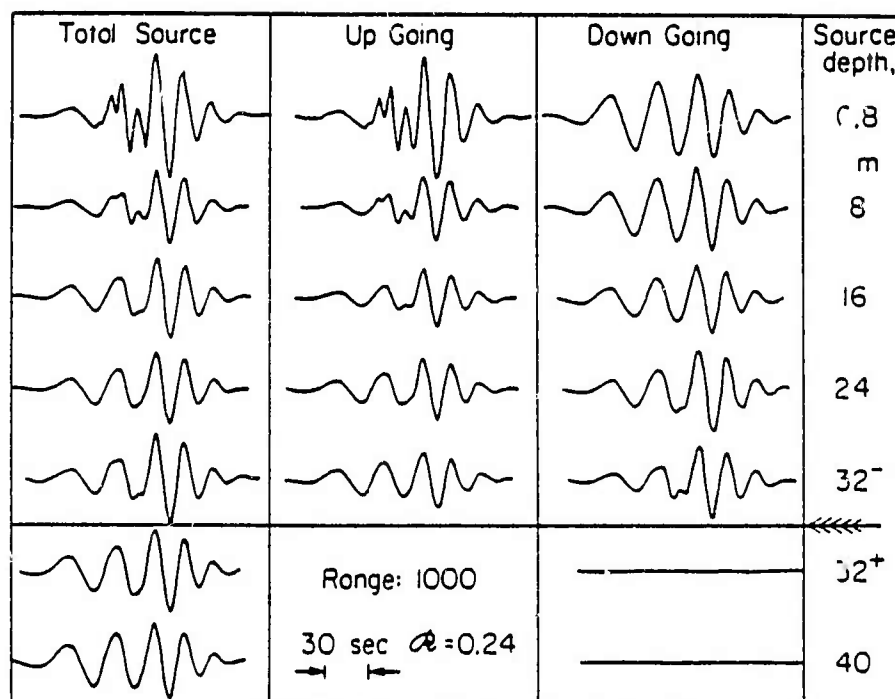


Figure 13. Rayleigh wave LRSM-LP seismograms at a range of 1000 km for the $R = 0.24$ model demonstrating the contribution of the upgoing and downgoing source radiation at a sequence of increasing source depths.

the synthetics are a measure of the accuracy in computing the four eigenfunctions at the top of the terminating half-space.

From Figures 12-15, we see that for models $R = 0.24$ and $R = 0.40$, the greater the source depth the less the high frequency content in the "total" and "upgoing" synthetics. Conversely, an increase in source depth increases the high frequencies in the "downgoing" synthetic with a maximum in high frequency content for a source at the bottom interface. This maximum high frequency content in synthetics for the downgoing source at the interface appeared to be the same as that for the upgoing source at some intermediate depth. This observation was true for all our one layer models.

The maximum PP amplitudes due to the total and separated source radiation for the three positive reflection coefficient models are shown in Figures 16, 17 and 18 as a function of source depth.

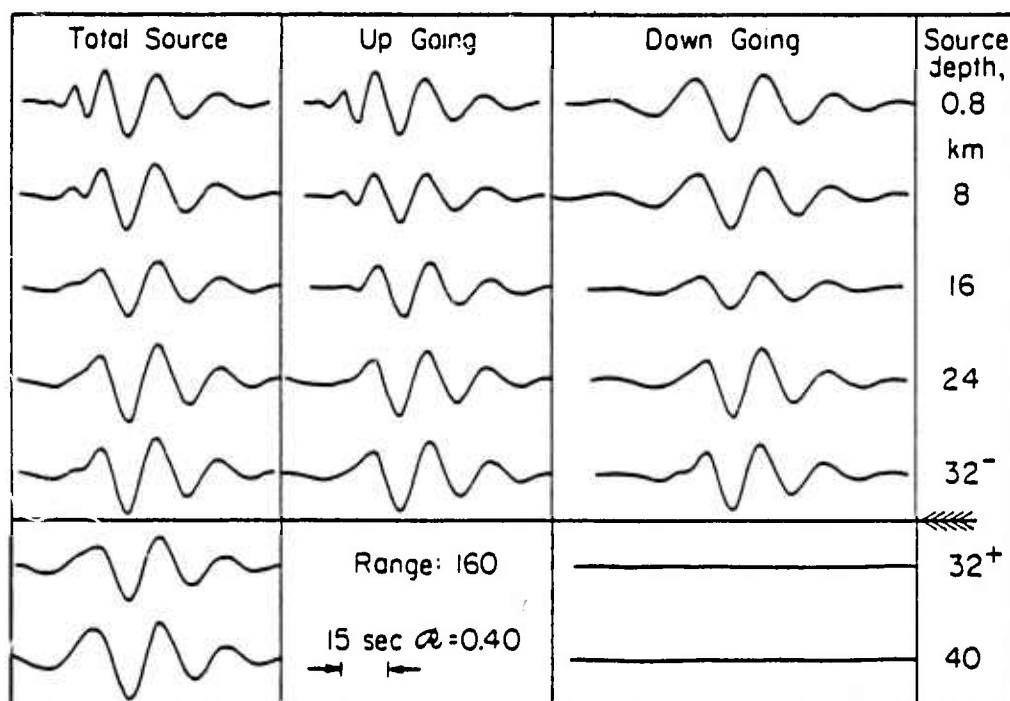


Figure 14. Rayleigh wave LRSM-LP seismograms at a range of 160 km for the $R = 0.40$ model demonstrating the contribution of the upgoing and downgoing source radiation at a sequence of increasing source depths.

In all three figures the amplitudes excited by the upgoing source radiation, labeled U, from a source in the upper layer are very close to those observed for sources in a half-space with the same elastic properties of the upper layer, i.e., HS0. These half-space amplitudes are shown in Figures 16, 17, 18 and 20 as dashed curves and labeled HS. Even though the difference is slight, the amplitude enhancement due to the lower half-space appears to be proportional to the value of the reflection coefficient, R .

For the contribution due to the downgoing source radiation, labeled D, the effect of R is very noticeable. Even though this is not basically a multiply reflected trapped wave phenomena like higher mode P-SV generation; the larger the positive reflection coefficient, the greater the contribution of the downgoing source wave to the excitation of the fundamental P-SV or Rayleigh wave mode. It is the contribution of the downgoing source radiation which cause the total source excitation, labeled T, to eventually

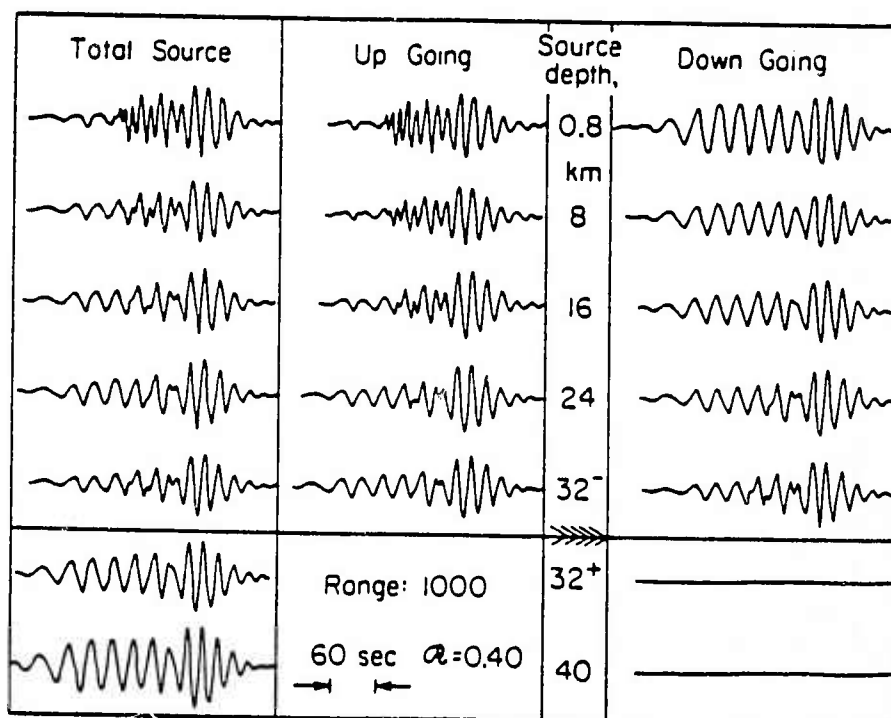


Figure 15. Rayleigh wave LRSM-LP seismograms at a range of 1000 km for the $R = 0.40$ model demonstrating the contribution of the upgoing and downgoing source radiation at a sequence of increasing source depths.

increase with increasing source depth in the upper layer.

As mentioned earlier, the high frequency content in the "downgoing synthetics" and the "upgoing synthetics" appeared to be similar at some intermediate source depth in the upper layer. This depth corresponds to the crossover depth of equal amplitude excitation of upgoing and downgoing source radiation, which is determined by the intersection of the U and D labeled curves in Figures 16, 17 and 18.

Since the positive reflection coefficients enhanced the excitation of the Rayleigh wave in one layer models over the half-space model and this enhancement appeared to be proportional to its value, we wondered what the effect of a negative reflection coefficient would be. In particular, we wanted to know if a negative value would reverse the polarity of the downgoing

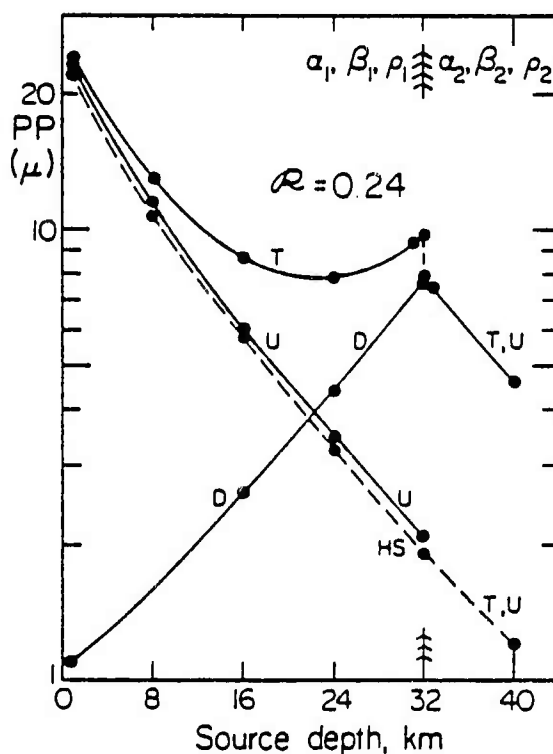


Figure 16. Maximum LRSM-LP Rayleigh amplitudes, PP, for the one layer half-space model, $R = 0.24$, versus source depth due to the upgoing, U, downgoing, D, and total, T, source radiation.

contribution so that the combination of the two Rayleigh waves would result in a completely different source depth behavior than the positive coefficient models.

As mentioned earlier, the negative R model has the same velocity structure as the standard one layer model, $R = 0.24$. The negative R value is obtained by a density reversal at depth. The upper medium density is the same as the positive models. The lower half-space density is reduced from the positive model values of 3.4 to a new value of 1.3 gm/cm^3 .

Rayleigh wave synthetics for the two models at a range of 160 km and for our standard source depth set as observed with the WWSSN-LP and LRSM-LP seismograph systems are shown in Figure 19. The Rayleigh wave signatures are different for the two models but not drastically so. Without the side by side

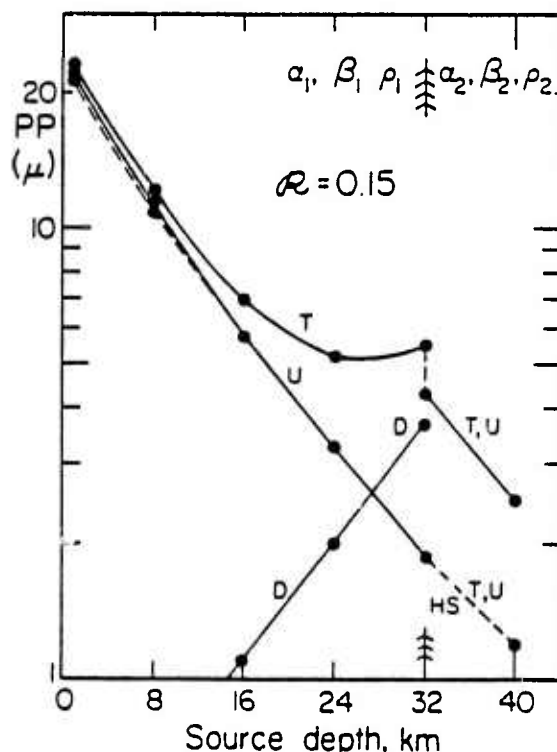


Figure 17. Maximum LRSM-LP Rayleigh amplitudes, PP, for the one layer half-space model, $R = 0.15$, versus source depth due to the upgoing, U, downgoing, D, and total, T, source radiation.

comparison, one would be hard pressed to delineate the differences.

The major differences in the two models is in the amplitude excitation for the total and separated source radiation as a function of source depth (Figure 20). The $R = -0.24$ model has a common feature with the positive R models in that the upgoing source radiation contribution decreases and the downgoing contribution increases with the increasing source depth in the layer. There are two notable differences. One is in the magnitude of the separated contributions and the other is that the upgoing contribution is greater than the total source contribution. The downgoing contribution is at least an order of magnitude smaller than the downgoing contribution to the positive R model. The upgoing contribution is significantly greater than the upgoing contribution to half-spaces and positive reflection coefficient models.

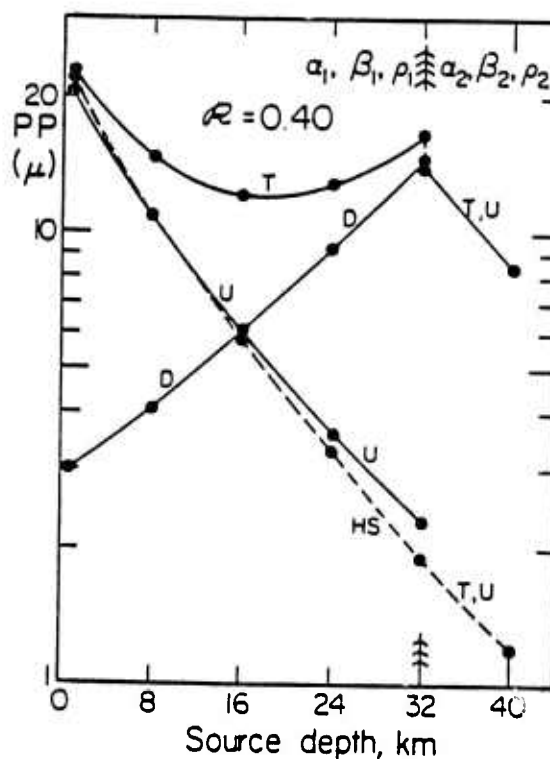


Figure 18. Maximum LRSM-LP Rayleigh amplitudes, PP, for the one layer half-space model, $R = 0.40$, versus source depth due to the upgoing, U, downgoing, D, and total, T, source radiation.

The postulated reversal in polarity between the separated contributions indeed occurred for this negative R model. This reversal caused the total source contribution to be less than the upgoing with a resulting difference in amplitude trend for increasing source depth with respect to the positive models. The two Rayleigh synthetics due to the vertically separated source radiation are shown in Figure 21. As before, the synthetics are scaled for best illustration.

The negative reflection coefficient model was used only to determine if it was possible to reverse the polarity between the two separated source Rayleigh waves. No further investigation of this class of models was undertaken since one layer negative R models are unrealistic earth models.

Because of the source depth behavior of all models,

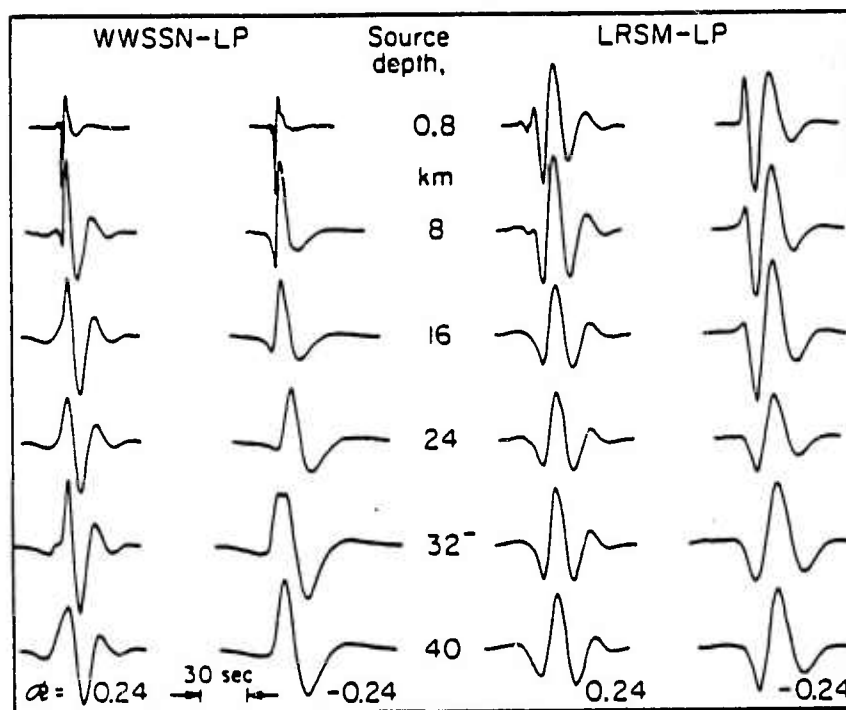


Figure 19. Rayleigh wave WWSSN-LP and LRSM-LP seismograms for models $R = 0.24$ and $R = -0.24$ at a range of 160 km and selected source depths.

especially the amplitude and frequency content of the Rayleigh synthetics as the separated sources approached the interface from above and below, we concluded that in one layer models at short ranges there are two fundamental Rayleigh waves. These upper layer surface waves are primarily generated by the interaction of the free surface with the upgoing source wave and of the returning downgoing source wave after reflection from the lower interface. The strength and polarity of the interface controlled Rayleigh wave depends on the sign and magnitude of the reflection coefficient. Although their effect was not isolated, it was felt that the contribution due to later multiple reflections would be reduced by their decrease in wave front curvature at the top and bottom of the wave guide.

NTS source region models. In the previous section, we restricted our discussion to that part of the Rayleigh wave train most influenced by the near surface. This was done by our choice of frequency band, by varying only the properties of the half-

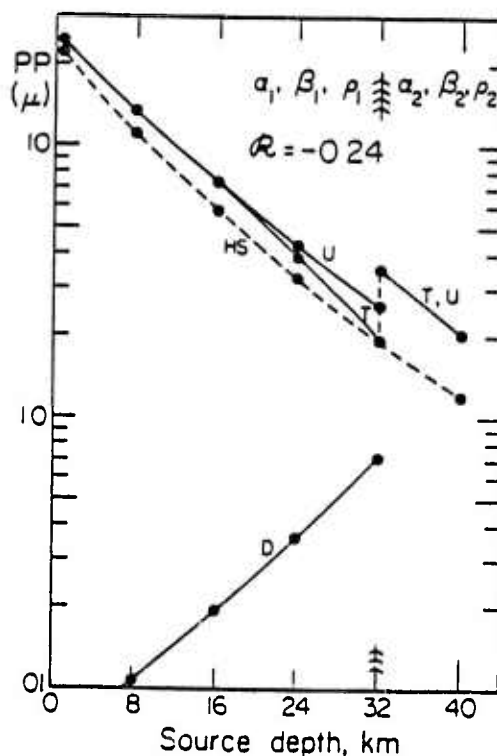


Figure 20. Maximum LRS-M-LP Rayleigh amplitudes, PP, for the one layer half-space model, $R = -0.24$, versus source depth due to the upgoing, U, downgoing, D, and total, T, source radiation.

space and by measurements at near ranges where dispersion effects had not yet dominated the wave signature. The reversal of Rayleigh wave polarity resulted from a negative reflection coefficient which was obtained with an unrealistic density depth contrast.

Reversal can also occur when the upper media are realistically varied and the Rayleigh wave is primarily controlled by the lower media. In fact, a reversal can occur in source structure models as reasonable as that of the NTS tuff region.

Earlier we mentioned several reasons to suspect that the upgoing waves from explosion sources might make a substantially smaller contribution to the far-field Rayleigh waves than is predicted by elastic wave theory. If so, what are the effects on the dependence of surface wave amplitude on source material and

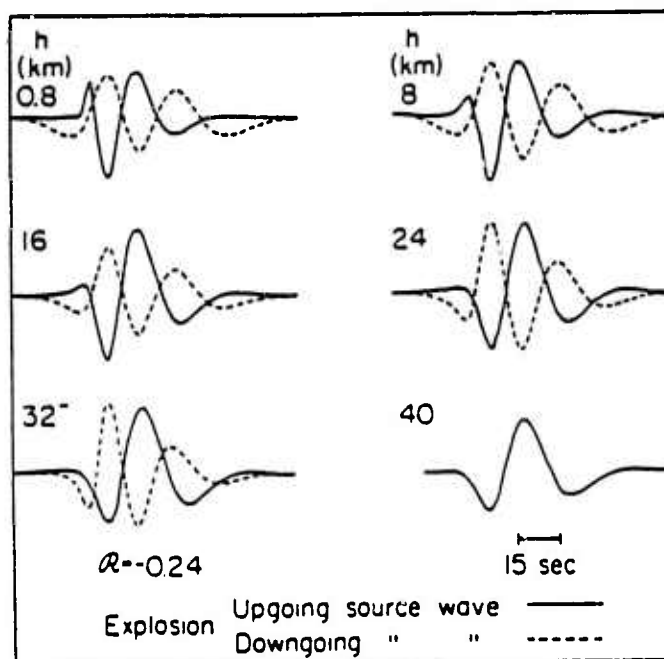


Figure 21. Rayleigh wave LRS-M-LP seismograms due to the upgoing source wave superimposed on those due to the downgoing source wave at a range of 160 km and selected source depths.

explosion yield? To address these questions we use an approximate Rayleigh wave synthesis technique described by Bache *et al.*, (1978a), that separates the contribution of the source region geology from that of the average propagation path. Separate crustal models are used for the source region and average path, with the vertical displacements computed from

$$\left\{ \bar{w}_0 \right\}_R = -4\pi i \mu k_1 \bar{\Psi}(\omega) K_{R1} A_{R1} T(\omega) H_0^{(2)}(k_2 r) \quad (2.33)$$

where quantities computed for the source region and average path are denoted by subscripts 1 and 2, respectively. The transition between the two structures is accounted for by the approximate transmission coefficient

$$T(\omega) = (c_2 A_{R2})^{1/2} / (c_1 A_{R1})^{1/2} \quad (2.34)$$

The derivation of this equalized horizontal energy flux expression is given in a later section.

Bache *et al.* (1978a) computed synthetic seismograms for explosions in three distinct source regions encountered at NTS, Climax Stock, where the source is in granite, Yucca Flat and Pahute Mesa, where the source materials are tuff or rhyolite. The Climax Stock and Yucca Flat source regions represent the extremes in material properties, and so will be used for our examples. The average path model is the western United States model 35-CM2 of Alexander (1963). The GRTM attenuation model and LRSM-LP instrument response were used in all the computations to be presented.

The remainder of this section is essentially a reevaluation by T.C. Bache of Bache *et al.* (1977 and 1978b). First, we construct theoretical amplitude-yield curves for explosions in the two different source regions. Synthetic seismograms were computed at a range of 3000 km for several yields (W) with the source depth (in meters) varying as $107 W^{1/3}$. All seismograms were computed with the same reduced displacement potential which was cube-root scaled to the yield. At the dominant periods, this source function is nearly constant and so can be characterized by its static level, $\Psi(\infty)$, at a particular yield. For these examples $\Psi(\infty) = 91 \text{ m}^3$ at 20 KT.

Selected synthetic seismograms are shown in Figure 22. We see that the waveform is essentially independent of yield and source depth and the source structure has only a minor effect. Consistent amplitude measurements are easily made and the M_s is computed from

$$M_s = \log A + 1.656 \log \Delta - 0.18 \quad (2.35)$$

where A is peak-to-peak amplitude in nanometers and Δ is the range in degrees.

Now let us look at the effect of suppressing all or part of the upgoing waves from the explosion source, which is illustrated by the examples in Figures 23 and 24. When the source is in granite the Rayleigh waves from the up and downgoing waves are almost exactly in phase and simply sum to make up the total seismogram. For the source in tuff (Figure 24) the situation is entirely different. The contributions from the up and downgoing waves are nearly opposite polarity (the phase difference is about 225 degrees at long period). Therefore, suppression of the upgoing waves actually increases M_s .

Synthetic seismograms analogous to those in Figure 22 were computed with the upgoing waves totally suppressed. As with the total source seismograms, the waveform is essentially independent of explosion yield and depth and looks like the "downgoing" examples in Figures 23 and 24. For each of the four cases we plot

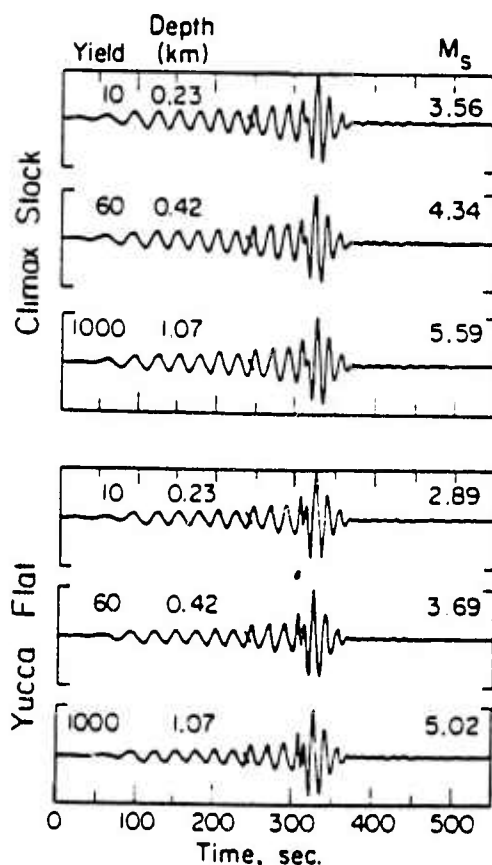


Figure 22. Typical vertical component synthetic seismograms at 3000 km for a range of yields at fixed scaled depth in structures representing NTS Climax Stock and Yucca Flat. The cycle at which amplitude was measured is indicated with a bar. The period of this cycle is 19.7 seconds for the Climax Stock seismograms and 18.8 seconds for the others.

M_s versus yield in Figure 25. The linear least squares fit to these data is shown with each set. The interesting features are the slope of the M_s on the source material and on the degree of suppression of the upgoing waves.

In all cases the slope is slightly greater than unity. This is due to two factors. First, the source level at 20 seconds is not precisely proportional to W , but increases slightly faster. If this were the only factor, the slope would be 1.02. The

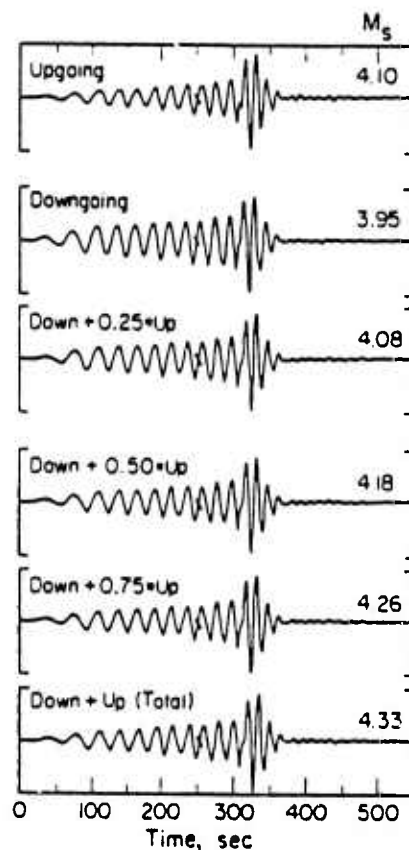


Figure 23. The relative contribution of the upgoing and downgoing source waves to one of the Climax Stock seismograms of Figure 22. Note that positive vertical is down for this set.

second factor is the depth dependence of the eigenfunctions; for example, K_{R1} is a slightly increasing function of depth for the tuff events. Of course, we have ignored any depth-dependent source coupling effects which are likely to be much more important. As an indication of these effects, we refer to Murphy (1977), who points out that the semi-empirical source model of Mueller and Murphy (1971) gives

$$\psi(\infty) \sim W^{0.76} \quad (2.36)$$

for NTS explosions at a constant scaled depth of burial. The M_s -log yield curves obtained using this model would have slopes that are 0.24 smaller than those in Figure 25.

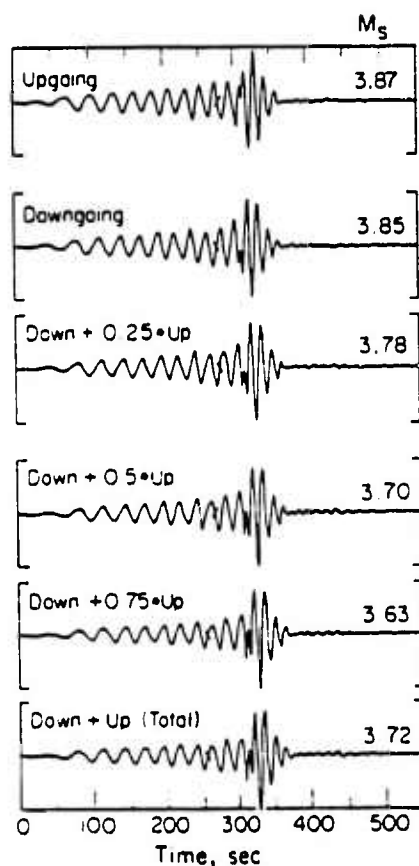


Figure 24. The relative contribution of the upgoing and downgoing source waves to one of the Yucca Flat seismograms of Figure 22. Note that positive vertical is down for this set.

The dependence of M_s on the source material is strongly influenced by the degree of suppression of upgoing waves. Bache *et al.* (1978a) discussed this dependence for a total RDP source. For a fixed path and source, the Rayleigh waves from explosions in different source materials are proportional to

$$G_s = \mu_s \frac{K_{R1} A_{R1}}{c_1} T(\omega) \quad (2.37)$$

This quantity was plotted versus period for the Climax Stock, Yucca Flat and Pahute Mesa source regions discussed in that paper. Comparing the fairly similar Yucca Flat and Pahute Mesa regions, Bache *et al.*, pointed out that $G_s \sim \mu_s$, leading to $M_s =$

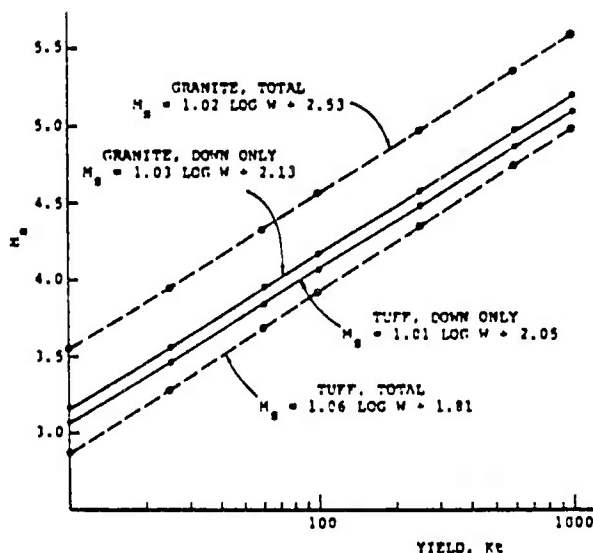


Figure 25. M_s versus yield for four data sets. The results of a linear least squares regression, assuming all error in M_s , are shown for each data set.

$\log [\mu_s \Psi(\infty)]$. However, for the dissimilar Climax Stock and Yucca Flat regions the dependence on μ_s is weaker. For the path models in that earlier paper or the 35-CM2 used here, this dependence is approximately

$$M_s \sim \log [\mu_s^{3/4} \Psi(\infty)] \quad (2.38)$$

It is interesting to compare this to the elementary scaling

$$M_s \sim \log [\mu_s^{-3/4} \Psi(\infty)] \quad (2.39)$$

derived from the half-space scaling in (2.31). This is inappropriate because it assumes that the entire path is changed with the local source medium and is valid only when the source layer path length and thickness are large enough to be effectively considered a uniform half-space. Limited source region results, e.g., an embedded transparent spherical source medium, predict an approximate dependence of $M_s \sim \log [\mu_s \Psi(\infty)]$ as above.

If the upgoing wave is suppressed, the dependence on the local shear modulus is much weaker. From the M_s values in Figures 23 and 24 we compute that total suppression of the

upgoing waves leads to the approximate scaling

$$M_s \sim \log [\mu_s^{1/8} \psi(\infty)] \quad (2.40)$$

Since the upgoing waves are almost certainly suppressed to some extent compared to the downgoing waves, the actual dependence on the local shear modulus should be somewhere between that in equations (2.39) and (2.40).

3. MIXED PATH RAYLEIGH WAVE TECHNIQUES

Rayleigh waves from events in close proximity, but in different source materials have been modeled by a variety of techniques. Before last year most of the techniques were based on approximations to matching the boundary conditions across a vertical plane separating two earth models, one for the source region and one for the remainder of the path. The formulation and suggested approximations were given by McGarr and Alsop (1967). If one assumes no mode conversion or reflection and normal incidence to the boundary, the approximation is equivalent to assuming that the total horizontal energy flux remains constant during the transmission of each mode across the boundary.

Recently a major effort has been placed on using the Knopoff-de Hoop elastodynamic integral representation, given by Burridge and Knopoff (1964), to couple nonlinear and laterally inhomogeneous source regions to laterally homogeneous propagation models. In the following sections we will give a brief derivation of the first technique and finish with a representation theorem application to Rayleigh waves generated by a vertical point force in a multilayered media.

3.1 Equal horizontal energy flux method

The assumption that there is no mode conversion is probably most appropriate for the fundamental mode Rayleigh wave since it is a true surface wave and only requires a free surface for its existence. On the other hand, higher mode Rayleigh waves and all Love modes are primarily constructive interfering multiply reflected and refracted body waves trapped in the crust upper mantle wave guide. In order for there to be minimal mode conversion for these waves, constructive interference angles should be nearly the same on both sides of the transition zone. Of course, in the presence of low velocity layers at periods for which the phase velocity of the fundamental Rayleigh wave is large enough to represent body waves in these layers, there may be considerable mode conversion even for the fundamental mode. Finite element results indicate that for an oceanic-continental transition zone this approximation is reasonable for the

fundamental Rayleigh wave and that there is considerable mode conversion for all Love wave modes.

The energy flux or transport is given by Biot (1956) as

$$W = 2UE\bar{w}_0^2 \quad (3.1)$$

where U is the group velocity and E is total potential or kinetic mode energy at a particular frequency normalized to the square of the surface vertical spectral displacement or in the case of Love waves surface horizontal displacement. Equating the energy flux across the boundary we have

$$\bar{w}_{02} = (U_1 E_1)^{1/2} / (U_2 E_2)^{1/2} \bar{w}_{01} \quad (3.2)$$

The expression can be further reduced to

$$\bar{w}_{02} = (c_2 \underline{A}_{R2})^{1/2} / (c_1 \underline{A}_{R1})^{1/2} \bar{w}_{01} \quad (3.3)$$

using the relation in Harkrider and Anderson (1966)

$$EU = \omega^2 (2c \underline{A}_R)^{-1} \quad (3.4)$$

where ω is the angular frequency, c is the phase velocity and $\underline{A}_R$ is the media spectral amplitude response. Specializing to the case of Rayleigh waves generated by an explosion, i.e., equation (2.24), $\Gamma = 0$ we have on the receiver side of the boundary at r_1

$$\left\{ \bar{w}_0 \right\}_{R2} = -4\pi\mu k_1 i\bar{\psi}(\omega) K_{R1-R1} T(\omega) H_0^{(2)}(k_1 r_1) \quad (3.5)$$

and as before

$$T(\omega) = (c_2 \underline{A}_{R2})^{1/2} / (c_1 \underline{A}_{R1})^{1/2} \quad (3.6)$$

In order to approximate the correct wavelength spreading in phase and amplitude in the receiver medium, we modify the distance factor to

$$H_0^{(2)}(k_1 r_1 + k_2 r_2) \quad (3.7)$$

where

$$k = \frac{\omega}{c} \quad (3.8)$$

In modeling events at NTS where source region structural differences are spatially limited compared to the propagation path length r_1 is usually set to zero and we have the relation used in (2.2).

Other mixed path approximations, which have been used, are to take the expression for an explosion in a uniform receiver type medium and replace $\mu \Psi(\omega) K_R(h)$ with values appropriate for the source medium. Even the best of these approximations is based upon assumptions which are questionable for limited source regions, e.g., that the source region wave front and the boundary are parallel.

3.2 Elastodynamic integral representation method

The Knopoff-de Hoop integral representation theorem can be used to extend near source calculations in laterally and vertically inhomogeneous regions to regional and teleseismic ranges using standard laterally homogeneous half-space codes (Harkrider *et al.*, 1981). We define a surface which encloses the heterogeneity of the source region and exterior to which the medium can be represented by a plane stratified medium. As long as the possible interaction of back reflected exterior radiation with the source region including the source itself can be neglected, i.e., a transparent source region, the Green's functions are those appropriate for the exterior region with no heterogeneities present. For a plane stratified half-space terminated at depth by a homogeneous half-space, the source containing surface can be a cylinder, i.e., canister, extending from the free-surface down into the terminating half-space. Various numerical techniques can then be used to determine the stress and displacement at a finite number of nodes on the surface. By spatially integrating the convolution of these nodal quantities with the appropriate surface wave Green's functions for ring sources in the laterally homogeneous region, we can obtain synthetic Rayleigh and Love waves for arbitrarily complex sources.

The vertical spectral displacement at the free surface of a multilayered elastic half-space can be expressed in terms of a surface integral of observed displacements and stresses as

$$\bar{w}_0 = \int_{\Sigma_{SES}} \{ \bar{G}_{3j} \bar{\sigma}_{jk}(\xi) - \bar{T}_{3kj} \bar{u}_j(\xi) \} n_k dS(\xi) \quad (3.9)$$

where

$$\bar{T}_{ikj} = \lambda \delta_{kj} \bar{G}_{ip,p} + \mu (\bar{G}_{ik,j} + \bar{G}_{ij,k}) \quad (3.10)$$

$\bar{G}_{3j}$: Spectral displacement in the vertical, x_3 , direction on the free surface due to a point force in the j direction at ξ .

$\bar{u}_j$: Components of the spectral displacements observed at ξ .

$\bar{\sigma}_{ij}$: Components of the stress tensor at ξ .

n_k : Unit vector normal to the surface Σ_{SES} , positive into the nonlinear source region.

Σ_{SES} : Surface excluding the nonlinear source region from the linear propagation media

and

$$\bar{G}_{ij,k} \equiv \frac{\partial}{\partial \xi_k} \bar{G}_{ij} = - \frac{\partial}{\partial x_k} \bar{G}_{ij} \quad (3.11)$$

Assuming an azimuthally symmetric distribution of displacements and stresses on Σ_{SES} , we integrate the integrand of equation (3.9) over a ring of radius r_o and at a depth z_o . The G_{ij} 's are the wave number integral solutions found in Harkrider (1964). Taking the residue contribution to obtain the Rayleigh waves, we obtain the following ring integration contributions to $\{\bar{w}_0\}_R$:

$$\begin{aligned} \left\{ \bar{w}_0 \right\}_R &= \int_0^h (\bar{u}_z \{\bar{T}_{zr}^o\} + \bar{u}_r \{\bar{T}_{rr}^o\} - \bar{\sigma}_{rz} \{\bar{G}_z^o\} - \bar{\sigma}_{rr} \{\bar{G}_r^o\}) \, adz_o \\ &+ \int_0^a (\bar{u}_z \{\bar{T}_{zz}^o\} + \bar{u}_r \{\bar{T}_{zr}^o\} - \bar{\sigma}_{zz} \{\bar{G}_z^o\} - \bar{\sigma}_{rz} \{\bar{G}_r^o\}) \, r_o dr_o \end{aligned} \quad (3.12)$$

with the ring Green's functions given as

$$\{\bar{T}_{zz}^o\} = -2\pi \left(\frac{i}{2}\right) \bar{y}_2^R(z_o) J_0(kr_o) \underline{A}_R H_0^{(2)}(kr) \quad (3.13)$$

$$\{\bar{T}_{zr}^o\} = -2\pi \left(\frac{i}{2}\right) \bar{y}_4^R(z_o) J_1(kr_o) \underline{A}_R H_0^{(2)}(kr) \quad (3.14)$$

$$\begin{aligned} \{\bar{T}_{rr}^o\} &= -2\pi \left(\frac{i}{2}\right) \{\mu \bar{y}_3^R(z_o) [J_0(kr_o) - J_2(kr_o)] \\ &+ 2 \frac{\lambda\mu}{(\lambda + 2\mu)} K_R(z_o) J_0(kr_o)\} k \underline{A}_R H_0^{(2)}(kr) \end{aligned} \quad (3.15)$$

$$\{\bar{G}_z^o\} = -2\pi \left(\frac{i}{2}\right) \bar{y}_1^R(z_o) J_0(kr_o) \underline{A}_R H_0^{(2)}(kr) \quad (3.16)$$

and

$$\{\bar{G}_r^o\} = -2\pi \left(\frac{i}{2}\right) \bar{y}_3^R(z_o) J_1(kr_o) \underline{A}_R H_0^{(2)}(kr) \quad (3.17)$$

with $r_o = a$ in the first integral which is over the canister's side and $z_o = h$ in the second integral which is over the canister's bottom.

An example of the technique is shown in Figure 27. The nodal displacement-stress values were calculated at Del Mar Technical Associates and the results were for a joint AFOSR project. The synthetics shown in the figure are the vertical

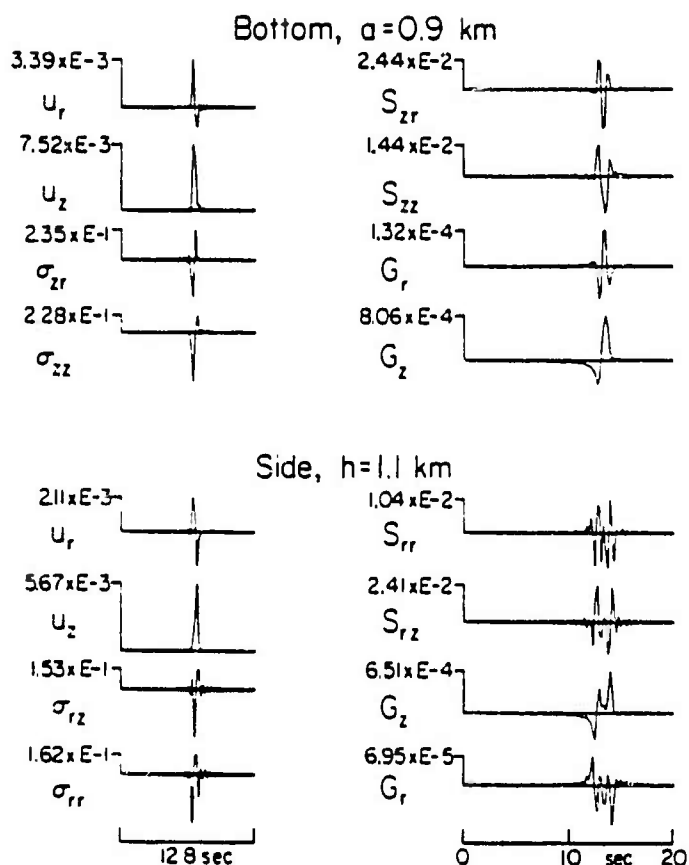


Figure 26. Nodal forcing functions and corresponding Green's functions for a ring on the bottom and a ring on the canister bottom for the HSO model.

Rayleigh wave displacements due to a vertical point source at a depth of 0.4 km in a homogeneous half-space at a range of 300.1 km. The nodal displacements and stresses are calculated for the bottom and sides of a cylindrical canister of radius 2.1 and vertical extent of 2.1 km. The vertical force is located on the axis with a band limited delta function history. The frequency range is 0.01 - 2.0 hertz.

The nodal values were calculated by two very different techniques. The SWIS code is a finite element code capable of calculating displacements and stresses in a vertically and laterally inhomogeneous medium. For practical purposes the lateral range is restricted to no more than five (5) source depths in this problem. The PROSE code is a full $(\omega-k)$ numerical integration solution to a source and receiver in a plane layered viscoelastic half-space. The two nodal value results are combined

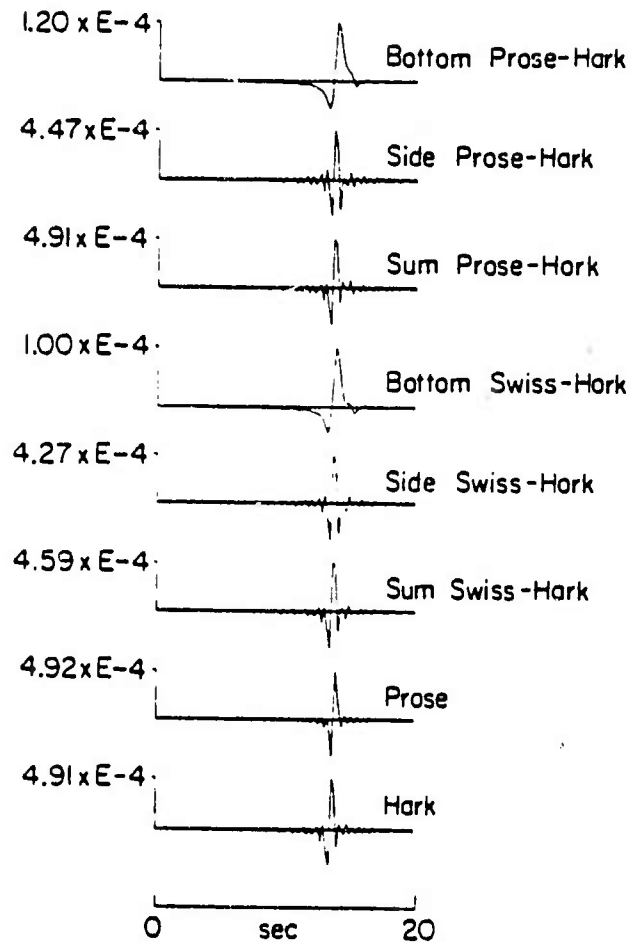


Figure 27. Contributions to the final solution of the summed convolutions from the bottom and side of the canister for the HSO model.

with ring source surface wave Green's functions, which propagate the values from the canister to receiver and are calculated by the plane layered surface wave code HARK.

Nodal displacement and stress forcing histories calculated by SWIS and their associated surface wave Green's functions from HARK are shown in Figure 26. The upper traces are for a ring of width 0.2 km and radius 0.9 km located on the bottom of the canister. The lower traces are for a ring of radius 2.1 km, depth 1.1 km, and width 0.2 km located on the side. These results are for the HSO homogeneous half-space discussed in section 2.2. The S's in the figure are the time domain equivalents of the spectral T's in the previous equations. The units are such that

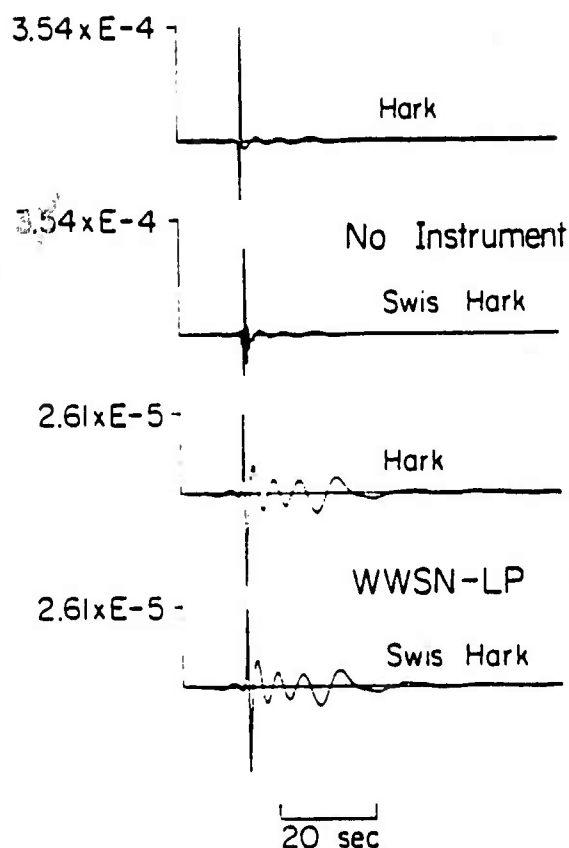


Figure 28. Comparison of the representation theorem results (SWIS-HARK) with the direct solution (HARK) for model CIT109 with and without WWSN-LP instrument.

the nodal forcing - Green's function convolutions when scaled by the ring areas, represent ground displacement at the receiver in microns due to a vertical point force of one dyne located at the source.

The resultant Rayleigh waves due to the bottom, side and total canister nodes are shown in Figure 27. The top three traces are formed using forcing functions generated by PROSE. The next three by the finite-element code SWIS. The bottom two traces are PROSE and HARK vertical displacement calculations at a range of 300.1 km from the vertical point force. HARK generates only the Rayleigh wave while PROSE includes the direct body waves which at this range are negligible compared to the Rayleigh wave. The comparison between signatures of PROSE-HARK, SWIS-HARK, and HARK is very good. This is due to the fact that propagation over the major part of the path is given by HARK calculations. The peak

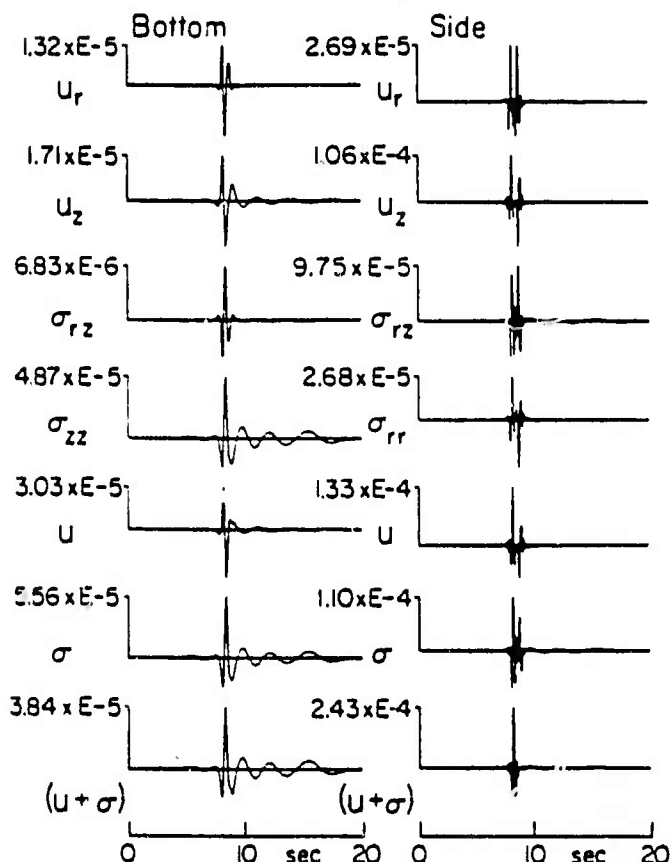


Figure 29. Contributions to the final solution for each forcing function on the side and bottom of the canister due to a vertical point force in model CIT109. Traces represent ground motion in microns.

amplitude values of PROSE-HARK, PROSE and HARK are in excellent agreement. The choice of a band limited delta function source history is a very severe test of the technique in that a slight shift of phase can give differences on the order of 10% in time domain side lobe amplitudes, i.e., the time spacing misses the narrower side lobe peak values. The differences between SWIS and PROSE as seen in SWIS-HARK and PROSE-HARK amplitude comparisons are not unexpected considering the difference in technique.

Earlier we stated that as long as the exterior reflected energy interaction with the source region is negligible, then the exterior Green's functions should be those appropriate for the propagation medium. As a test of this, we coupled the above canister driving functions, i.e., nodal displacements and stresses

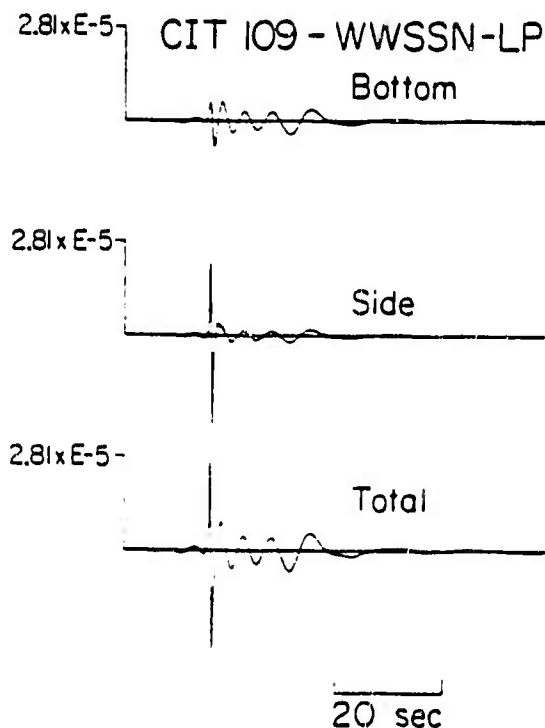


Figure 30. Results for CIT109 showing the contributions of the bottom and side of the canister as seen through a WWSSN-LP instrument.

for a vertical point force in a simple half-space, to the Rayleigh Green's functions for a multilayered crust-mantle model of the western U.S., CIT109. The negligible reflection criteria to be satisfied in this case since the surface layer of CIT109 was the same elastic medium as the interior source region.

The results are shown in Figure 29 at a range of 1500 km with a WWSSN-LP instrument and without. The comparison of the direct calculation HARK with the SWIS forcing or driving function result as seen through the long period instrument is outstanding both in absolute amplitude and shape especially before and after the maximum peak to trough amplitude. The two signals are virtually indistinguishable when overlayed.

The contribution of each forcing function on the bottom and side of the canister is shown in Figure 29. The traces represent surface vertical displacement at the receiver. The high frequency and maximum amplitude contribution comes from the canister side and the long period from the bottom. One might expect this type of frequency partition because of the mean depth of the contrib-

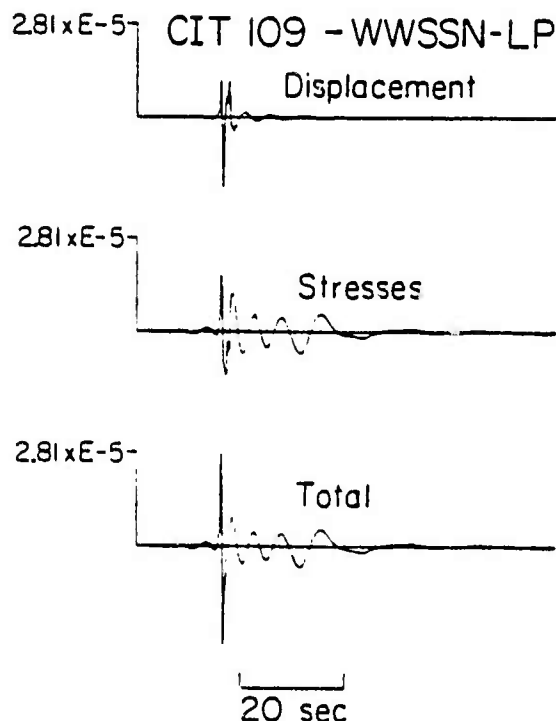


Figure 31. Contributions of the displacement and stress forcing functions to the CIT109 WWSSN-LP seismogram.

uting canister surfaces. On the bottom, the stresses make a larger contribution than the displacements and on the side the displacements are more important although neither dominate by more than a factor of two.

In Figure 30, the bottom, side, and total canister contributions to the Rayleigh wave WWSSN-LP seismograms are shown. As in the synthetic without an instrument (Figure 29), the high frequency and maximum amplitude contribution is from the side of the canister. From Figure 31, we see that the high frequency portion of the WWSSN-LP seismogram requires both displacement and stress values on the canister, whereas the long period energy comes predominately from the stresses.

4. SUMMARY

In the preceeding sections, the need for the theoretical inclusion of different source regions, various techniques for this, and some of the results from coupling near source phenomena into surface wave generation were presented. The techniques reviewed were vertical separation of source radiation, mixed path

approximations, and source region coupling through the elastodynamic representation theorem.

In closing, it is important to realize that different source regions relatively close to each other can theoretically not only exhibit significant differences in teleseismic Rayleigh wave amplitudes but also reversals in polarity. This can be caused by local source region structure as well as change in tectonic environment.

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Body wave amplitude patterns and upper mantle attenuation
variations across North America

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BODY WAVE AMPLITUDE PATTERNS AND
UPPER MANTLE ATTENUATION VARIATIONS
ACROSS NORTH AMERICA

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Summary. A broad band amplitude study of P and SH waves from deep South American events recorded at WWSSN stations in the United States provides constraints on upper mantle variations in attenuation. The events were selected to minimize contamination due to source complexity, radiation pattern, and source structure. The prevalent feature in the short period and long period bands is that the SH amplitudes show the same regional variations as the P amplitudes, with the regional variations of short periods being similar, but enhanced relative to the variations at long periods. The amplitudes do not indicate a systematic regional difference between the East Coast and Rocky Mountain provinces, but both short and long periods are enhanced at the Gulf Coast and midwestern stations, probably due to thick sedimentary bed receiver structure. Since detailed receiver structures for individual stations are poorly known, we used the working hypothesis that the regional amplitude patterns are caused by lateral variations in upper mantle attenuation. Both a constant Q and a frequency dependent attenuation operator were considered, with the relative amplitudes and waveforms in short period and long period bands used to estimate the acceptable range of Δt^* in the constant Q model, or the acceptable range of the parameter τ_m , which describes the high frequency roll-off of the absorption band in the frequency dependent model. Time domain modeling indicates that the short period P and SH amplitudes and waveforms in the 1 to 10 sec period range do not unambiguously demand frequency dependent attenuation,

allowing for the uncertainty in source frequency content and long period absorption band amplitude. The short period P amplitudes can be fit with a range in $\Delta t_{\alpha}^* = 0.5$ sec or a range in τ_m from 0.001-0.25 sec, with either single parameter variation consistent with the general short period SH amplitude behavior. A technique for determining the absolute value of t^* and possible τ_m values appropriate for the 5-20 sec band was applied to selected data. A value of $t^* = 0.8$ sec and small τ_m were found for travel paths from the deep events to North America, and lower t^* or large τ_m values were found for paths to southern African stations. Long period amplitude variations cannot be explained by frequency dependent models with only a single high frequency roll-off of the absorption band, but values of Δt_{β}^* determined from the long period SH amplitudes do exceedingly well in predicting the long period P amplitudes under the assumptions of constant Q and $t_{\beta}^*/t_{\alpha}^* = 4.0$. The total range in long period Δt_{β}^* is about 3 to 4 sec. Though receiver effects may cause some of this variation between stations, the data make it clear that such effects must produce amplitude behavior similar to that caused by regional variations in attenuation. It is shown that the Δt_{β}^* values determined from the long periods do not predict the short period amplitudes in detail, but the predictions are not prohibitively outside the range of the short period amplitude variations. Spectral analysis of the long period SH waves was also performed, yielding a slightly greater range of Δt_{β}^* than the time domain results, but the resulting values

of Δt_p^* do not predict the amplitude variations well, indicating the large effects of long period receiver functions on both the amplitude and spectral behavior.

1 Introduction

This study examines the regional amplitude behavior of body waves over a range of periods from one to 20 seconds. The short period and long period WWSSN seismographs in the contiguous United States serve as a broad array for this purpose. Short period and long period P and SH wave amplitudes from deep South American events in Peru and Argentina provide a large data set, which has been carefully screened to minimize factors such as source complexity, trends due to radiation pattern, and source structure. Present knowledge of the receiver structure under each WWSSN station is inadequate for removing receiver effects from the amplitude data, which precludes direct interpretation of the amplitude behavior in terms of upper mantle scattering or attenuation. The general impact of receiver effects is to increase the total range in relative amplitude behavior, particularly between amplitudes recorded at receivers on thick sedimentary layers and amplitudes recorded at stations in tectonically active or shield regions. Bearing this in mind, we sought to determine the bounds on the possible range of lateral attenuation under the United States by finding attenuation models which reproduce the

waveforms and amplitudes of the data. The range in period of the data, as well as the simultaneous use of P and SH phases, provides constraints on the attenuation models. Both constant Q and frequency dependent attenuation models have been examined, and time domain and frequency domain methods applied.

Numerous studies have indicated systematic regional differences in body wave amplitudes and frequency content between the eastern and western United States. The amplitude studies include measurement of short period P wave amplitudes (Cleary 1967; Evernden & Clark 1970; Butler & Ruff 1980); short period and long period P wave m_b (Booth, Marshall & Young 1974); m_b versus M_s (Ward & Toksöz 1971; Evernden & Filson 1971; Marshall, Springer & Rodean 1979); and short period P and S amplitudes (Der, Masse, & Gurski 1975). The majority of amplitude studies have utilized LRSM stations, and these yield a general tendency for body wave magnitudes to be .26 magnitude units larger in the East relative to the West. Spectral analysis of short period P waves (e.g. Der & McElfresh 1977) and of short period S waves (Der, Smart & Chaplin 1980) have produced evidence for lateral variations in upper mantle attenuation with average $\Delta t_{\alpha}^* = .25$ sec and $\Delta t_{\beta}^* = .67$ sec between the East and West Coasts. Such lateral variations of attenuation, combined with receiver effects, may account for much of the m_b variations at LRSM stations (Der, McElfresh & Mrazek 1979).

The amplitude studies utilizing WSSN data of Butler & Ruff (1980) and Burdick (1978) failed to detect significant east to west

differences in amplitudes, though individual stations showed distinctive amplitude patterns and some regional trends. These studies were more selective in choosing simple, coherent nuclear explosion and earthquake data than most previous studies. This selectivity is necessary for analysis of amplitude trends in the time domain, as it minimizes the effects of source complexity and radiation pattern, which can otherwise only be limited by utilizing very large data sets. Analysis of long period WWSSN body wave spectra indicate regional variations in attenuation of $\Delta t_p^* = 7.0$ sec (Solomon & Toksöz 1970; Mikumo & Kurita 1968; Teng 1968). The resulting Δt^* values differ from, and exceed, those determined at short periods. This last point has been asserted as evidence for frequency dependence of attenuation, since the long period values of t^* would prevent observation of 2-4 Hz energy in short period P and S waves.

The amplitude data gathered in this study does not clearly indicate a simple east to west regional amplitude variation in the short period band. The major amplitude anomaly is in the Great Plains and Gulf Coast regions, for both short period and long period waves, consistent with the results of Butler & Ruff (1980) and Der et al. (1979). As shown in this paper, the amplitude variations are grossly consistent with regional variations in upper mantle attenuation, and this consistency argues against dismissing the amplitude behavior as being purely the result of varying receiver structure until further constraints are placed on the

regional variations. A companion study will present data from other azimuths and a more detailed assessment of the individual receiver structures.

The evidence for lateral variations in attenuation from spectral studies have principally utilized data in the .5-10 Hz band (Der et al., 1980). The relatively high energy levels of frequencies greater than 1 Hz are inconsistent with frequency independent t^* values of around 1 sec that are often adopted in the long period body wave band (10-30 sec). This indicates that either t^* is frequency dependent within the body wave band, or the long period body wave t^* values are inappropriate. There are large uncertainties in the absolute t^* levels throughout the body wave band, though the t^* values for periods shorter than one second must be low enough to allow the high frequency energy to be observed. Because of this uncertainty it is not obvious at what periods frequency dependent effects may be apparent. In this paper we analyze the amplitude and waveform behavior of the WWSSN body waves in terms of their implications for either frequency independent or frequency dependent attenuation in the period range 1-20 sec. We first consider frequency independent, but laterally varying, t^* models to check the consistency of such models with the data in this period range. In this analysis, due to the tradeoffs between source time function and attenuation, the absolute levels of t^* chosen for the time domain models are unimportant. The results indicate that the WWSSN body waves are fairly consistent with

frequency independent attenuation assuming $t_p^* = 4t_a^*$, as long as significant lateral variations in t^* are permitted. This result follows from the pervasive tendency for the amplitude variations of P and SH to track one another in the 1-20 sec band.

In order to accommodate the possibility that t^* does vary with frequency in this period range, we also adopt a frequency dependent Q model after Minster (1978a, 1978b). This model introduces a roll-off in the absorption band so that short period signals are less attenuated than long periods compared with the constant Q model. We find that such models can be consistent with the WWSSN data, but the data do not clearly demand frequency dependence. This procedure is somewhat sensitive to the choice of t^* , but both the waveform modeling and absolute t^* determinations that are performed indicate that we have chosen appropriate long period t^* values. The presence of the absorption band roll off ensures that these models are not inconsistent with the observations of significant energy at frequencies greater than 1 Hz. Since we lack detailed receiver corrections we cannot determine reliable station dependent attenuation parameters, rather the principle gain from this modeling is in demonstrating the consistency of the WWSSN amplitudes and waveforms with laterally varying attenuation of either frequency dependent or frequency independent models.

2 Data

2.1 DATA SELECTION

Twenty five WSSN stations in the United States have been operating over at least several years since the establishment of the network in 1962-1963. After 1965, all stations have been operating with uniform instrumentation, with short periods peaked at one second and long periods peaked at 15 seconds, except for the station at Golden, GOL, which has continued to operate its long period instrument with a response peaked at 30 seconds. The gain settings vary widely for both short period and long period instruments, ranging from $\times 6250$ to $\times 400000$ and from $\times 750$ to $\times 6000$ respectively. Thus, there is considerable variation in the amplitude of recorded body wave phases and in the reliability with which they may be measured and digitized. The data selection process has been such as to only accept data with good signal to noise ratio, resulting in fairly uniform data quality.

Since this study is aimed at upper mantle and receiver variations, the earthquake data have been carefully screened to minimize source complications. Because of presumed large upper mantle attenuation, only deep earthquakes have been analyzed, so as to clearly observe short period SH phases. We selected deep South American events because the resulting range in azimuth to North American stations is relatively small, spanning 32° from Argentina and 50° from Peru, ensuring that the deep mantle paths were similar for each event and between stations. The two clusters of deep

earthquake activity in Peru and Argentina range in distance across the United States from $42-71^\circ$ and $60-90^\circ$ respectively. Typically, the data collected for a particular event span smaller ranges in azimuth and delta since the event may not be recorded at all stations, and due to omitting any SH data contaminated by SKS or ScS, which interfere beyond 80° .

In selecting the data, several criteria were imposed. Most importantly, we considered only simple, impulsive events with clear body wave arrivals. We also required adequate coverage of a particular event to measure amplitudes at eastern and western stations. Another basic constraint was that first motions had to be consistent across the "array", with demonstrable phase coherence. These criteria were imposed with the intention of maximizing the reliability of the amplitude behavior for this relatively small data set by limiting source effects. Several events with focal mechanisms determined by other researchers were included in the data, and the first motions at all WWSSN stations were checked for comparison with events for which published mechanisms were not available. It was found that the source orientations are naturally divided between Peru and Argentina, with deep events in Peru favoring stable P wave radiation across North America, and events in Argentina favoring stable SH wave radiation. All deep events recorded since 1964 with $m_b > 4.9$ were examined, utilizing the combined WWSSN film libraries of Caltech, Menlo Park, and UCLA. Table 1 provides a listing of NOAA catalog hypocenters

of acceptable events and the phases selected for examination. Figure 1 shows a gnomonic projection of the WWSSN stations used and the earthquake epicenters. All of the events for which long period SH waves were obtained ($m_b = 5.5 - 5.8$) were used in the short period SH analysis. The events for which long period P waves were large enough to measure reliably ($m_b = 5.5 - 6.2$) generally had few measurable short period amplitudes. The short period P waves were collected from events with $m_b = 5.0 - 5.3$, and were all confined to the Peruvian zone.

Focal mechanisms of eight of the 19 events used in this study were available. The P and SH mechanisms for six of the events, as determined by Stauder (1973) using first motion and S wave polarization, are shown in Figure 2. In each lower hemisphere nodal projection, the take-off directions of stations at the extremes in azimuth and distance used for that particular event are shown. With the exception of the 1968 event, the mechanisms are all for Argentinan events used in the SH analysis. It is clear that for these normal-faulting events, the conditions producing favorable SH radiation are the proximity of one of the P nodal planes to the projection of stations on the western continental margin. Only one event, that of July 25, 1969, of the eight events used in the SH analysis has a P nodal plane that actually intersects the array. Thus, any high frequency directivity effects caused by rupture along the nodal plane should be confined to the westernmost stations. The 1968 event occurred in northern

Argentina and has a focal mechanism similar to the Peruvian events used in the P wave analysis, (Stauder 1975). For instance, the focal mechanism of the November 3, 1965 event was given by Khattri (1969) as $\phi = -9^\circ$, $\delta = 52^\circ$, $\lambda = -88^\circ$. Focal mechanisms for the same events shown in Figure 2 were given by Isacks & Molnar (1971), with most solutions being very similar and the least constraint being on the northeast trending P wave nodal plane, which does not affect the location of the SH radiation nodes significantly.

Of all the data collected, the short period P waves were from the events for which the focal mechanisms were least constrained. To circumvent this, only events in Peru, for which the larger events are known to have favorable P radiation patterns, were used. The short period P amplitudes from four Argentinan events, including the events of January 17, 1967 and December 20, 1966, were measured and compared with the pattern from the Peruvian events. These amplitudes all showed a stable east to west decreasing trend relative to the Peruvian amplitudes, consistent with the known nodal proximity of the Argentinan P waves. This indicates that the Peruvian amplitudes are indeed free of such bias. This comparison was also done for the long period P data with similar results. Synthetics computed with the focal mechanisms shown in Figure 2 indicate that the range in amplitude variation caused by radiation pattern should be less than 25% for any given event, for all phases used in this study. Use of several events probably reduces this source of error considerably.

In the course of measuring amplitudes, the calibration pulse and calibration current recorded on the seismograms were checked to ensure correct gain setting. Though the only viable test of this was by amplitude measurement of the calibration pulse, the results give some confidence in the stability of the WSSN instruments over the sixteen years spanned by the data set. Generally, short period calibration pulse amplitudes fluctuated by about 5% with rare fluctuations as large as 30%. The long period fluctuations were generally on the order of 10% with occasional deviations of up to 30%. Data with exceptionally bad calibration amplitudes were excluded.

2.2 SH DATA

The only detailed study of short period SH phases across North America has been performed by Der et al. (1980), who rotated LRSM recordings into SV and SH. They then fit lines to the log amplitude spectra to infer Δt_p^* variations. The LRSM instruments are responsive to slightly higher frequencies than the WSSN short period instruments, but the major difference is in the faster recording speed of the LRSM instrument, which allows for more accurate digitization and rotation of the SH component. For WSSN data, this stage of processing is undoubtedly the source of some amplitude and waveform error, though the relatively long period

nature of the SH observations offsets this somewhat. We selected events with adequate signal to noise ratio to enable accurate digitization. The East Coast stations recorded SH predominantly on the EW component for the events used, and required little rotation, but western stations required greater rotation. Fortunately, the very low background noise level of western stations and the high gains that they operate at, permit good confidence in the resulting rotated phases. The SH data collected for this study constitute an important new data set of high quality and are discussed below.

Figures 3 through 10 present plots of the rotated short period and long period SH phases. Station data are not presented when film records could not be located; the event was not recorded; the trace was too faint to reliably digitize; or long period transients obscure the arrival. For the 1964 and 1965 events, most WWSSN stations were operating with long period instruments peaked at 30 seconds, and few arrivals were recoverable. The amplitudes in all of these figures have been normalized to unit peak amplitude in order to better portray differences in complexity and frequency content.

In Figure 3 only short period waveforms are shown. A five second leader ahead of the SH arrival indicates the signal to noise ratio. The length of digitized signal may vary due to location on the seismogram and omission of later ScS arrivals. This event has the largest body wave magnitude of those used for the SH analysis, yet has simple, high frequency pulses which are rather coherent

from station to station. Note the strong second arrival at BEC, a feature that occurs frequently in this data set and represents a receiver effect. Note also that ATL, LUB, OXF and RCD have the highest frequency SH waves. GOL, ALQ, TUC and GSC appear somewhat lower in frequency content, consistent with the observations of Der et al. (1980). ScS phases are present in the coda of some western stations, but are always at least several seconds after the SH phase. The receiver structure at DUG is exceedingly complicated judging from the ringing nature of the coda, but this ringing is somewhat enhanced by the arrival of the strong second pulse which may be ScS. The amplitude ratio of SH to SV varies from about one in the West to more than two in the East for this event. The radiation patterns for SV indicate rapid amplitude variations across the array for this and the other Argentinian events, thus no attempt has been made to incorporate SV data in this study other than to note the greater complexity of the SV signals, which presumably results from receiver structure.

The data in Figure 4 includes a few long period SH signals, but only ALQ and TUC had changed the peak response to 15 seconds at this date. The similarity of signals between instruments peaked at 30 and at 15 seconds is due to the similarity in response over the 10 to 20 second band, which spans the dominant periods of these long period SH signals. The amplification difference over this range averages only 20%. The numbers shown are the gain corrected amplitude ratios of the short period to long period signals. The

amplitudes measured were the zero line to peak of the first half cycle of the long period SH, which was chosen to avoid receiver structure complexity later in the signal, and the peak to peak measurement of the first $1\frac{1}{2}$ cycles of the short period signal. This latter amplitude was measured rather than the first peak to trough amplitude because it was subject to less scatter between events. Note that the short period to long period amplitude ratio can vary by a factor of three, even between closely adjacent stations such as SCP and OGD or TUC and ALQ. The strong second arrival at OGD is not present at SCP, thus some of the short period amplitudes may be diminished by scattering at OGD. There is evidence that such is the case at TUC also, where both the long periods and short periods show strong arrivals about 10 seconds after SH. As proves true for all of the events, ATL and LUB have very high frequency signals. The strong second arrival at DUG is either a receiver phase or ScS.

More complete long period and short period SH data sets are shown in Figures 5 through 9. The data in Figures 5 and 6 indicate that these were very similar events as suggested by the focal mechanisms of Figure 2. Of particular interest are the high frequency content and large short period to long period ratios for stations OXF, FLO, JCT and LUB. Strong second arrivals are apparent at BEC, SHA and TUC. ScS arrives in the SH overshoot at BOZ, GSC, DUG, and RCD, but is sufficiently delayed to reliably measure the first half cycle of SH. There is no obvious

distinction in frequency content between the Rocky Mountain stations and East Coast stations; the only signals with consistently different frequency content being those recorded by the midwestern and Gulf Coast stations which are known to overlie sedimentary structures (Butler & Ruff 1980).

These figures show that the long period waveforms at OXF and ATL stand out as being exceptionally long period. Signals at GOL were recorded with a peak instrument response of 30 seconds, so they too are broadened. Burdick (1978) found large amplitude ratio anomalies associated with ATL and OXF, and his data included the events in Figures 6 and 7. He correctly attributed some of the OXF anomaly to a substantial drop in long period amplitude relative to adjacent stations, presumably a strong receiver effect. The high frequency arrivals at OXF are also associated with relatively high short period amplitudes. In this data ATL does not show anomalous amplitude ratios, only very broad long period waveforms.

The event of July 25, 1969, shown in Figure 8, has very complete coverage, but slightly greater SH wave complexity. This complexity is not manifested in the long period signals, but is rather clear in the short periods. There appears to be a second pulse in the short period waveforms, smaller than the first and a few seconds later. Despite this, the waveforms are still rather coherent, and demonstrate the same characteristics as the other events. For this event, the P nodal plane appears to trend north through JCT, but it is difficult to discern any obvious effects of

having the nodal plane actually segment the array.

The focal mechanism for the event in Figure 9 was not well constrained by first motions. However, a comparison with Figure 3 shows that the short period waveforms at common stations ATL, GSC, TUC and ALQ are virtually identical, and thus, the mechanism is probably much the same as for the 1964 event shown in Figure 2. The distinctive feature in Figure 9 is the very large amplitude ratio at DAL. This value is the only one determined for DAL, a station with an erratic recording history, and is thus somewhat suspect. The high frequency of these simple SH pulses, relative to the other events, does tend to enhance any visual difference between the East and West Coast stations. This may be consistent with Der et al.'s (1980) assertion that higher frequency sources seem to show more dramatic differences in East versus West Coast signals.

The last event included in the SH data set (Figure 10) is a small event that was located slightly north of the other events. The signal to noise ratio is substantially lower due to the small size of the event and no long period signals were retrievable. Though the waveforms are not quite as stable as desired, it is of interest to determine whether there is resolvable source region bias in the data, which will be discussed later. The basic constraint on the focal mechanism of this event was that the western P waves were close to being nodal, as was the case for the other events used to obtain SH phases.

2.3 P DATA

The amplitude data set for long period P consists of six events, four located in Peru, and two located in northern Argentina, as shown in Figure 1 and Table 1. All events were checked for consistency with favorable P-wave radiation patterns, which was facilitated by three published focal mechanisms of events used. The gain setting of WWSSN instruments required examination of slightly larger magnitude events than for the other phases, but all long period waveforms were selected to be as simple and as impulsive as possible. An example of the P wave data is shown in Figure 2 of Mikumo & Kurita (1968). Stations at all azimuths from 30° to 90° were checked to ensure similarity in P and S waveforms in order to avoid excessive source complexity. Several of the events had relatively strong second arrivals about ten seconds after P, which may indicate some source complexity, but this did not contaminate the amplitude measurements, which were of the peak of the first half cycle of the P arrival. Measurement of this peak minimized crustal and multiple source effects. The events had very good signal to noise ratio and thus amplitudes could be accurately measured with an error of approximately 10%. The short period amplitudes of these events were generally too large to measure, however, they were systematically checked for evidence of excessive source complexity. As would be expected based on the relative

magnitudes, the short period signals of the events used for long period P wave amplitudes were more complicated than those of the data used for short period P wave amplitudes. However, they were far less complicated than for larger deep events which were excluded from any measurement.

The short period P amplitudes from deep events constitute a somewhat sparse data set. The focal mechanism and source complexity constraints limited the number of acceptable events to only five. However, the much larger data set of carefully selected P waves from shallow events in South America presented by Butler & Ruff (1980) and Butler et al. (1979) provides a useful comparison and supplement. In the analysis described below, the deep and shallow data sets were treated separately and the results compared to verify that the small data set from deep focus events does not introduce systematic errors. Butler & Hart (1979) show many of the shallow P wave waveforms employed in the shallow source analysis. Our P waves from deep sources were generally of significantly higher frequency content and few could be reliably digitized, though amplitudes could be measured with an accuracy of 10-20%. The amplitude measured was the peak of the first half cycle to first trough measurement which was found to be most stable and free of receiver effects by Butler & Ruff (1980). All of the new short period P data were from closely clustered hypocenters in Peru. The high frequency nature of the data precludes any general observation of frequency differences between the eastern and western stations,

though, again, the midwestern and Gulf Coast stations tend to be higher frequency.

3 Relative Amplitudes

The amplitude data were normalized to a common gain and then amplitudes from Argentina events were corrected for geometric spreading to a distance of 70° , while those from Peru were corrected to 60° . This amounted to no more than a 15% correction for most stations, using the expressions of Kanamori & Stewart (1976). We then employed the technique described by Butler & Ruff (1980) to normalize each data set. In this procedure we took the set of observations of a particular phase, O_{ij} , for I events and J stations, and from the I events chose a reference event k . Scale factors a_i were found to minimize the least squares error for each event $i \neq k$:

$$\min \sum |a_i O_{ij} - O_{kj}|^2.$$

Let B_k be the average amplitude of the master event k :

$$B_k = J^{-1} \sum_j O_{kj}.$$

The total error for all events I in the source region is then:

$$\beta_k^{-1} \sum_i \sum_j |\alpha_i o_{ij} - o_{kj}|^2. \quad (1)$$

By iteratively allowing each event i to be the master event k , factors α_i were found which best minimized the total error (1).

The short period P amplitudes are shown in Figure 11. For this and all similar figures, the ordinate is a log scale indicating relative amplitude, and stations are ordered in a west to east arrangement as viewed from the South American azimuth. The scatter in amplitudes at each station is relatively small, less than a factor of 1.5 around the mean, and individual events do not appear to have significant trends across the array relative to the station means. The lowest signal amplitudes are observed at TUC, ALQ and OGD, with GOL and SCP being relatively low. High amplitudes are concentrated in the Texas stations LUB and DAL as well as at FLO, RCD and COR. The midwestern and Gulf Coast stations tend to record the highest amplitudes, with the East Coast and West Coast stations scattering around the same level. This general pattern is consistent with the results presented by Butler & Ruff (1980). The total range in mean amplitudes between stations is a factor of six.

A similar basic pattern, but one containing a greater total range in variation of a factor of 10, is shown by the short period SH amplitudes in Figure 12. With the exception of ALQ, the Rocky Mountain stations record relatively low amplitudes, though the eastern stations ATL, OGD and WES are similarly low. Again,

midwestern and Gulf Coast stations are generally enhanced. The contact between stations recording high amplitudes and stations recording lower amplitudes is more abrupt along the Rocky Mountain front than in the East, as is the case for the P waves. The 1975 event in northern Argentina which was located north of the other epicenters may have a slight east to west amplitude trend relative to the other events, but this is not sufficiently outside the scatter of the data, nor is the focal mechanism well enough constrained, to detect significant source region bias in the amplitudes. The scatter at stations OXF, GEO and SHA tends to be greater than for most stations, and there is a single, very high amplitude observation at DAL. This scatter and anomalously high amplitude may indicate strong receiver effects that are acutely sensitive to incidence angle.

The long period data are presented in Figures 13 and 14. As is expected for these longer period waves, the amplitudes demonstrate less overall variation between stations and less individual station scatter. The striking features of the long period P data in Figure 13 are the very low amplitudes of OXF, which deviate from the regional trend, and the enhancement of DAL, FLO, COR and DUG. The total range in amplitude variation is only a factor of two, which indicates remarkable stability of the WWSSN long periods over the 12 years spanned by the data, and the relative insensitivity of long period P to receiver or attenuation variations. The long period SH amplitude pattern (Figure 14) has

larger variation between stations, with very low amplitudes at TUC, FLO, OXF and WES, and slight amplitude enhancement at LUB and JCT. The anomalously low long period amplitudes of OXF produce much of the amplitude ratio anomaly noted by Burdick (1978).

To facilitate comparison of the data sets, the mean and standard error of the mean were computed for each station and the P and SH observations plotted together. The relative mean values are given in Table 2. The short period data are shown in Figure 15. The two data sets have been linearly scaled to minimize the separation between them. The first order feature apparent in the data is that the short periods track one another well, with high amplitudes in the central states and fairly similar behavior in the eastern stations. In the West, the amplitudes of SH are consistently lower than the P waves relative to the rest of the pattern, with the notable exception of ALQ. The western station SH waves are the lowest in the SH pattern, though GSC appears to record higher amplitudes than the Rocky Mountain stations. The stations DUG and BOZ show the greatest discrepancy between P and SH behavior, though there is only a single SH data point at BOZ.

The long periods are plotted in Figure 16, and again the outstanding feature is the degree to which they track. The largest deviation in behavior occurs at FLO, where the SH waves are much lower amplitude, but there is only a single P measurement. Comparing the short period and long period P variations shows that the patterns track closely though LUB, JCT and RCD do not show long

period enhancement. The short periods at GSC, BOZ, SCP and OGD do not track the long period trends very closely either, but the rest of the stations do. The SH data also show strong correlation between short and long period behavior except for FLO, OXF, OGD and ALQ.

The tendency of the amplitudes of the various phases to track one another is the type behavior expected for attenuation controlled amplitude variations. The more highly attenuated SH phases have larger regional variations than the P waves, consistent with the commonly accepted notion that attenuation is dominated by losses in shear. There are clearly substantial amplitude variations introduced by receiver structure, and the order of magnitude of these variations may actually exceed the scatter between some stations. Consider the short period high amplitudes recorded at midwestern and southern stations, or the long period behavior of FLO and OXF, both of which are such large effects that they dominate the amplitude patterns. Butler & Ruff (1980) examined various receiver structure induced amplifications for short period P waves and concluded that a factor of two amplification due to horizontal sedimentary layer receiver structures is reasonable. This may account for much of the midwestern short period anomaly, but cannot explain all of the other variations. The long period variations at OXF are very large, and it is difficult to assess what type of receiver structure may be responsible, though it must clearly have

three-dimensional structure. The fact that the short periods are relatively enhanced and have complicated high frequency codas, indicates that this anomaly is no conventional attenuation effect. The azimuthal behavior of this anomaly will be considered in a following paper, but it should be noted here that the occurrence of the anomaly at adjacent stations FLO and OXF in the Mississippi Valley indicates that this is not the result of faulty instrument emplacement. The studies of Burdick & Langston (1977); Langston & Blume (1977); and Langston (1977,1978) have detected the presence of dipping structures beneath the stations LON and COR utilizing long period body waves. Due to their distance from Argentina these stations were not used in the SH analysis, but COR does show slight long period and short period P enhancement, while LON does not.

4 Source variations

The similarity of focal mechanism and earthquake location of the data presented above requires some consideration of possible systematic variations in amplitudes that are not due to attenuation in the upper mantle. Though we can fairly well rule out systematic trends due to focal mechanism based on the tests mentioned above, there remains a question about bias introduced by comparing Peruvian and Argentinan sources, and in utilizing data from a narrow range in depth. The selection of deep events for analysis helps reduce potential bias due to source region slab

heterogeneities that could cause azimuthal variations. In Figure 17 we compare the short period P wave data collected in this study with the results compiled by Butler et al. (1979), which includes 15 shallow and intermediate depth South American events, the majority of which were located in Argentina, well south of the deep data epicenters. The amplitude behavior is very similar at the eastern and central stations. OXF shows greater enhancement for the shallow events, and the single shallow source reading for BOZ is low relative to the three readings from deep events, which may be due to receiver effects. Taking the ratio of the mean values shows that there is at most a factor of two total range in relative amplitude behavior for stations for which there is adequate data. The only systematic trend in the ratios is a tendency for BKS, COR, LON and GSC to record slightly larger amplitudes from deep sources, but this is not a robust feature, and of little import for this work as these stations were generally too distant to obtain SH data, with the exception of GSC. There is certainly no obvious systematic source bias such as that found between the much closer together test sites in southern and northern Novaya Zemlya by Butler & Ruff (1980).

As another test, we divided the long period P data set into Peruvian and Argentinan subsets, and checked for systematic source effects. The data are shown in Figure 18, and though the statistical samples are small in this case, there is again no systematic source region effect apparent in the data. The

amplitude ratios show no resolvable trends and a total range in variation of two, though much less variation for the majority of stations. The similarity between patterns from different depths and source regions allows some confidence in comparison of the amplitude patterns in this paper.

In the foregoing discussion of the amplitude data set, several instances of probable receiver dominated effects have been mentioned. Comparison of the amplitude patterns with data from other azimuths identifies stations with rapidly varying behavior, which is more likely due to near receiver effects, but a detailed comparison of all the phases presented here with other azimuths is beyond the scope of this paper. Butler et al. (1979) have considered the azimuthal sensitivity of short period P waves, and their conclusions indicate substantial azimuthal receiver effects at LON and BOZ, and lesser variations at most other stations. However, the general pattern of short period amplification in the central states is common to all azimuths, as is the lack of strong east to west amplitude differences. It should be noted that Butler & Ruff (1980) found that scattering, as reflected in the horizontal components of the P waves, does not appear to be responsible for the short period P amplitude pattern.

5 Attenuation models

The question of upper mantle regional variations and frequency

dependence of attenuation is a controversial and fundamental problem in seismology. Most studies have failed to overcome the uncertainties in source frequency content, receiver and propagational effects, and instrumental band width to present unambiguous conclusions. In this study we primarily sought to test compatability and consistency of a range of attenuation variations with the South American body wave data set, rather than to make absolute attenuation determinations that require complete knowledge of the source and receiver. In order to accomodate the tradeoffs between source time function and attenuation, a broad range in source frequency content was considered, as was a broad range in long period absorption band amplitudes, in the time domain modeling. Comparison with the observed short period and long period data places bounds on these variables, however there is inevitably an uncertainty in reference baseline selection. To partially circumvent this, we utilized results from the literature as well as performing an absolute t^* determination, to minimize arbitrariness in baseline selection. Spectral ratio techniques were also employed and compared with the time domain models.

Two basic attenuation models have been considered, with only single parameters varied in each. These are the constant Q , Futterman model (Futterman 1962) parameterized by

$$t^* = \int_{\text{Ray path}} Q^{-1}(s) V^{-1}(s) ds,$$

and a frequency dependent model parameterized by a high frequency roll-off of the relaxation spectrum given by τ_m^{-1} (Minster 1978a,b). In the latter model, for small attenuation, Q is given by

$$Q^{-1}(f) = \frac{2}{Q_m} \tan^{-1} [2\pi f(\tau_M - \tau_m) / (1 + (2\pi f)^2 \tau_M \tau_m)] \quad (2)$$

Q_m is the quality factor in the absorption band. The points at which Q^{-1} is half of its maximum value occur for $f = (2\pi\tau_M)^{-1}$ and $(2\pi\tau_m)^{-1}$. We fix τ_M at 1000 sec, well outside the body wave pass band of our data in all of the models. The constant Q model has frequently been adopted in time domain modeling of body waves at frequencies lower than 1 Hz, and we tested how well this model agrees with our broad band data. The motivation for considering the frequency dependent model are the numerous observations of body wave frequencies that would not be seen if the long period Q values applied to high frequencies as well (eg. Der et al. 1980; Sipkin & Jordan 1979). It is shown in the present study that τ_m values which allow two to four Hz energy to pass through the mantle have strong effects on one to five second energy as well. Thus, the WWSSN signals constrain acceptable frequency dependence of this type. The long period level of the absorption band is described by $t^* = t_u/Q_m$ with t_u being the travel time, whether or not a τ_m roll-off is present. When τ_m is allowed to vary, the t^* level is held constant so that the model amplitude variation can be attributed entirely to changes in τ_m . It is, of course, possible

to permit greater flexibility by simultaneously varying both t^* and τ_m , but this is not warranted by our data, because we are presently unable to correct for the broadband receiver effects. Thus, this analysis does not encompass attenuation models such as proposed by Solomon (1972), which would allow short periods and long periods to vary independently. Such a refined model must await a better understanding of the long period amplitude and spectral effects apparent in the data.

The test of constant Q attenuation was performed as follows. Large amplitude, high frequency P and SH data were compared with synthetics generated for a wide range of source models and t^* values. We adopted the constraint that relative P and SH absorption is given by $t_p^*/t_s^* = 4.0$, which implicitly assumes that all attenuation is caused by shear losses in the mantle (Anderson et al. 1965). The validity of this assumption is based on absolute t^* determinations such as those by Helmberger & Hadley (1980) and Burdick (1978), and relative attenuation measurements by Solomon & Toksöz (1970) and Der et al. (1980), as well as much work done with surface waves. Once the high frequency, high amplitude signals were adequately modeled by a particular source model and t^* combination, differential attenuation, Δt^* , was applied by convolving this synthetic signal with Futterman filters, and the amplitude behavior relative to the initial model was recorded. As a consistency check, the data that were initially modeled were carefully digitized and also convolved with differential

attenuation filters, and the amplitude decrease relative to the original signal was recorded.

Figure 19 shows some of the results of this procedure, where the amplitude behavior relative to the initial signal (either reference synthetic model or digitized data) are plotted versus Δt_{α}^* . The curves for LP SH and SP SH were generated using $\Delta t_{\beta}^* = 4\Delta t_{\alpha}^*$. The theoretical source time functions for all of the models shown except for SP P was a symmetric triangle of one second total duration. The initial t^* for the LP P model was $t_{\alpha}^* = 0.8$ sec, while it was $t_{\beta}^* = 3.2$ sec for the short and long period SH models, though as discussed below, the absolute t^* values are not important in this procedure. Selection of this source model produced short period to long period SH amplitude ratios in the range of those observed in the data. The deep short period P waves were from smaller events of considerably higher frequency content, and a shorter source time function of 0.2 seconds duration and $t_{\alpha}^* = 0.6$ sec were used for the synthetics. Also shown in Figure 19 are the curves representing the decrease in amplitude when digitized data were convolved with differential attenuation operators. Other data produced very similar curves. These curves are closely tracked by the models, which indicates that the initial source frequency content has been adequately accounted for. For a fairly large range of alternate source model- t^* combinations similar curves can be found that fit the data behavior as well as the ones shown. Thus, this procedure is relatively insensitive to the particular

parameters selected; what is important is the initial frequency content. For the range in Δt_Q^* shown, the synthetic waveforms are compatible with the range in observed waveforms and amplitudes. As has been shown by Butler and Ruff (1980), the change in waveform is not dramatic even for large changes in t^* . This waveform behavior is discussed in more detail below.

Figure 19 clearly shows the insensitivity of long period waves to even large changes in t^* . The range in Δt_Q^* required to span a factor of two in amplitude is 1.5 sec for long period P, which produces a factor of 3 variation in long period SH amplitudes. Thus, it is clear that regional variations of at least this range would have to be invoked to explain the observed amplitude behavior with constant Q attenuation alone. The relative amplitude behavior of the short periods is, of course, much more sensitive to Δt_Q^* , at least as long as there is significant high frequency information in the signal. It is clear from the short period P and SH curves that it will be possible to explain the observed tendency of short period P and SH amplitudes to track closely by Δt^* variations, and that for the total range in Δt^* needed to fit the observed long period amplitude variations, the predicted short period variations would be on the order of two times greater than observed.

A somewhat different procedure had to be adopted for the frequency dependent modeling. The effect of increasing τ_m from zero is to permit progressively more high frequencies to pass through to the receiver, thus the amplitudes increase with τ_m . To

accommodate this, we initially modeled low amplitude, low frequency content data, and then for a range in t^* and source time functions, generated synthetics with attenuation operators that had increasing τ_m . The amplitude behavior was then measured relative to the $\tau_m = 0.001$ sec reference, at which the attenuation operator is very similar to a Futterman operator with the same t^* . τ_m was allowed to increase only as much as would still generate waveforms compatible with the data. This restricted the acceptable range in t^* , τ_m , and source models significantly, unlike for the constant Q model, where t^* and reasonable source duration trade off almost indistinguishably.

The results for short period P waves are shown in Figure 20, where the amplitudes were measured relative to a signal generated with a Futterman operator of the t^* value indicated for each curve. The upper three curves were generated by delta function sources convolved with short period instrument and attenuation operators of varying t^* and τ_m . These demonstrate the predictable result that the relative enhancement in amplitude caused by increasing τ_m is greater for larger values of t^* , at least over this small range in t^* . The lower curves show that for a given t^* , the higher the source frequency content the higher the relative enhancement caused by increasing τ_m . The three triangular source models and t^* values of 0.6-0.8 sec produce waveforms spanning the range of observed variations for a given event when τ_m varies from 0.001 to 0.25 sec. This variation in frequency dependence produces a range in

amplitudes of a factor of from 4 to 6, comparable with the range observed in the data. Larger values of τ_m produce unacceptably spiky waveforms for almost any simple triangular, trapezoidal, or Savage-type (Savage 1966) source model.

Figure 21 shows a similar set of curves for short period SH waves, which have somewhat more complicated behavior. Again, the upper three curves were generated using a delta function source, but over the range in τ_m up to 0.3 sec the effect of varying t_p^* is different than for the P waves. This is because of the greater range in t_p^* considered, as well as the larger dominant period of the SH waves due to longer source duration and higher t_p^* . It takes larger τ_m values to overcome the longer period nature of the source, but there is ultimately a cross-over point at which the relative enhancement of the higher t_p^* signal is greater, as seen in models (B) and (C). For the trapezoidal and Savage source models examined, fairly large t_p^* values of around 3.0 sec and τ_m values less than 0.25 sec were in best agreement with the data. For the January 17, 1967 and the December 20, 1966 events we checked the source models shown in Figure 21 against several short period and long period P signals to help constrain the range of acceptable models. For $\tau_m > 0.25$ sec the models shown in Figure 21 no longer produce acceptable short period P waves. For lower reference t_p^* values one requires source models that look like Futterman type attenuation filters to produce acceptable waveforms for any substantial range in τ_m , but there is again the unavoidable

tradeoff between reference t^* and source time function. Because simple trapezoidal time functions and Savage source models are theoretically reasonable source representations for these small, deep events, we tend to prefer the results with $t_\alpha^* = 0.8$ sec and $t_\beta^* = 3.2$ sec.

The variation in long period amplitudes produced by varying t_m from 0.001-0.25 sec are relatively minor, as shown in Figure 22. For $t_\alpha^* = 0.8$ sec, and a source time function that adequately models the long period P waves, only an amplitude variation of a factor of 1.2 is produced. The same is true of models that fit the SH data, regardless of source model or t_p^* selection. It is apparent that changes in the long period level of the relaxation spectrum produced by varying t^* are the only way to produce significant amplitude variations for these periods.

To more clearly compare the results of the synthetic models, we have employed the procedure of utilizing the data from one phase to predict the amplitude variations of another. This tests the degree to which the amplitudes behave consistently with lateral variations in attenuation. The curves in Figures 19-22 are readily adapted to this procedure. Figure 23 shows the results of predicting the short period SH data from the short period P data. The Δt_α^* results were obtained by taking the P wave amplitude means and determining the Δt_α^* for each station, as read from Figure 19, that yield the variations from the highest amplitude station. Then, for each station, that Δt_α^* predicted the short period SH

behavior from Figure 19, and the relative amplitudes were plotted against the data. Similarly, Figure 20 was used to determine the τ_m values needed to produce the P pattern, assuming a particular source model (in this case source D), and the resulting τ_m values then predicted the SH amplitudes using Figure 21 and a particular source model (in this case C with $t_B^* = 3.2$ sec). The predicted amplitudes were plotted with the data after linearly minimizing the scatter at each station. Some stations may lack either observed values or predictions depending upon the coverage of the data set.

It is clear that in a gross sense, describing the P wave amplitude variations by either Δt_α^* or a range in τ_m predicts the general behavior of the SH amplitudes. This basically reflects the fact that the amplitudes track, but more specifically, the degree to which they track is similar to what would result from lateral variations in attenuation. Of course, it may be possible to find receiver functions that produce a similar effect. The more important point to draw from Figure 23 is that the short period WWSSN amplitudes do not clearly demand nor preclude frequency dependent attenuation. Since the amplitude and waveform behavior have been checked for consistency throughout, this means that one cannot base a strong argument for or against frequency dependence from the short period amplitude and waveform behavior alone. There do not appear to be any strong trends in the data which suggest alternate selection of t_B^*/t_α^* ratios or different reference models.

Though this is clearly a crude prediction, it is of some

interest to look at the individual stations a bit closer. The predicted amplitudes at DUG are larger than observed in the SH data, but this station does have complex SH wave codas which may account for this discrepancy. BOZ has strong azimuthal variations (Butler & Ruff, 1980) and only a single SH observation, so its behavior is uncertain. Though ALQ is predicted to be on the level of TUC and GOL, which are well predicted, the SH amplitudes are significantly higher than expected. It is difficult to explain this with attenuation. Both OXF and AAM have high frequency SH waves and complex codas and their amplitudes are underpredicted. Most other stations, with the exception of SCP, are quite well predicted.

The entire analysis described above was reproduced using the shallow and intermediate depth P data of Butler *et al.* (1979) with appropriate source models and Δt_Q^* , and τ_m behavior as in Figures 19 and 20. The predictions are very similar, despite the longer period sources and different t_Q^* level of the shallow data. Table 3 list the Δt_Q^* and τ_m results for each P data set. The similarity in numbers indicate the stability of the process. The actual numbers are intended only to convey an appreciation of the range in Δt_Q^* and τ_m indicated by the data, and are not to be taken as definitive station values.

Since the short period data is most subject to large receiver effects, we have also attempted to test the compatibility of the long period P and SH data, and the relative short period-long

period behavior. By analogous procedures to those outlined above, we have predicted long period P using the long period SH amplitudes, with the results shown in Figure 24. Only a Δt^* comparison is made, since the frequency dependent model cannot encompass the observed amplitude variations. The resulting values of Δt_p^* are listed in Table 3, and as expected, a large range in Δt_p^* is necessary to explain the data. The interesting result is that the P waves are rather well predicted, with the largest single exception being at station FLO. This success attests to the degree to which the long periods track compatibly with constant Q attenuation and $t_p^*/t_q^* = 4.0$.

In order to compare the short period data with the long periods, we have exercised some selectivity in only comparing the East Coast and Rocky Mountain stations after determining Δt_q^* and Δt_p^* values appropriate for this subset. This was done to avoid the large anomalies of OXF and FLO, as well as the probable sediment amplifications in the central states. This also enables a more direct comparison of the two subregions. In Figure 25 the predicted long period P amplitudes for these subregions are shown, with the success of the prediction even better than for the entire data set. Figure 26 shows the predicted short period P amplitudes based on the long period SH Δt_p^* values, and Figure 27 shows the predicted short period SH pattern. There is a fair amount of mismatch in the P data, with several eastern stations being overpredicted, whereas the SH data is slightly better matched.

Considering the crudeness of this test, the most important feature is that although the details of the short period amplitude patterns are not well predicted, the overall range of amplitudes is not adequately outside the range of possible receiver effects to argue for frequency dependence. The Δt_{β}^* values involved in this comparison are as large as 3.8 sec, so with smaller variations expected if receiver corrections for long periods can be applied, it seems that the fit will only improve. In this test it is difficult to argue against frequency dependent effects as well, for allowing for a second parameter in the attenuation variations enables one to fit the short periods much better, but this is not warranted by the present data set, at least until receiver corrections can be made.

The effects of possible frequency dependence are important to resolve, because if there is an appreciable impact on the short period WSSN amplitudes, then source studies or nuclear test yield estimates using these phases may be in significant error. Consider Figure 28 which shows short period P wave data from one of the shallow events studied by Butler & Ruff (1980), and some of the short period SH waves from this study, as well as synthetics produced by varying t^* and τ_m . The source time function for all of the synthetics was a symmetric triangle of one second duration, with a reference $t_{\alpha}^* = 1.0$ sec and $t_{\beta}^* = 3.2$ sec. Values of τ_m and Δt^* are shown which produce relative amplitude variations in increments of 20% of the initial source model amplitude. This

demonstrates two important features; one is the relatively minor change in waveform even for large values of Δt^* , and the other is the difference in resulting waveforms for the same source time function and t^* level between the constant Q and frequency dependent models. The frequency dependent synthetics could be brought into closer agreement with the data by using a longer period source time function or larger t^* values, which indicates the importance of attenuation in constraining the source model. When a longer period source is adopted, the waveform changes for the frequency dependent operator are as small as shown here for the constant Q model. The P and SH data shown in this figure do suggest a correlation between low frequency content and low amplitude, such as would be expected for either attenuation model, but this is not always observed in the data.

Since the details of the amplitude variations in the frequency dependent attenuation synthetic tests are sensitive to the t^* of the long period relaxation band, the value of which is controversial, we attempted to make an absolute t^* determination with the best quality event in this data. The event of January 17, 1967, was selected for this, and in addition to the SH data, several high quality short period P and long period P waves were digitized. We examined the body wave complexity and stability at all azimuths for this event and found it to be as simple and well behaved as any we know of. This simplicity and the small size of the event allowed us to assume that the source spectrum is almost

the same for the P and S waves, and thus that the only difference between P and S observed at a station is in absolute amplitude governed by relative radiation pattern and geometric spreading, differential attenuation between P and S, and differences in receiver effects which are assumed to be small. This point source approximation tends to bias t^* determinations toward larger values, but given the small size of the source involved this effect should not be excessive. Then, having interpolated, detrended and tapered the data, we applied differential attenuation operators to the short period and long period P waves, varying the attenuation until the amplitude ratio $R_{sp,lp} = \max \text{amp}|(\text{filtered } P_{sp,lp})| / \max \text{amp}|SH_{sp,lp}|$ was the same for short periods and long periods. In determining the optimal differential attenuation operator we actually minimized the parameter γ :

$$\gamma = [(C_{sp} + C_{lp})^2/4 + (1 - |R_{sp} - R_{lp}|^2/|R_{sp} + R_{lp}|)^2]$$

where C_{sp} is the normalized cross correlation coefficient of the filtered short period P and observed short period SH signals, and C_{lp} is the corresponding long period normalized cross correlation coefficient. This parameter, γ , which is just a weighted measure of the crosscorrelations and amplitude fits, weights the amplitude match more heavily, but provides some control on the waveform fit. The normalized cross correlations were only performed over the first pulse half cycle in order to increase their sensitivity.

Given the differential operator that minimizes γ , $F^* = t_\beta^* - t_\alpha^*$, one can determine the absolute value of t_α^* under the assumption that $t_\beta^*/t_\alpha^* = 4.0$.

The results of this experiment are listed in Table 4, and shown in Figure 29. For North American stations, the results indicate t_α^* values of 0.8-1.0 sec, with smaller values yielding very poor amplitude agreement. It must be emphasized that this procedure is most sensitive to the dominant periods of the signals, and hence the resulting t^* values are appropriate for the period range 5-20 sec. Note that the P waveforms map rather well into the SH waveforms, over at least the first few cycles. This indicates the validity of some of the assumptions made in the procedure. The resulting t_α^* values are in agreement with the results of Burdick (1978), who included this event in his determination of $t_\beta^* = 3.2$ sec. Two African stations were also included in this test, PRE and SDE, with the interesting result that $t_\alpha^* = 0.4$ sec. This may indicate significant differences in upper mantle attenuation beneath southern Africa relative to the United States, or possible directivity effects in the source. The comparison for the African stations was done with SV phases since the radiation pattern is more favorable for these than for SH, as shown in Figure 2.

We also allowed the differential attenuation operator to be frequency dependent according to equation (2). Selecting a reference t_α^* of 0.8 sec and varying t_m produced the t_m values in Table 4 and the waveforms in Figure 29. The BLA amplitudes were

not well fit, though the short period normalized cross correlation is better in this case, and ALQ and a similar station DUG are only well fit with higher t_{α}^* values so that no τ_m values were determined for the latter two. The fits at the remaining stations are as good or better than for the Futterman operator test, and the African stations have predictably high τ_m values. This procedure is certainly subject to errors in digitization and the assumptions about the source and receivers. However, the consistency of the results indicates the validity of selecting long period t_{α}^* values of 0.8-1.0 sec as done in the time domain modeling.

6 Spectral ratio results

As an alternate approach to determining Δt^* variations, we have applied the spectral ratio technique to the long period SH data presented above. Under the assumption that the source spectrum and receiver crustal transfer functions are the same, the spectral ratio of a given body wave phase recorded at two stations by similar instruments will be a function only of the differences of attenuation along the raypaths:

$$\frac{A_i(f)}{A_j(f)} = g_{ij} \frac{\exp(-\pi f t_i^*)}{\exp(-\pi f t_j^*)}$$

where g_{ij} is a constant correction for geometric spreading. Thus, the natural log of the spectral ratio gives

$$\log(A_i(f)/A_j(f)) = C_{ij} - \pi f \Delta t_{ij}^*$$

Fitting a line to the log ratio enables determination of Δt_{ij}^* . For the distance range between stations in this data, existing laterally homogeneous Q_B models would not cause more than a Δt_B^* variation of 0.1 sec (Lundquist 1980), so we did not attempt to correct the values to any standard model as was done by Solomon & Tcksöz (1970). Data with ScS in the time window were excluded. We took a 30 second time window interpolated in .25 second intervals, removed the mean, detrended the signal, tapered it with a 10% cosine termination filter, and computed the spectra. The spectra were smoothed with a running mean over a window of 0.04 Hz, and the log ratio was performed and divided by π . A least squares fit line was found for the ratioed spectra. We performed the analysis with and without Gaussian band pass filtering and found that filtering the data does not significantly affect the results. However selection of the frequency window over which to fit the ratio slope does. The spectral slopes were determined over the period range from 6.3 to 25.0 sec. In Table 5a we list the resulting values of Δt_B^* and standard deviations with SCP as the reference station. A band pass filter of 1-60 seconds has been applied to eliminate digitization and rotation noise. The spectral ratios for the January 17, 1967 event are shown in Figure 30, and the exceptional quality of the ratios indicates that source complexity and receiver

structure are not important in the bulk of the data. This figure is representative of the quality of all of the ratios obtained. Several stations, BEC, SHA and TUC do have strong second arrivals which produce some interference peaks in the ratios, and the results for these stations are questionable. In Table 5b we list the results for data band pass filtered over the band 5-25 seconds, with the spectral slopes being determined between 6.7 and 14.3 seconds. The choice of band pass filter again does not significantly affect the results. The variation in resulting values indicates the sensitivity of such measurements, and the degree of arbitrariness in selecting a suitable window.

These Δt_p^* determinations can be compared with the time domain results above and the results of Solomon & Toksöz (1970), bearing in mind that we have removed a factor of π from our slopes and errors. Despite the very smooth ratios that were found, there is significant scatter between events. Table 6 lists the mean Δt_p^* values and standard deviations from this study, as well as the results of Solomon & Toksöz (1970) with π removed from their numbers. The large Δt_p^* values at ATL and OXF are consistent with the broadened nature of their waveforms mentioned above. GOL, even after instrument correction, has a very large Δt_p^* value, as does WES. The remainder of the data have Δt_p^* values within the range found in the time domain analysis, but a comparison with Table 3 shows that individual stations, such as FLO, GEO and OGD, have significantly different results. Comparison with Solomon &

Toksöz's (1970) values, which were dominated by SV determinations, shows similar large deviations. The high quality of this data set indicates that the new results are more reliable, but they are still subject to large error bars. There is no obvious east-west trend in the results listed in Table 6.

In order to provide a comparison of the observed long period SH amplitude behavior with that predicted from the spectral ratio results, we utilized Figure 19 to predict the relative amplitude patterns from the results in Table 6. Figure 31 shows the observed data versus the pattern predicted for the Δt_{β}^* values from Table 5b, and from the results of Solomon & Toksöz (1970). There is large scatter in the comparison, which suggests that either the spectral ratio techniques are subject to large uncertainties due to the processing procedure, or the amplitudes are profoundly affected by receiver structures which do not disturb the amplitude spectra excessively. Probably the reality is somewhere between, with it being very difficult to dismiss the spectral evidence for attenuation variations, but almost equally difficult to accommodate as much amplitude variation as needed for receiver functions affecting 15 second period SH waves.

7 Discussion

The tests discussed above were not intended to provide accurate station dependent attenuation operators, rather they

quantitatively examine the consistency of the relative amplitude behavior throughout the band from one to 20 seconds. The variations in long period behavior are perhaps the most suggestive that attenuation mechanisms are controlling the basic patterns, with the P and SH amplitudes behaving as predicted for constant Q and $t_p^*/t_q^* = 4.0$. The discrepancy between the time domain amplitude results and the spectral analysis do argue for very large receiver effects in the long period data, though the P and SH relative behavior must still be similar to that produced by attenuation. Either technique indicates very large Δt_p^* values of up to 4.0 sec, consistent with the general results of Solomon & Toksöz (1970), but we feel that a better understanding of the frequency dependent receiver effects will lower this range considerably. We plan to address this problem by considering azimuthal sensitivity and details of the signal codas in a forthcoming paper.

Time domain modeling of the short period P and SH waves does not unambiguously resolve frequency dependence of attenuation, emphasizing as it does the longer periods in the waveform. Analysis of short period and long period digital recordings may permit more reliable spectral analysis of the broadband frequency content, but we did not attempt such techniques for this data because of the limitations of hand digitization and rotation of the SH components. To first order, the P and SH amplitude behavior is consistent with regional variations in attenuation, with many of the deviations from this consistency occurring at stations with

apparently large receiver effects. As it may be possible to generate receiver structures which produce relative P and SH amplitude behavior similar to that seen for the attenuation models, this does not conclusively indicate variations in attenuation, rather it demonstrates the general consistency of that hypothesis.

Comparison of the the long period and short period amplitudes indicates that individual station behavior is not entirely consistent with frequency independent attenuation variations, although the inconsistency is not such as to make a strong case for frequency dependence. The range of Δt^* values found by Der et al. (1980) for LRSM stations is within the range determined by our amplitude data, however we do not detect systematic regional differences between the East Coast and Rocky Mountain stations. The structure of the short period amplitude pattern is dominated by the high amplitudes of the central states, as also demonstrated by Butler & Ruff (1980). Much of this trend can be attributed to amplification due to sedimentary receiver structures, and further analysis is required to resolve to what degree differential attenuation governs the pattern.

Conclusions

Short period and long period P and SH amplitude patterns across WWSSN stations in the United States tend to follow the same regional trends. Stations in the midwestern and Gulf Coast states

record large amplitude, high frequency arrivals relative to the East Coast and Rocky Mountain stations which record similar amplitudes. Efforts have been made to ensure that the amplitude data is free from bias due to source complexity, radiation pattern, and source structure. The relative amplitude and waveform behavior are tested for compatibility with constant Q and frequency dependent lateral attenuation variations. The short period data do not resolve frequency dependence of attenuation, and either a range of Δt_{α}^* of 0.5 sec or a range in τ_m of 0.001-0.25 sec proves consistent with the P and SH behavior in the 1 to 7 sec period range. The long period amplitude behavior is very consistent with Δt_{β}^* variations of up to 4.0 sec, though these variations are not in close agreement with the short period amplitudes. A technique was developed for determining absolute t_{α}^* levels, and indicates that $t_{\alpha}^* = 0.8-1.0$ sec for the deep South American events recorded in North America, and lower t_{α}^* values or significant frequency dependence for paths to southern African stations. Spectral analysis of the long period SH data yields a large range in Δt_{β}^* of up to 4.0 sec, though the individual station values are not in close agreement with the amplitude behavior. This indicates that long period receiver functions may be more significant than usually assumed.

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Figure Captions

- Figure 1. A gnomonic projection (all great circles are straight lines) showing the epicenters of the deep events in Peru and Argentina and the location of WWSSN stations used in this study.
- Figure 2. P and SH equal area nodal projections in the lower hemisphere for some of the events used in this study. For each event, projections of stations at the extremes of azimuth and distance for the data from that event are shown.
- Figure 3. Rotated short period SH components of the December 9, 1964 event recorded at WWSSN stations in the United States. Peak amplitudes are normalized to unit amplitude.
- Figure 4. Short period and long period SH components of the March 5, 1965 event. Only stations ALQ and TUC have long period response peaked at 15 seconds, the rest are peaked at 30 seconds. The numbers given are the gain corrected short period to long period amplitude ratios.
- Figure 5. Short period and long period SH components and amplitude ratios of the December 20, 1966

event. All stations except GOL have long period response peaked at 15 seconds for this and the following figures.

Figure 6. Short period and long period SH components and amplitude ratios of the January 17, 1967 event.

Figure 7. Short period and long period SH components and amplitude ratios of the September 9, 1967 event.

Figure 8. Short period and long period SH components and amplitude ratios of the July 25, 1969 event.

Figure 9. Short period and long period SH components and amplitude ratios of the January 3, 1973 event.

Figure 10. Short period SH components of the February 11, 1975 event.

Figure 11. Relative amplitudes of short period P waves from deep events in Peru recorded at WWSSN stations across the United States.

Figure 12. Relative amplitudes of short period SH waves from deep events in Argentina recorded at WWSSN stations across the United States.

Figure 13. Relative amplitudes of long period P waves from deep events in Peru and Argentina recorded at WWSSN stations across the United States.

Figure 14. Relative amplitudes of long period SH waves from deep events in Argentina recorded at WWSSN

stations across the United States.

Figure 15. The mean and standard error of the mean of the short period P and SH amplitudes from South America recorded at WWSSN stations across the United States.

Figure 16. The mean and standard error of the mean of the long period P and SH amplitudes from South America recorded at WWSSN stations across the United States.

Figure 17. Comparison of the means and standard errors of amplitudes of short period P waves from deep and shallow earthquakes in South America recorded at WWSSN stations across the United States. The shallow source amplitude data is from Butler et al. (1979).

Figure 18. Comparison of the means and standard errors of amplitudes of long period P waves from deep events in Peru and Argentina.

Figure 19. Amplitude of body waves versus Δt_{α}^* . The amplitudes are given relative to the reference signal amplitude of either observed data or synthetic model with initial reference attenuation given by t_1^* . Curves B and C and the LPSH and SPSH curves are generated with $\Delta t_{\beta}^* = 4\Delta t_{\alpha}^*$. The similarity in behavior of the

attenuated data and corresponding model indicates that the models have an appropriate spectral content. The synthetic time functions are trapezoids described by $T(\text{rise time, constant duration, fall time})$ in seconds.

Figure 20. Amplitude of short period P wave synthetics generated with frequency dependent attenuation operators. For each curve t^* is held constant and τ_m is varied. Curves (A) are generated with delta functions and the indicated attenuation operators and the other curves are generated with trapezoidal sources. All signals are convolved with a short period WSSN instrument response $I_{sp}(t)$. The amplitudes are given relative to the $\tau_m = 0.001$, Futterman operator amplitude.

Figure 21. Amplitude of short period SH wave synthetics generated with frequency dependent attenuation operators. For each curve t^* is held constant and τ_m is varied. Curves (A) are generated with delta functions and the indicated attenuation operators, and the other curves are generated with trapezoidal sources. All signals are convolved with a short period WSSN instrument response $I_{sp}(t)$. The amplitudes are

given relative to the $\tau_m = 0.001$, Futterman operator amplitude.

Figure 22. Amplitude of long period P wave ($t^* = 0.8$ sec) and SH wave ($t^* = 3.2$ sec) synthetics generated with frequency dependent attenuation operators, as in Figures 20 and 21. $I_{1p}(t)$ is the long period WWSSN instrument response.

Figure 23. Observed short period SH amplitudes versus predicted SH amplitudes utilizing the station values of Δt_α^* and $\Delta \tau_m$ which reproduce the short period P amplitudes.

Figure 24. Observed long period P amplitudes versus predicted P amplitudes utilizing the station values of Δt_β^* which reproduce the long period SH amplitudes.

Figure 25. Observed long period P amplitudes versus predicted P amplitudes utilizing the station values of Δt_β^* which reproduce the long period SH amplitudes in East Coast and western stations.

Figure 26. Observed short period P amplitudes versus predicted P amplitudes utilizing the station values of Δt_β^* which reproduce the long period SH amplitudes in East Coast and western stations.

Figure 27. Observed short period SH amplitudes versus predicted SH amplitudes utilizing the station values of Δt_g^* which reproduce the long period SH amplitudes in East Coast and western stations.

Figure 28. Amplitude and waveform variations versus Δt^* and changes in τ_m for short period P and SH. The short period P data are from a shallow Argentinan event, and the SH data are from a deep Argentinan event. Synthetics for a triangular source time function of one second duration are shown with the reference $t_a^* = 1.0$ sec for the P synthetics and $t_b^* = 3.2$ sec for the SH synthetics. The variation in attenuation needed to produce a factor of five range in amplitudes for each case is shown. The frequency independent models provide a good fit to the data, whereas a longer time function or larger reference t^* would be needed for the frequency dependent models to match the data.

Figure 29. Determination of t_a^* by obtaining a differential attenuation operator which maps the P waves into the SH waves. The t_a^* values which produce the best fitting SH synthetics are shown along with the synthetics. The τ_m

value which produces the best fitting SH synthetics assuming $t_{\alpha}^* = 0.8$ sec are also shown with the corresponding synthetics.

Figure 30. Spectral ratios of long period SH waves, with station SCP as reference, for the January 17, 1967 event. The numbers shown are the least square determinations of the spectral ratio slope with one standard deviation. The data were bandpass filtered with a Gaussian filter of (1,60) seconds.

Figure 31. Comparison of observed long period SH amplitude behavior with that expected given the Δt_{β}^* values determined by spectral ratios in this work and by Solomon and Toksöz (1970).

Table 1. Earthquake data set.

Date	Origin Time	Lat., °S	Lon., °W	Depth, km	m_b	Phase
9 Dec. 1964	13:35:42.4	27.5	63.2	586.	5.9	SH(sp)
22 Dec. 1964	00:24:48.7	9.5	71.3	614.	5.3	P(sp)
18 Feb. 1965	22:32:19.6	9.9	71.2	594.	5.2	P(sp)
5 Mar. 1965	14:32:19.2	27.0	63.3	573.	5.5	SH(sp,lp)
3 Nov. 1965	01:39:03.1	9.1	71.4	593.	6.2	P(lp)
20 Dec. 1966	12:26:54.6	26.1	63.2	586.	5.7	SH(sp,lp)
17 Jan. 1967	01:07:54.3	27.4	63.3	588.	5.6	SH(sp,lp)
27 Mar. 1967	08:26:34.6	8.9	71.4	605.	5.3	P(sp)
9 Sep. 1967	10:06:44.1	27.7	63.1	578.	5.8	SH(sp,lp)
11 Oct. 1967	20:28:10.2	10.3	71.2	585.	5.0	P(sp)
23 Aug. 1968	22:36:51.3	22.0	63.5	537.	5.8	P(lp)
25 Jul. 1969	06:06:42.4	25.6	63.3	579.	5.5	SH(sp,lp)
13 Dec. 1971	16:19:00.1	8.9	71.6	614.	5.3	P(sp)
12 Jan. 1972	09:59:10.3	6.9	71.8	580.	5.9	P(lp)
21 Jan. 1972	19:18:57.2	6.7	71.9	562.	5.6	P(lp)
3 Jan. 1973	02:58:16.7	27.7	63.3	563.	5.6	SH(sp,lp)
25 Oct. 1973	14:08:59.5	22.0	63.7	529.	6.1	P(lp)
11 Feb. 1975	21:56:49.5	20.7	62.9	562.	5.3	SH(sp)
9 Apr. 1977	04:04:12.5	10.0	71.2	564.	5.5	P(lp)

Table 2. Relative mean amplitudes

Station	spP			spSH			lpP			lpSH		
	Mean	S.E.M.	N	Mean	S.E.M.	N	Mean	S.E.M.	N	Mean	S.E.M.	N
BKS	1.13	0.18	4	-	-	0	1.12	0.03	6	1.11	0.00	1
COR	2.17	0.58	3	-	-	0	1.39	0.10	5	-	-	0
LON	0.60	0.06	5	-	-	0	0.98	0.07	4	-	-	0
GSC	1.00	0.16	2	0.51	0.03	7	0.88	0.11	4	1.03	0.10	4
MSO	-	-	0	0.31	0.00	1	-	-	0	-	-	0
DUG	0.98	0.01	3	0.24	0.02	5	1.24	0.06	6	1.04	0.07	3
TUC	0.38	0.02	5	0.21	0.02	8	0.76	0.09	6	0.60	0.05	6
BOZ	0.71	0.04	3	0.19	0.00	1	0.82	0.08	2	0.69	0.00	1
ALQ	0.37	0.03	5	0.60	0.04	7	0.89	0.06	5	1.09	0.09	4
GOL	0.48	0.10	3	0.26	0.02	7	1.04	0.06	5	1.02	0.07	4
RCD	1.76	0.49	2	1.09	0.00	1	1.13	0.00	1	1.00	0.00	1
LUB	1.64	0.50	3	1.99	0.16	4	0.97	0.08	4	1.30	0.18	2
JCT	-	-	0	1.39	0.19	4	1.06	0.02	4	1.30	0.18	4
DAL	2.22	0.41	3	6.07	0.00	1	1.52	0.10	3	1.56	0.18	2
FLO	1.79	0.00	1	0.94	0.16	4	1.42	0.00	1	0.47	0.03	4
OXF	0.90	0.14	3	1.76	0.27	7	0.62	0.04	5	0.52	0.11	3
SHA	1.53	0.00	1	1.60	0.74	2	1.03	0.02	6	1.26	0.06	4
AAM	0.62	0.09	2	1.13	0.34	2	1.00	0.11	3	0.97	0.08	4
ATL	0.72	0.11	4	0.31	0.02	5	0.84	0.05	5	0.85	0.12	4
BLA	0.90	0.09	3	0.91	0.14	4	1.06	0.08	6	1.35	0.18	4
SCP	0.46	0.06	4	0.75	0.06	6	1.03	0.06	6	1.44	0.07	5
GEO	0.78	0.04	2	0.47	0.13	5	0.98	0.14	3	1.09	0.15	3
OGD	0.34	0.04	2	0.32	0.05	5	0.72	0.09	4	1.07	0.09	5
WES	0.53	0.00	1	0.34	0.09	3	0.79	0.06	5	0.56	0.07	2
BEC	-	-	0	0.53	0.11	2	0.73	0.01	4	0.69	0.15	4

Table 3. Model results for Δt^* , τ_m .

Station	Deep spP		Shallow spP		$\Delta t_{\beta}^{*1.5H}$, sec
	Δt_{α}^* , sec	τ_m , sec ($t_{\alpha}^*=0.6$)	Δt_{α}^* , sec	τ_m , sec ($t_{\alpha}^*=1.3$)	
AAM	-0.11	0.034	-0.05	0.081	1.46
ALQ	0.11	0.009	0.06	0.061	1.01
ATL	-0.16	0.068	-0.12	0.100	2.03
BEC	-	-	-	-	2.92
BKS	-0.31	0.110	-0.21	0.115	0.94
BLA	-0.23	0.088	-0.22	0.119	0.24
BOZ	-0.14	0.064	0.41	0.001*	2.98
COR	-0.49	0.280	-0.46	0.180	-
DAL	-0.50	0.265	-0.50	0.240	-0.23
DUG	-0.26	0.096	-0.20	0.112	1.19
FLO	-0.46	0.200	-0.50	0.212	4.80
GEO	-0.18	0.072	-0.12	0.100	1.00
GOL	-0.01	0.032	0.15	0.047	1.25
GSC	-0.25	0.100	-0.15	0.107	1.24
JCT	-	-	-0.28	0.131	0.38
LON	-0.10	0.049	0.21	0.035	-
LUB	-0.41	0.180	-0.53	0.021	0.38
MSO	-	-	-0.08	0.090	-
OGD	0.13	0.001*	0.15	0.047	1.07
OXF	-0.28	0.089	-0.52	0.200	4.30
RCD	-0.44	0.190	-	-	1.34
SCP	0.00*	0.029	0.00*	0.072	0.00*
SHA	-0.39	0.165	-	-	0.47
TUC	0.09	0.012	0.05	0.064	3.54
WES	-0.05	0.040	0.05	0.069	3.90

* Reference station

Table 4.

Absolute t_{α}^* determinations

STA	t_{α}^*	R	C_{sp}	C_{lp}	γ
BLA	1.00	0.060	0.972	0.997	0.984
GEO	0.72	0.115	0.986	0.993	0.988
SCP	0.79	0.101	0.975	0.997	0.984
ALQ	0.96	0.047	0.999	0.986	0.992
DUG	1.03	0.074	0.935	0.962	0.949
PRE	0.38	0.648	0.989	0.907	0.948
SDB	0.39	0.568	0.905	0.949	0.929

 τ_m determinations with $t_{\alpha}^* = 0.8$

STA	τ_m	R	C_{sp}	C_{lp}	γ
BLA	0.05	0.140 0.073 _{sp} 0.073 _{lp}	0.986	0.998	0.560
GEO	0.04	0.108	0.985	0.994	0.989
SCP	0.01	0.100	0.976	0.996	0.984
PRE	0.15	0.446	0.959	0.956	0.958
SDB	0.17	0.402	0.872	0.983	0.930

Table 5a. Δt_{β}^* from long period SH spectral ratios in the band 6.3-25.0 sec.[†]

Station	Dec. 20, 1966	Jan. 17, 1967	Sep. 9, 1967	Jul. 25, 1969	Jan. 3, 1973 [*]
AAM	-	1.43 ± 0.02	1.23 ± 0.03	0.02 ± 0.02	1.69 ± 0.01
ALQ	0.38 ± 0.03	1.30 ± 0.04	-	-	0.40 ± 0.06
ATL	-	6.32 ± 0.07	-	3.44 ± 0.09	4.85 ± 0.03
BEC	2.85 ± 0.08	1.75 ± 0.09	-	1.20 ± 0.06	2.52 ± 0.08
BLA	1.06 ± 0.02	1.15 ± 0.02	-	1.21 ± 0.02	1.14
DAL	-	-	-	-	-0.51 ± 0.04
FLO	0.66 ± 0.06	0.94 ± 0.04	0.39 ± 0.04	1.34 ± 0.06	-
GEO	-	2.33 ± 0.01	2.20 ± 0.04	4.40 ± 0.03	-
GOL	4.74 ± 0.09	4.25 ± 0.08	3.97 ± 0.09	4.52 ± 0.08	-
JCT	3.15 ± 0.04	1.41 ± 0.02	-	2.25 ± 0.02	1.49 ± 0.03
LUB	-	-0.28 ± 0.02	-	0.65 ± 0.03	-
OGD	1.03 ± 0.04	1.95 ± 0.01	-	3.16 ± 0.04	3.48 ± 0.04
ONF	-	5.87 ± 0.05	4.55 ± 0.03	3.19 ± 0.08	-
SHA	-	-1.37 ± 0.06	-1.01 ± 0.09	-2.79 ± 0.04	-1.74 ± 0.05
TUC	1.99 ± 0.05	1.74 ± 0.07	3.11 ± 0.02	3.92 ± 0.06	3.70 ± 0.05
WES	-	-	5.01 ± 0.05	5.02 ± 0.05	-

[†] Reference station: SCP

^{*} BLA reference corrected to SCP

[‡] Strong second arrival in signal

Table 5b. Δt_b^* from long period SH spectral ratios in the band 6.7-14.3 sec.[†]

Station	Dec. 20, 1966	Jan. 17, 1967	Sep. 9, 1967	Jul. 25, 1969	Jan. 3, 1973 [*]
AAM	-	1.08 ± 0.01	0.49 ± 0.02	-0.80 ± 0.02	1.15 ± 0.01
ALQ	-0.20 ± 0.02	1.15 ± 0.04	-	-	-1.03 ± 0.05
ATL	-	6.65 ± 0.03	-	3.60 ± 0.08	4.24 ± 0.02
BEC [‡]	3.08 ± 0.08	2.19 ± 0.08	-	1.73 ± 0.04	3.49 ± 0.07
BLA	0.69 ± 0.01	1.15 ± 0.01	-	0.93 ± 0.01	0.89
DAL	-	-	-	-	-1.17 ± 0.02
FLO	-0.96 ± 0.04	0.75 ± 0.02	0.15 ± 0.03	0.08 ± 0.04	-
GEO	-	2.30 ± 0.01	2.30 ± 0.03	4.32 ± 0.01	-
GOL	5.71 ± 0.05	5.25 ± 0.03	5.97 ± 0.05	4.56 ± 0.01	-
JCT	3.89 ± 0.02	2.07 ± 0.02	-	2.41 ± 0.01	1.08 ± 0.02
LUB	-	0.47 ± 0.01	-	1.40 ± 0.01	-
OGD	0.28 ± 0.02	1.85 ± 0.01	-	3.15 ± 0.03	3.11 ± 0.01
ONF	-	6.52 ± 0.01	5.70 ± 0.01	2.35 ± 0.04	-
SHA [‡]	-	-1.16 ± 0.06	0.30 ± 0.08	-2.73 ± 0.04	-1.51 ± 0.04
TUC [‡]	0.67 ± 0.03	1.11 ± 0.03	2.63 ± 0.02	3.85 ± 0.04	3.61 ± 0.02
WES	-	-	5.07 ± 0.01	4.31 ± 0.04	-

[†] Reference station: SCP

^{*} BLA reference corrected to SCP

[‡] Strong second arrival in signal

Table 6. Average Δt_{β}^* from spectral ratios⁺.

Station	N	Table 5a	Table 5b	Solomon & Toksöz (1970)
AAM	4	1.1 ± 0.7	0.5 ± 0.9	0.99
ALQ	3	0.7 ± 0.5	0.0 ± 1.1	0.06
ATL	3	4.9 ± 1.4	4.8 ± 1.6	-1.08
BEC ^f	4	2.1 ± 0.7	2.6 ± 0.8	-
BLA	3	1.1 ± 0.1	0.9 ± 0.2	-
DAL	1	-0.5	-1.2	-
FLO	4	0.8 ± 0.4	0.0 ± 0.7	-3.72
GEO	3	3.0 ± 1.2	3.0 ± 1.2	2.26
GOL	4	4.4 ± 0.3	5.4 ± 0.6	-
JCT	4	2.1 ± 0.8	2.4 ± 1.2	1.43
LUB	2	0.2 ± 0.7	0.9 ± 0.7	-0.95
OGD	4	2.4 ± 1.1	2.1 ± 1.4	-
OXF	3	4.5 ± 1.3	4.9 ± 2.2	-1.43
SHA ^f	4	-1.7 ± 0.8	-1.3 ± 1.2	-2.42
TUC ^f	5	2.9 ± 1.0	2.4 ± 1.4	0.83
WES	2	5.0 ± 0.1	4.7 ± 0.5	3.02

⁺ reference station SCP^f strong second arrival

DEEP SOUTH AMERICAN EVENTS

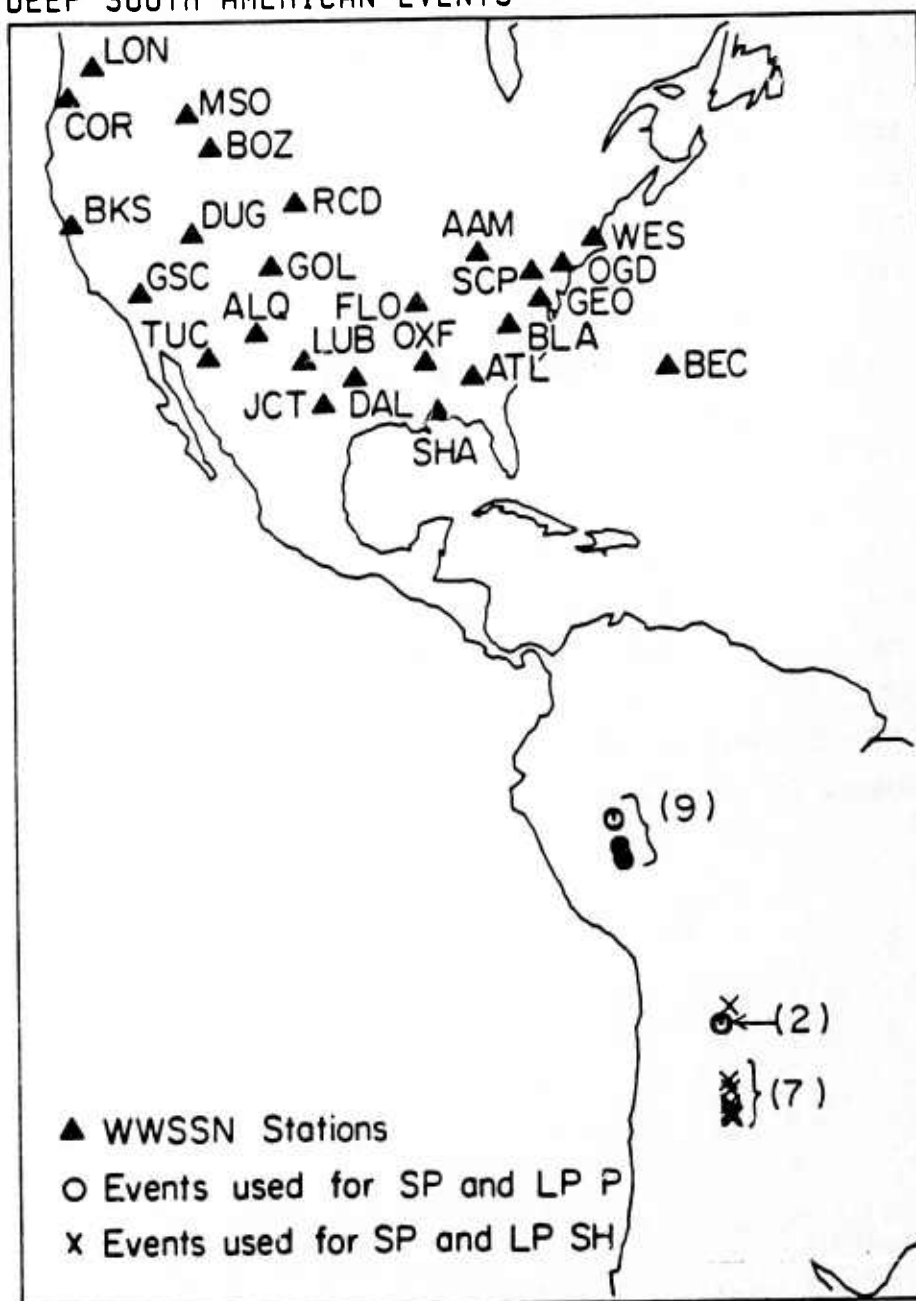


Fig. 1

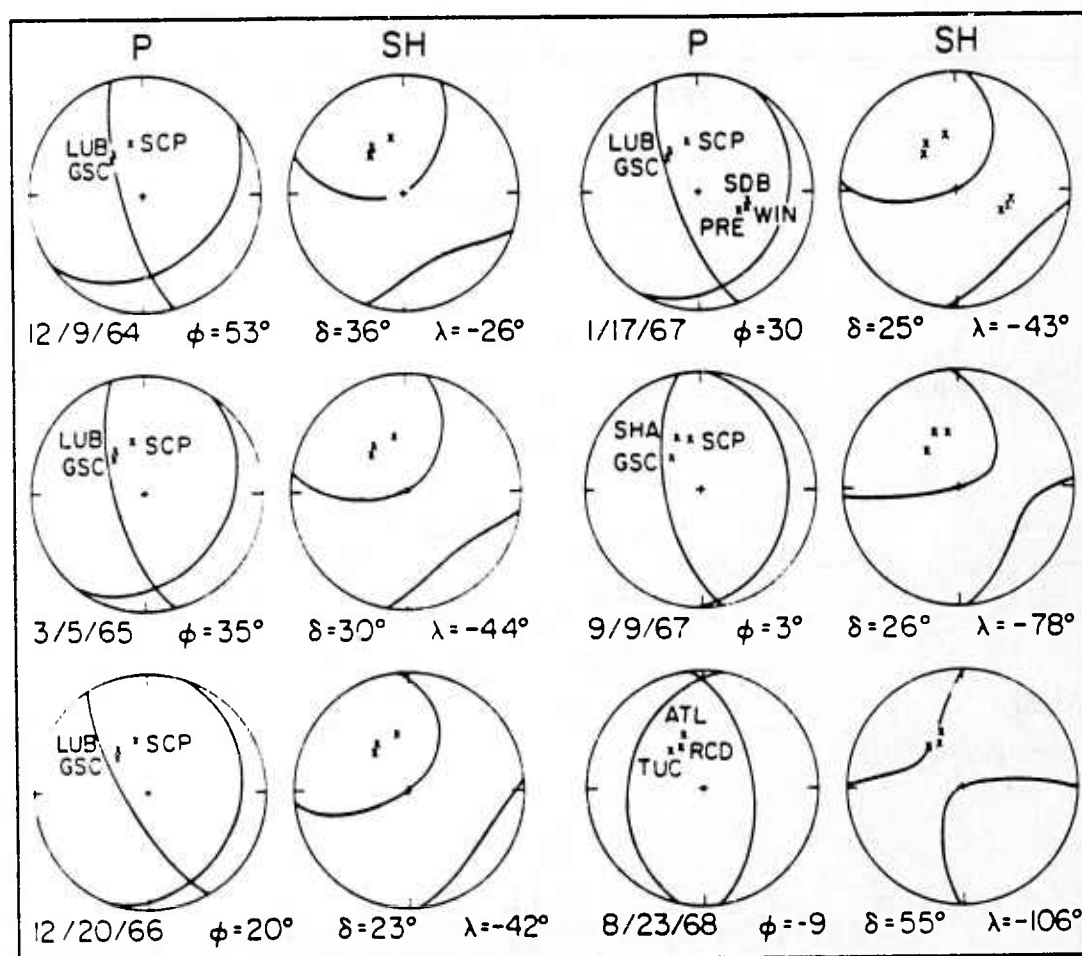
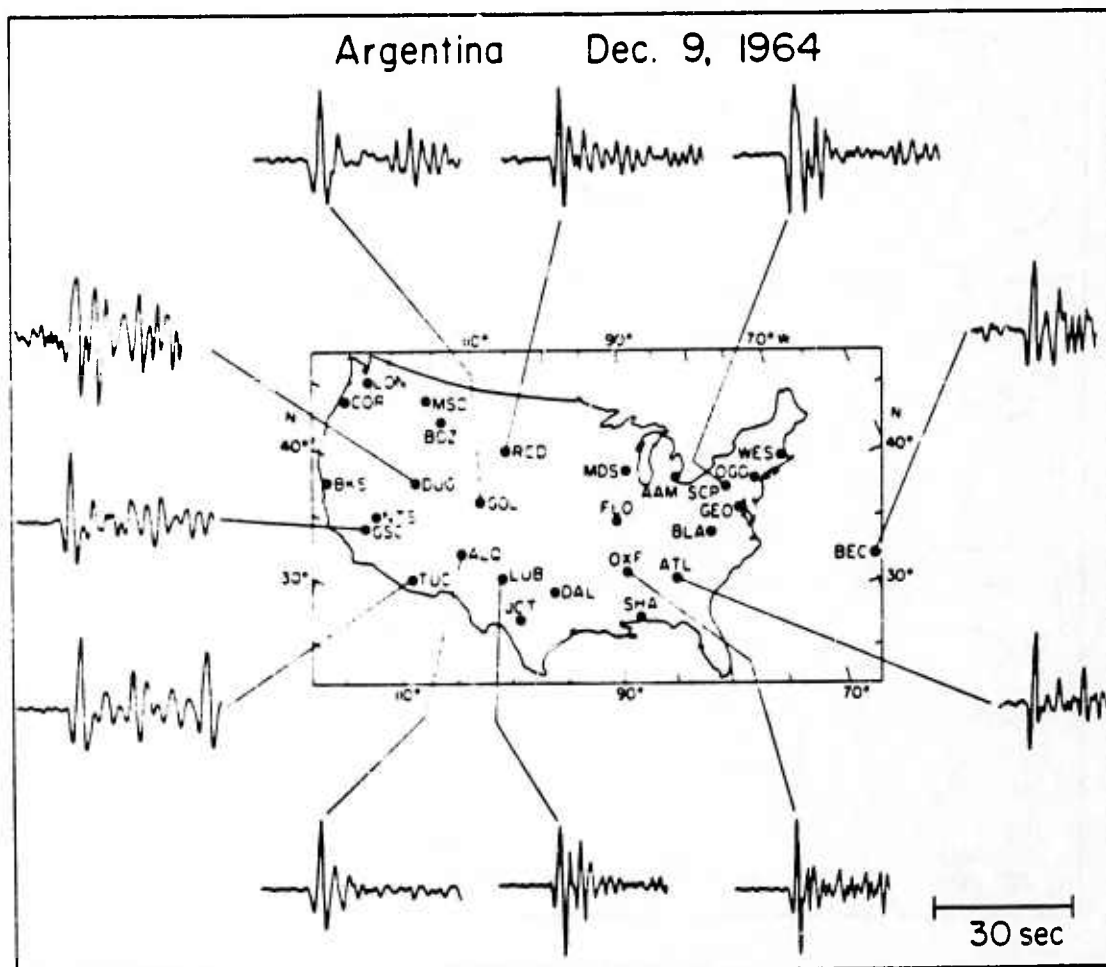


Fig. 2

*Fig. 3*

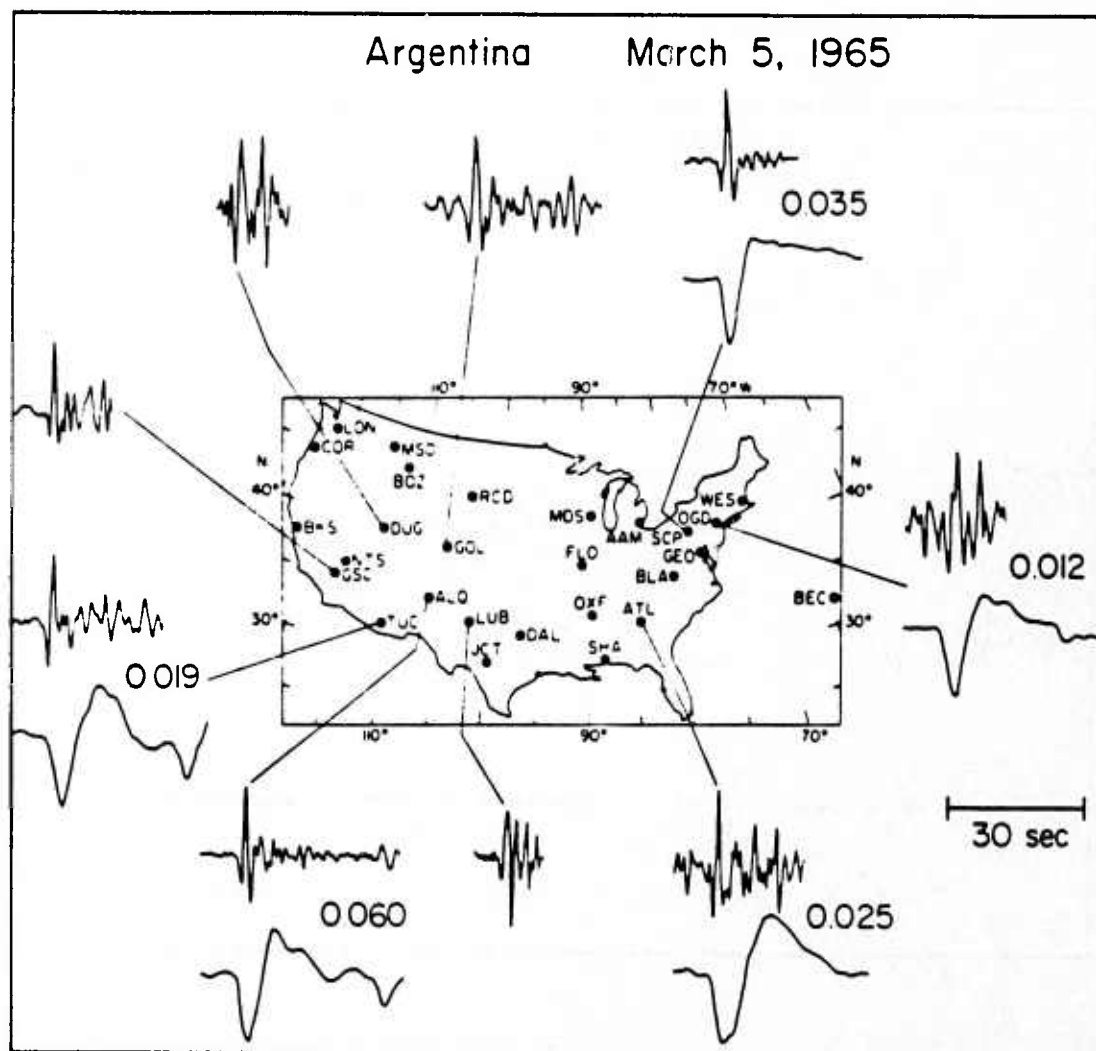
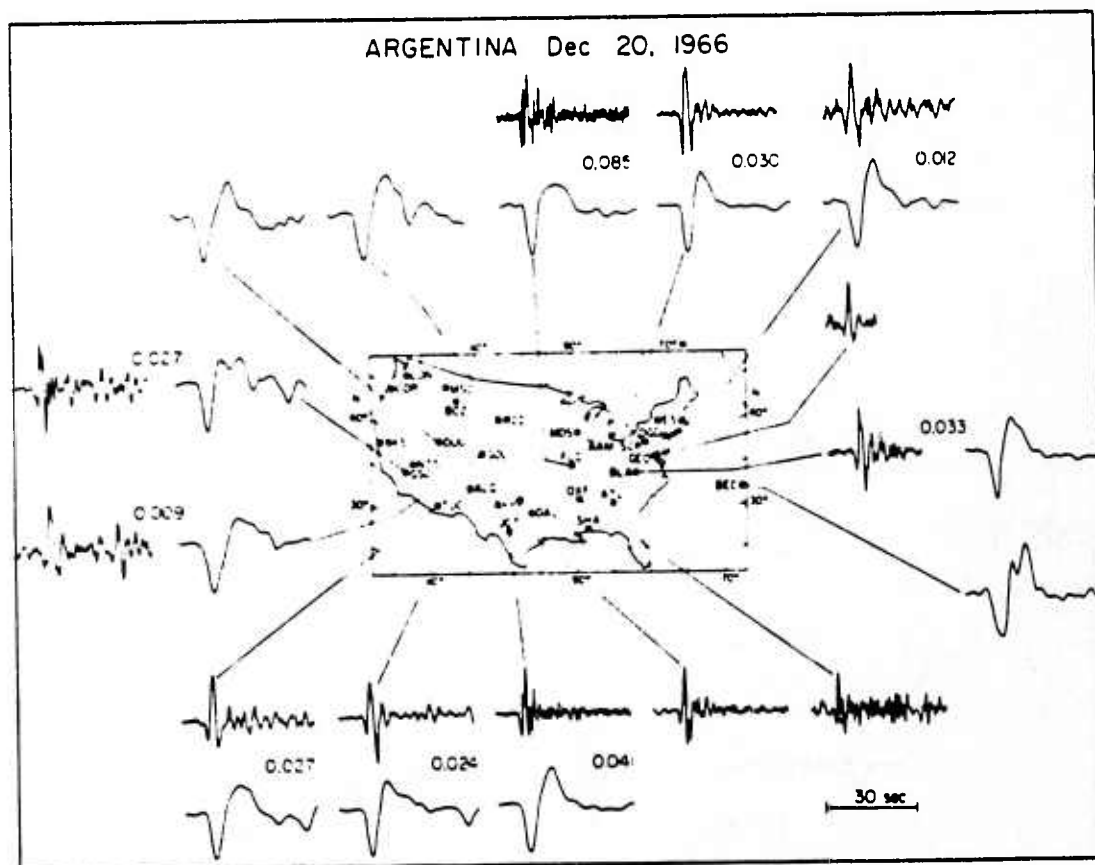


Fig. 4

*Fig. 5*

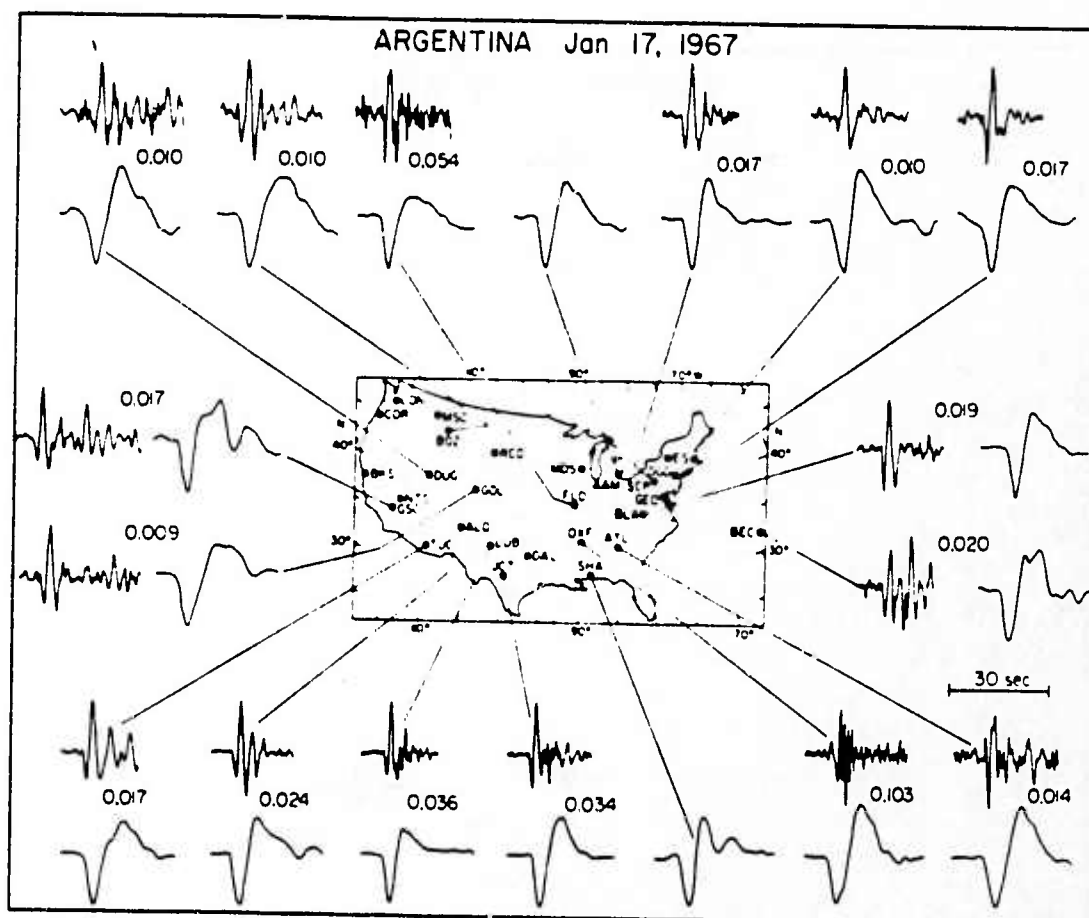
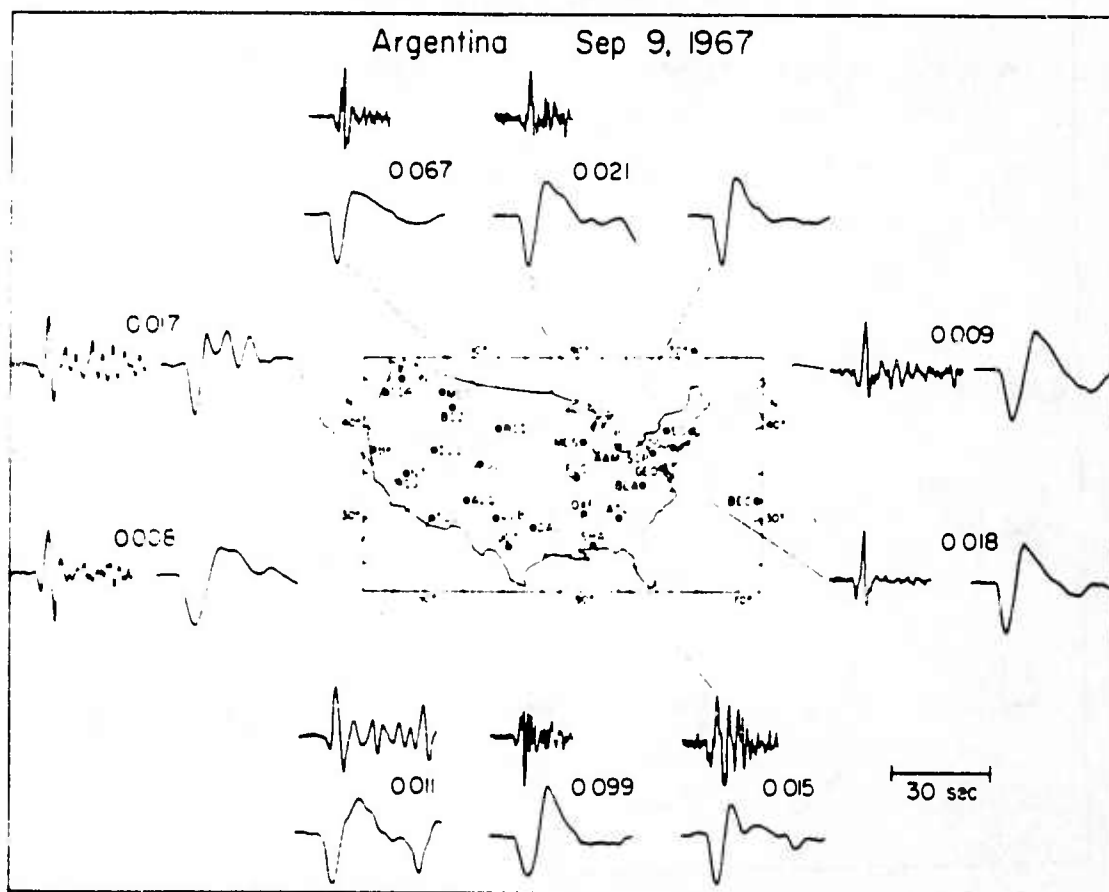
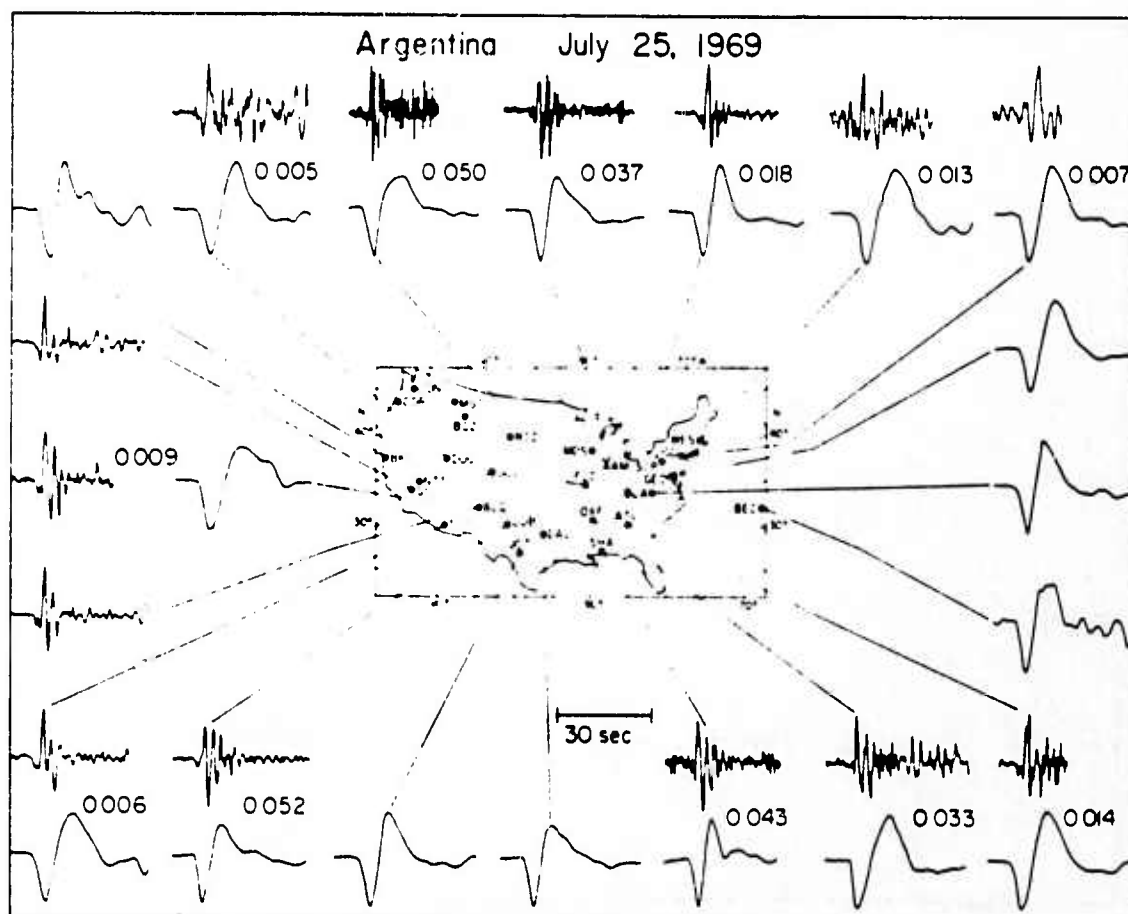
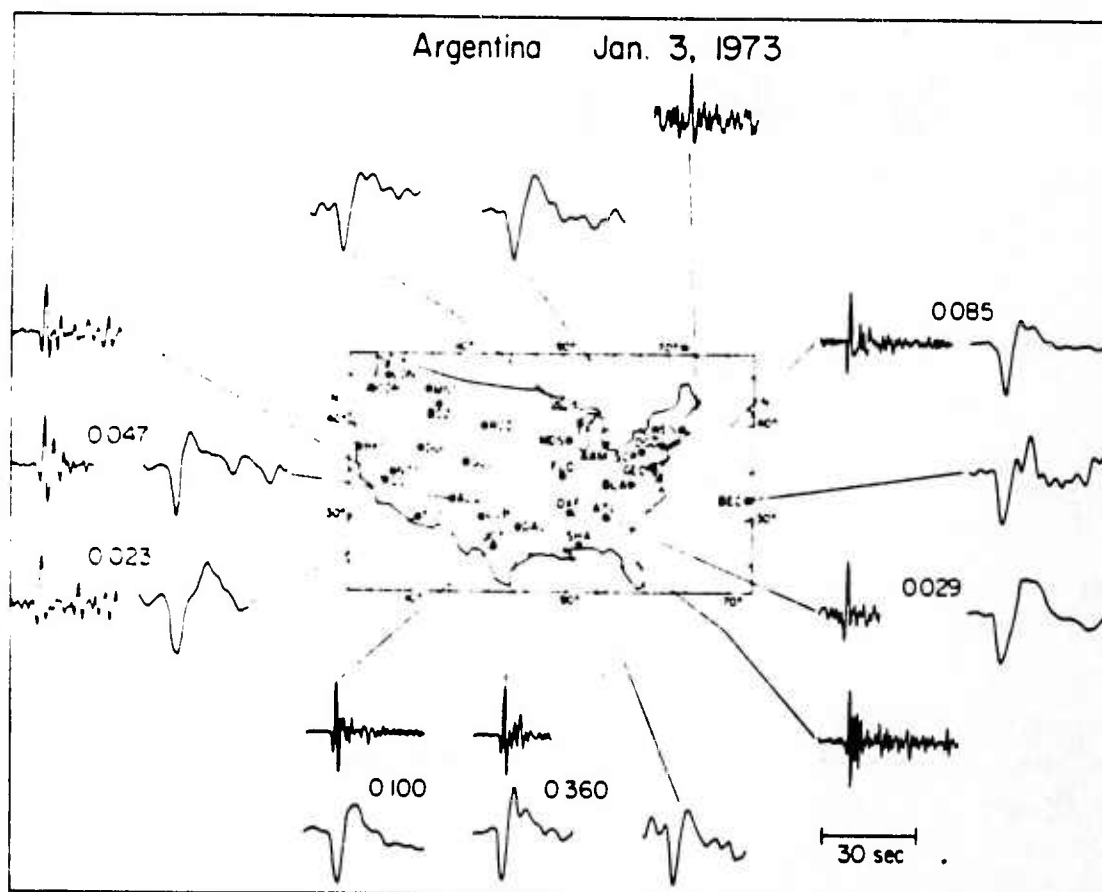
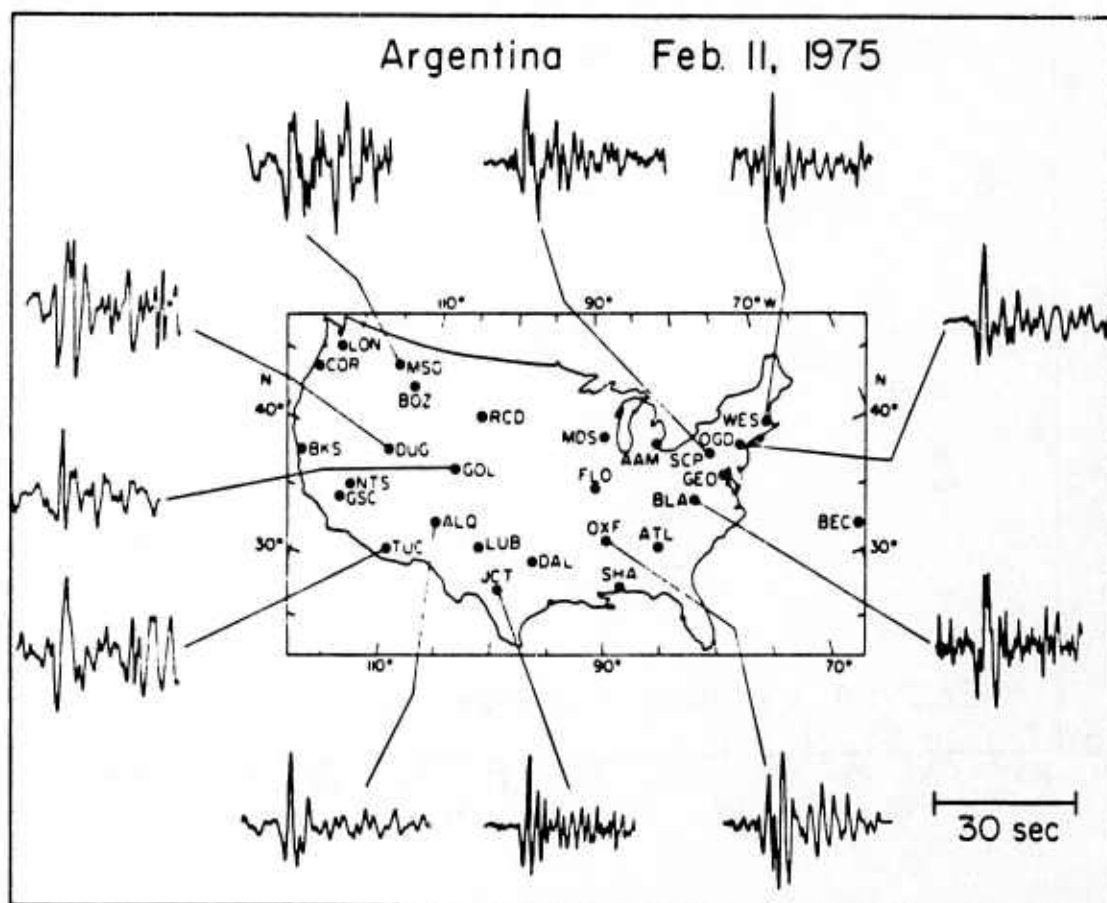


Fig. 6

*Fig. 7*

*Fig. 8*

*Fig. 9*

*Fig. 10*

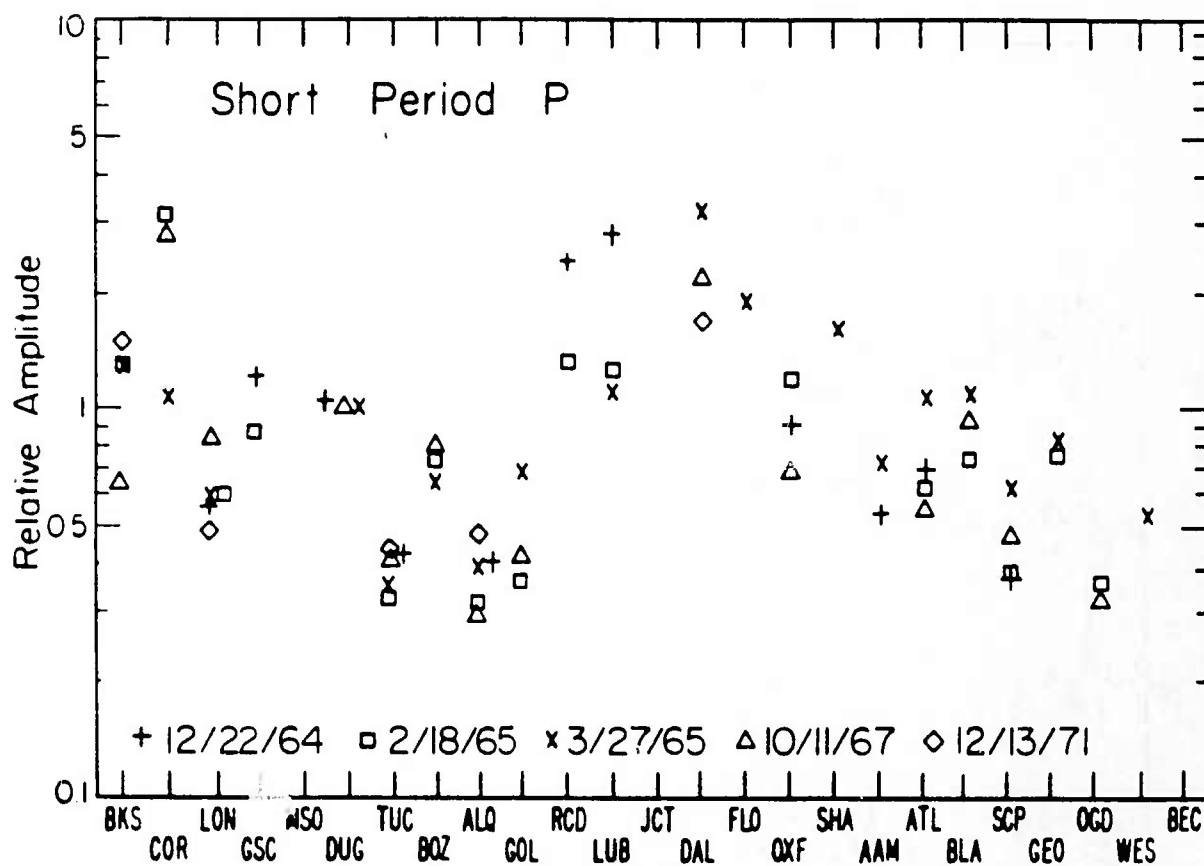
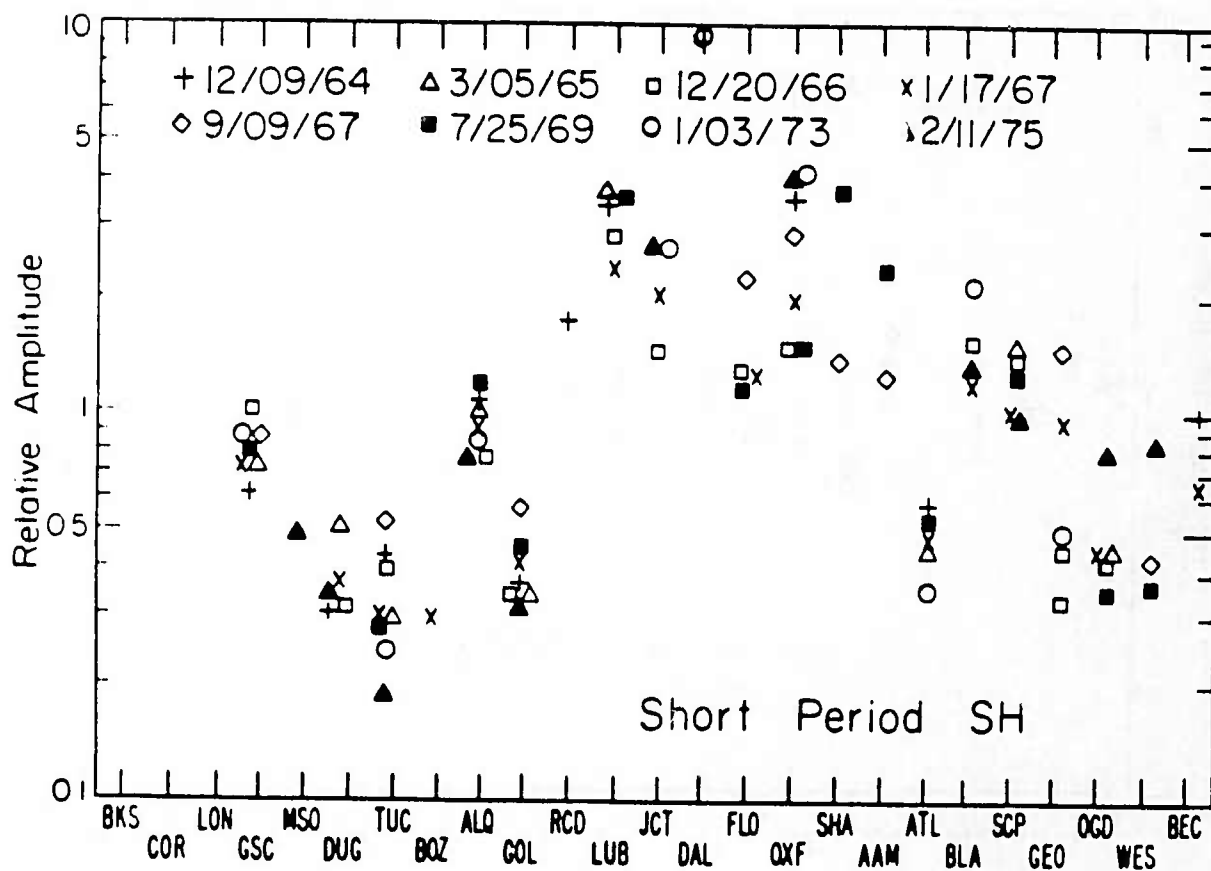


Fig. 11



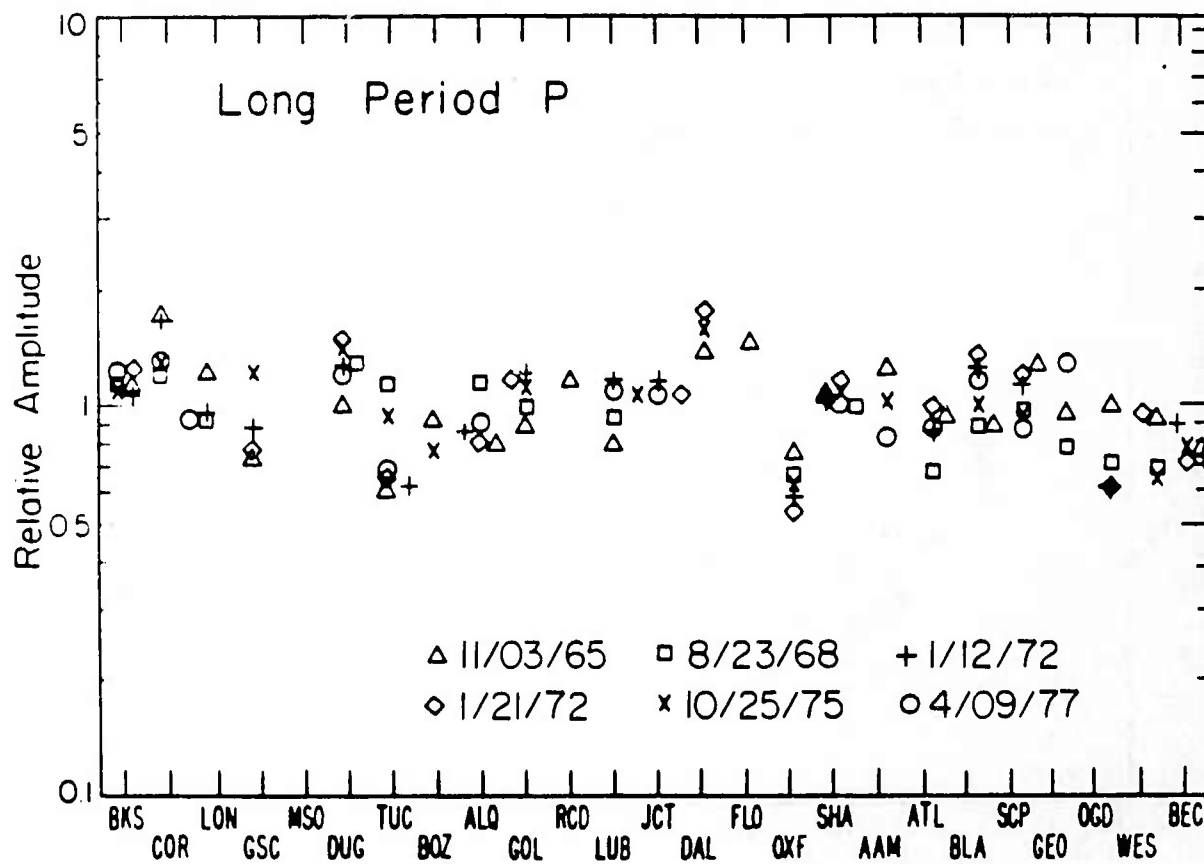


Fig. 13

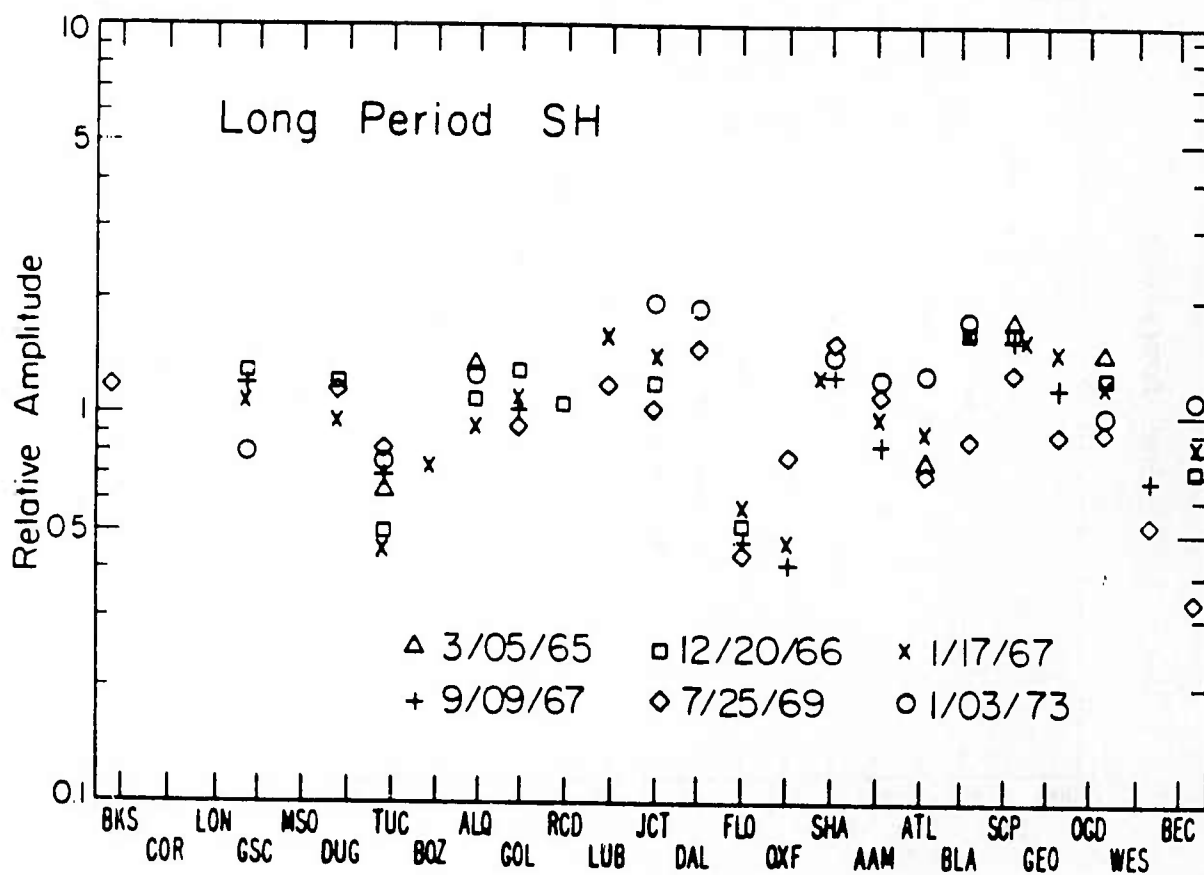


Fig. 14

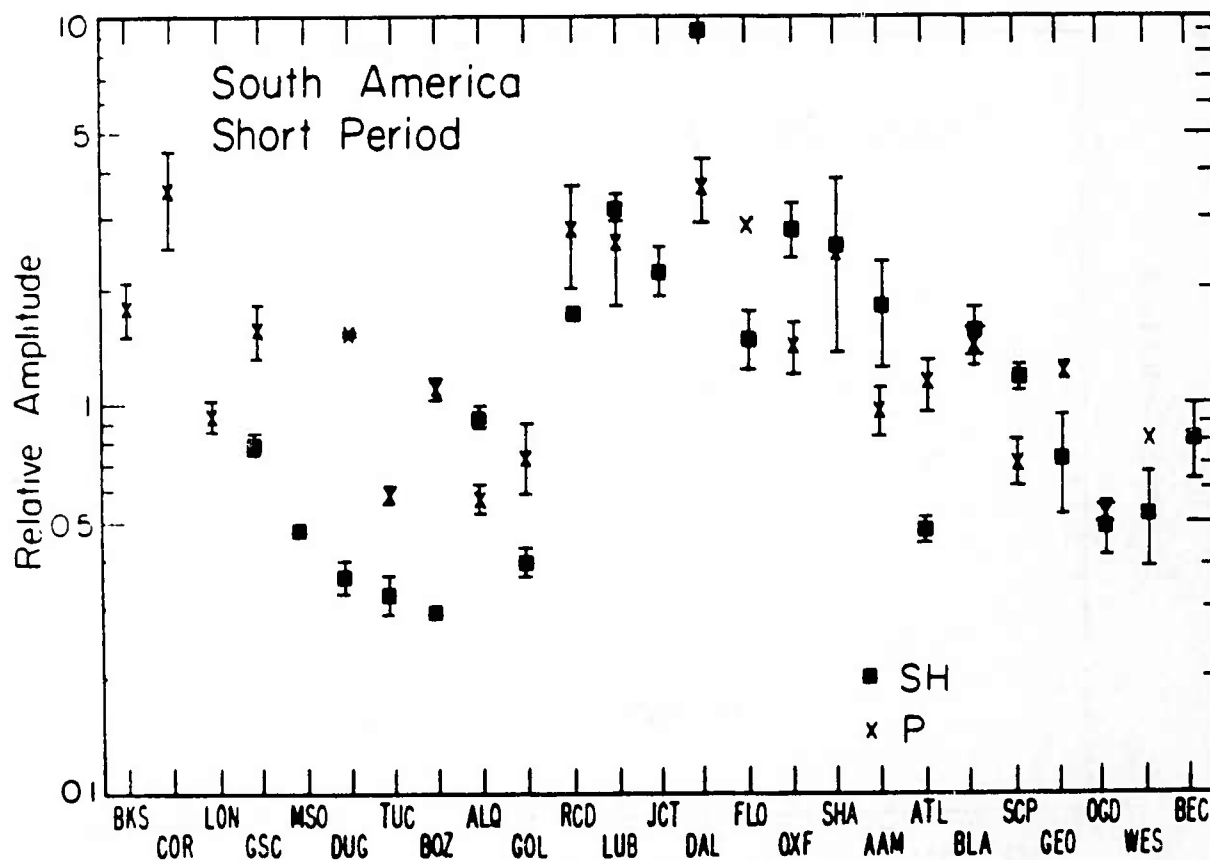


Fig. 15

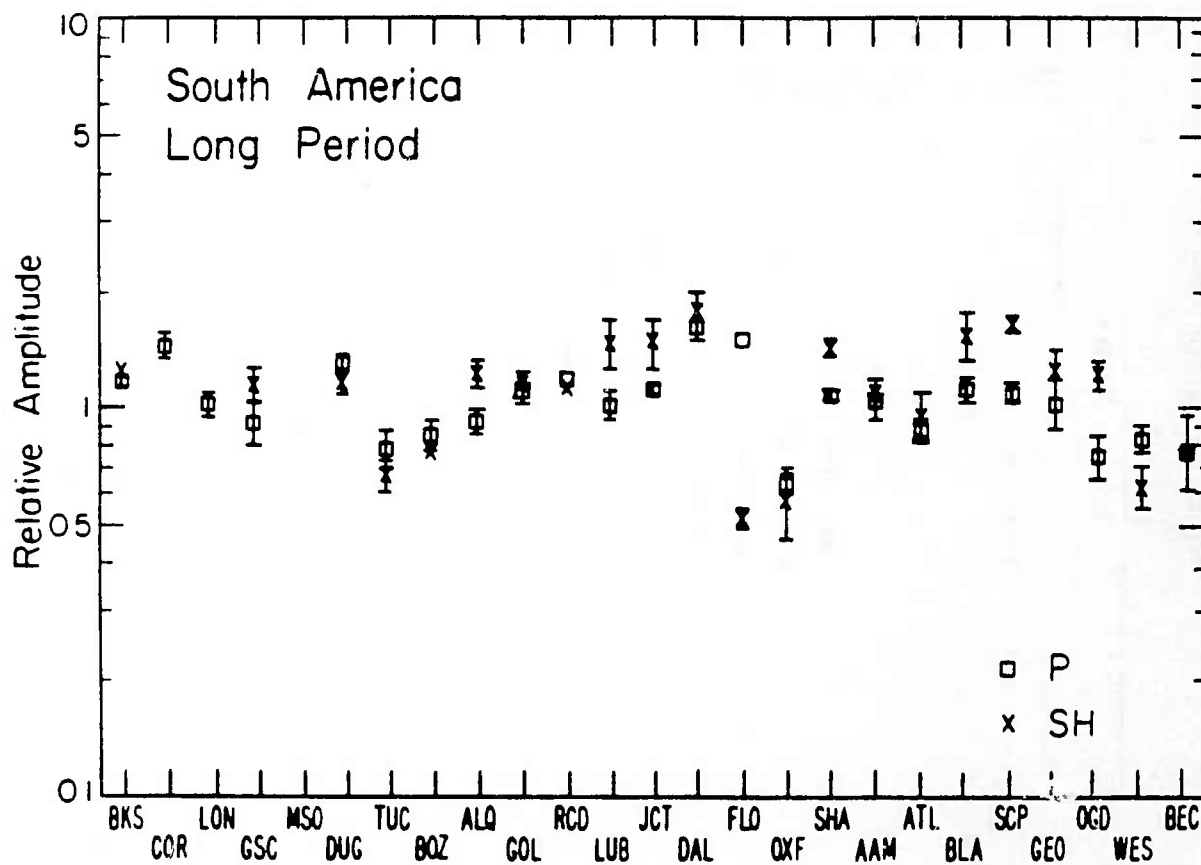


Fig. 16

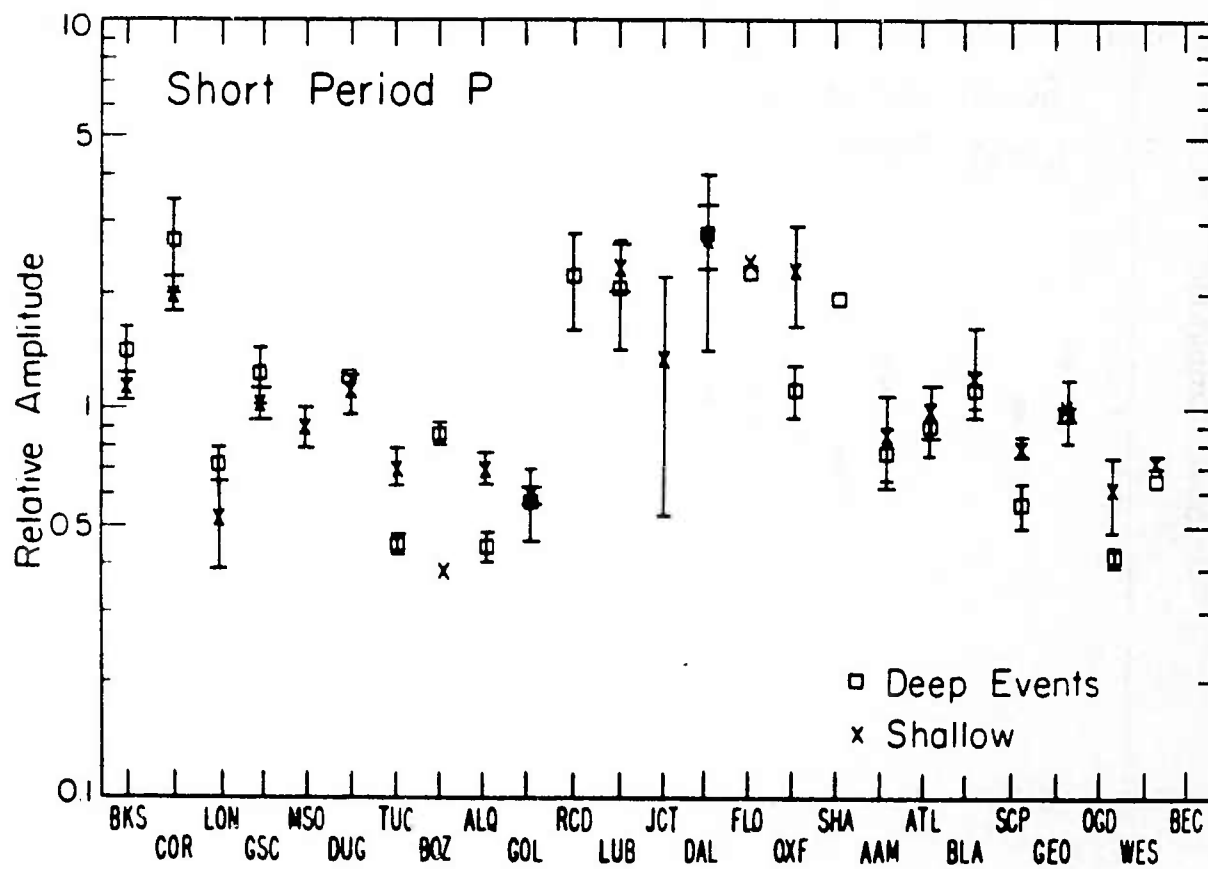


Fig. 17

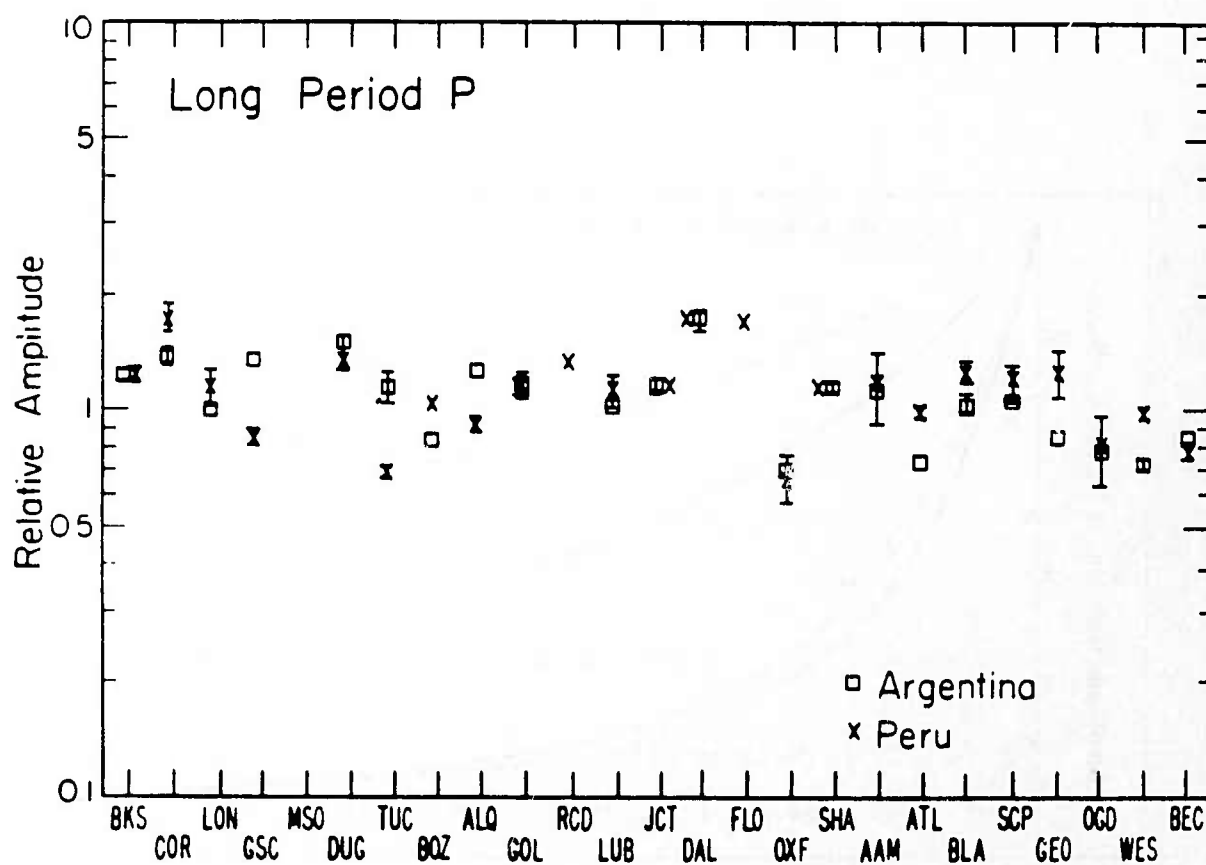


Fig. 18

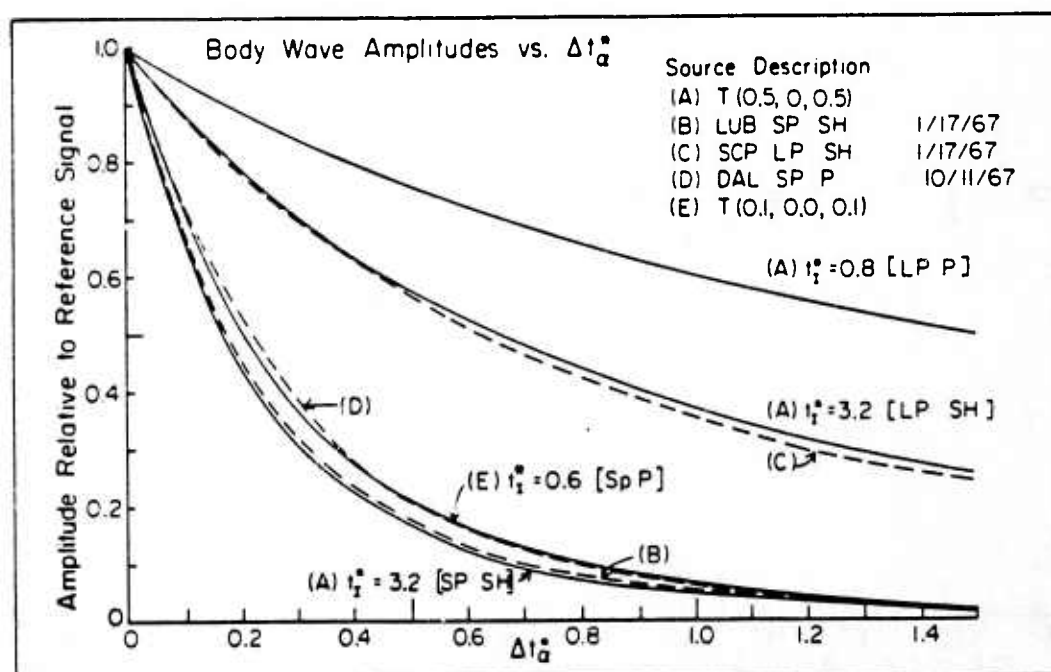


Fig. 19

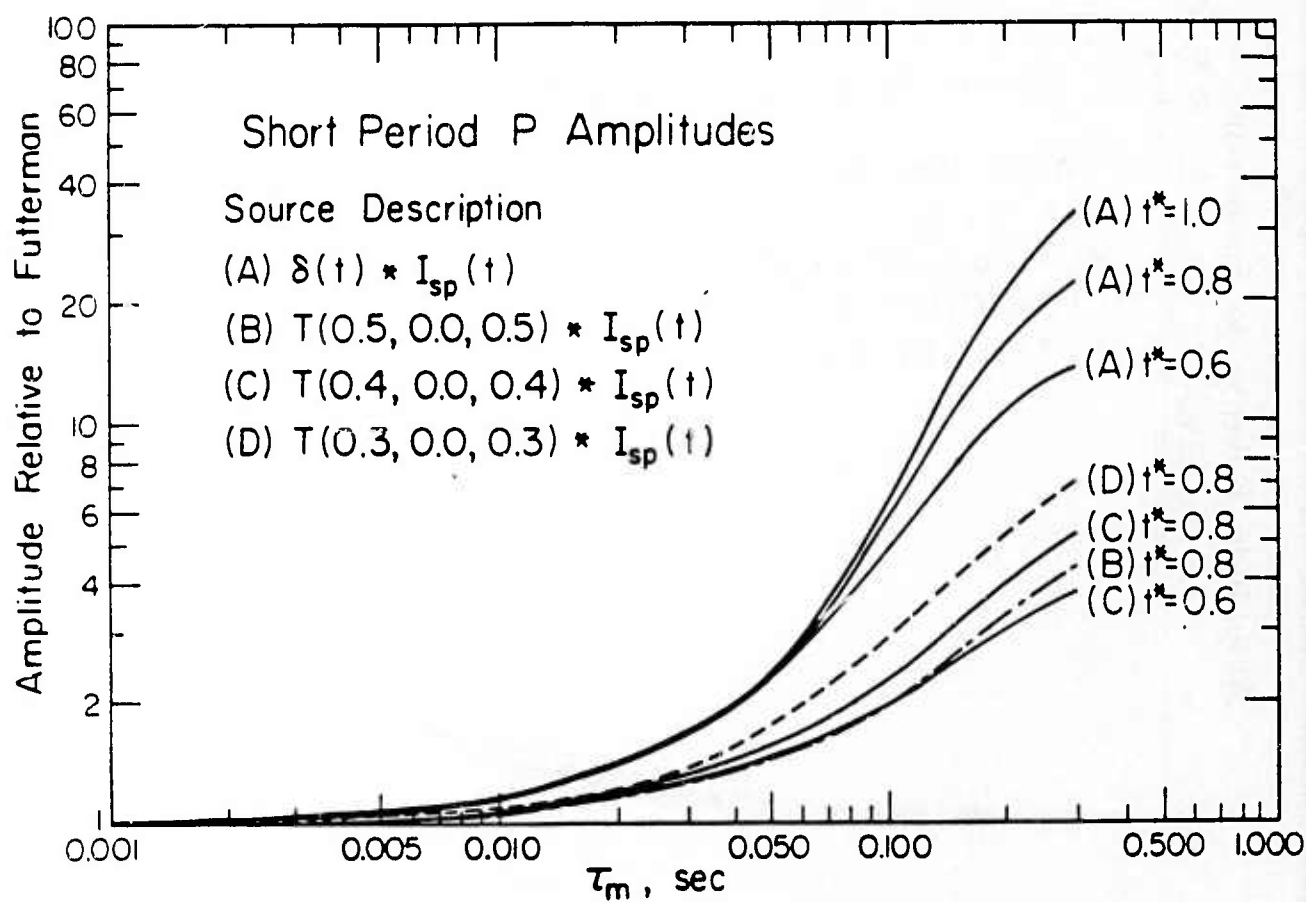


Fig. 20

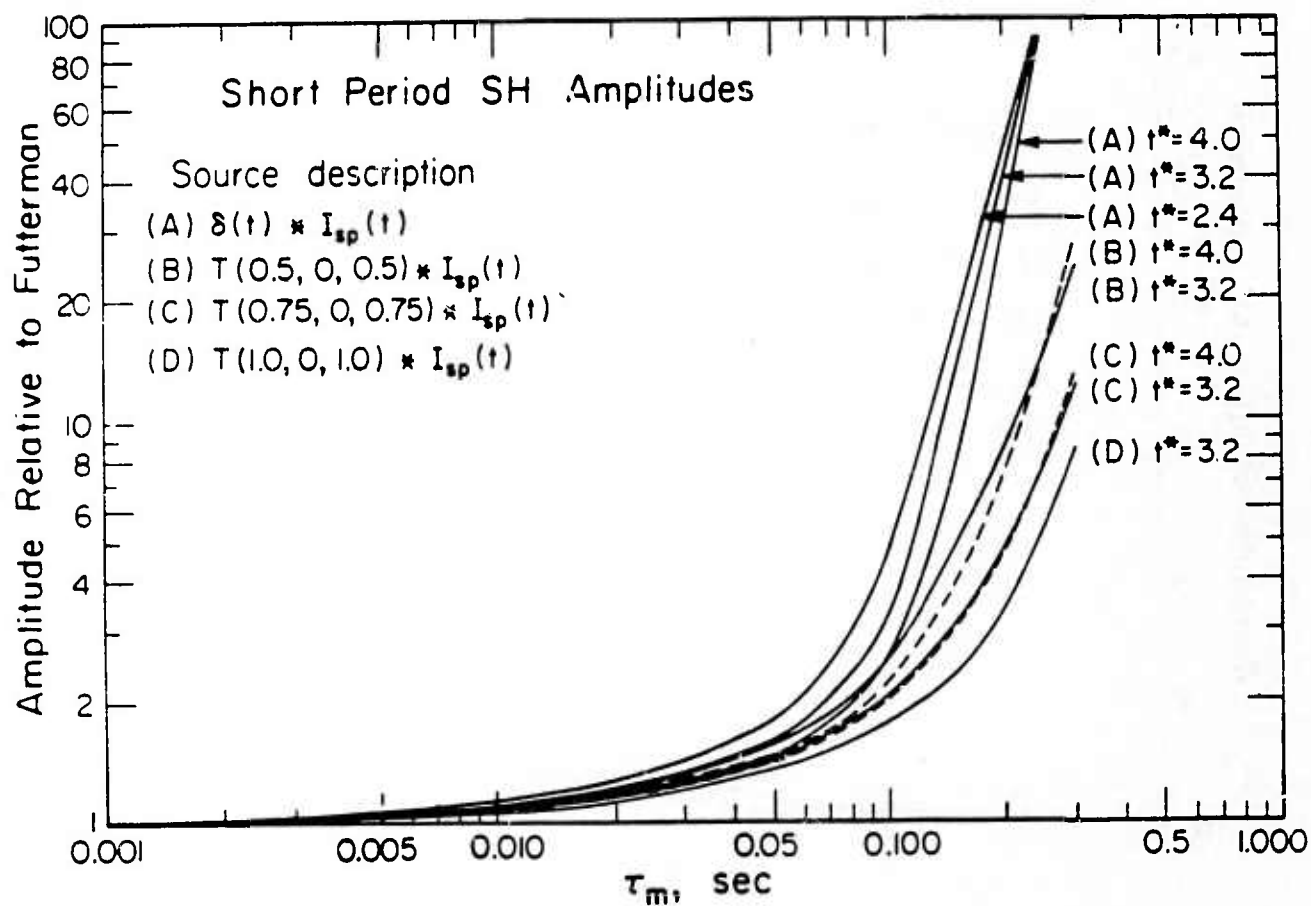


Fig. 21

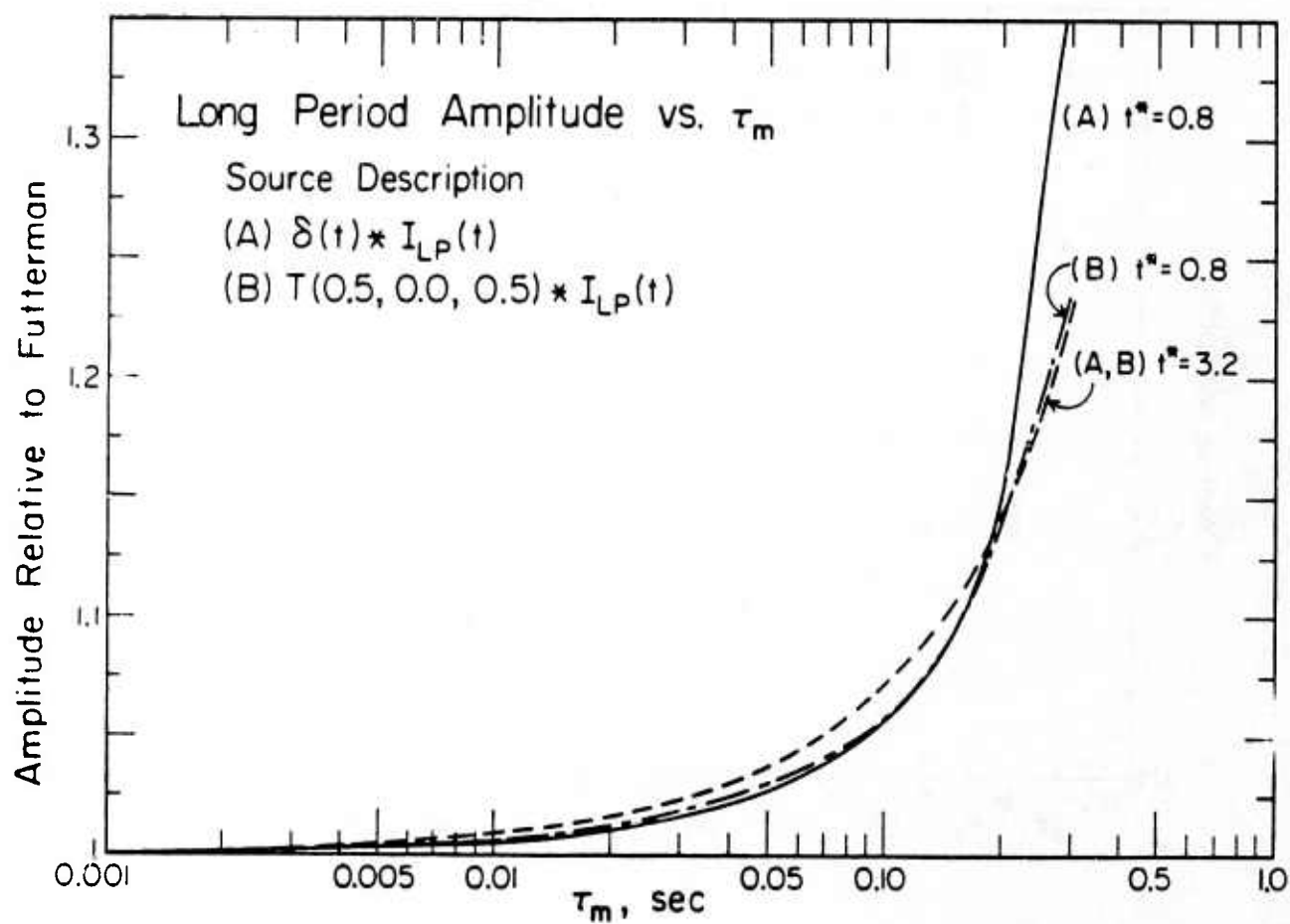


Fig. 22

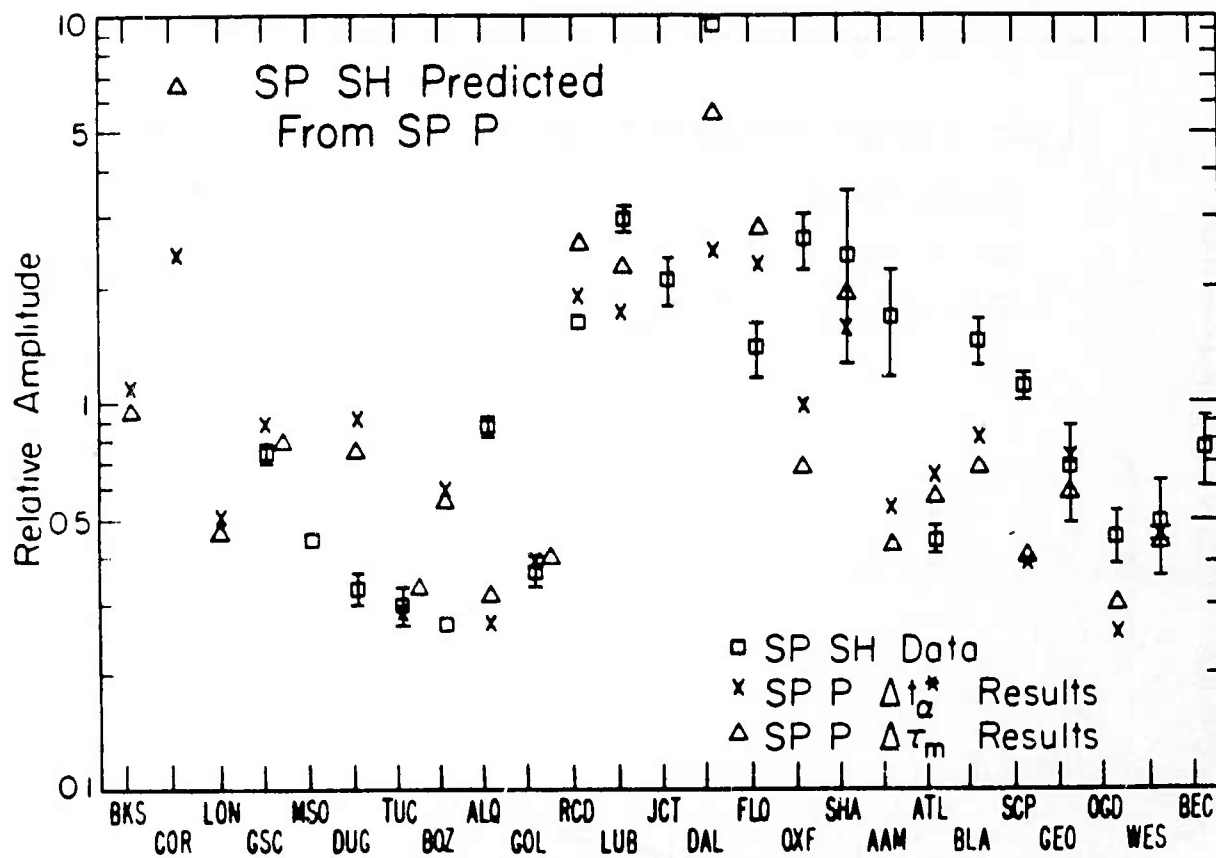


Fig. 23

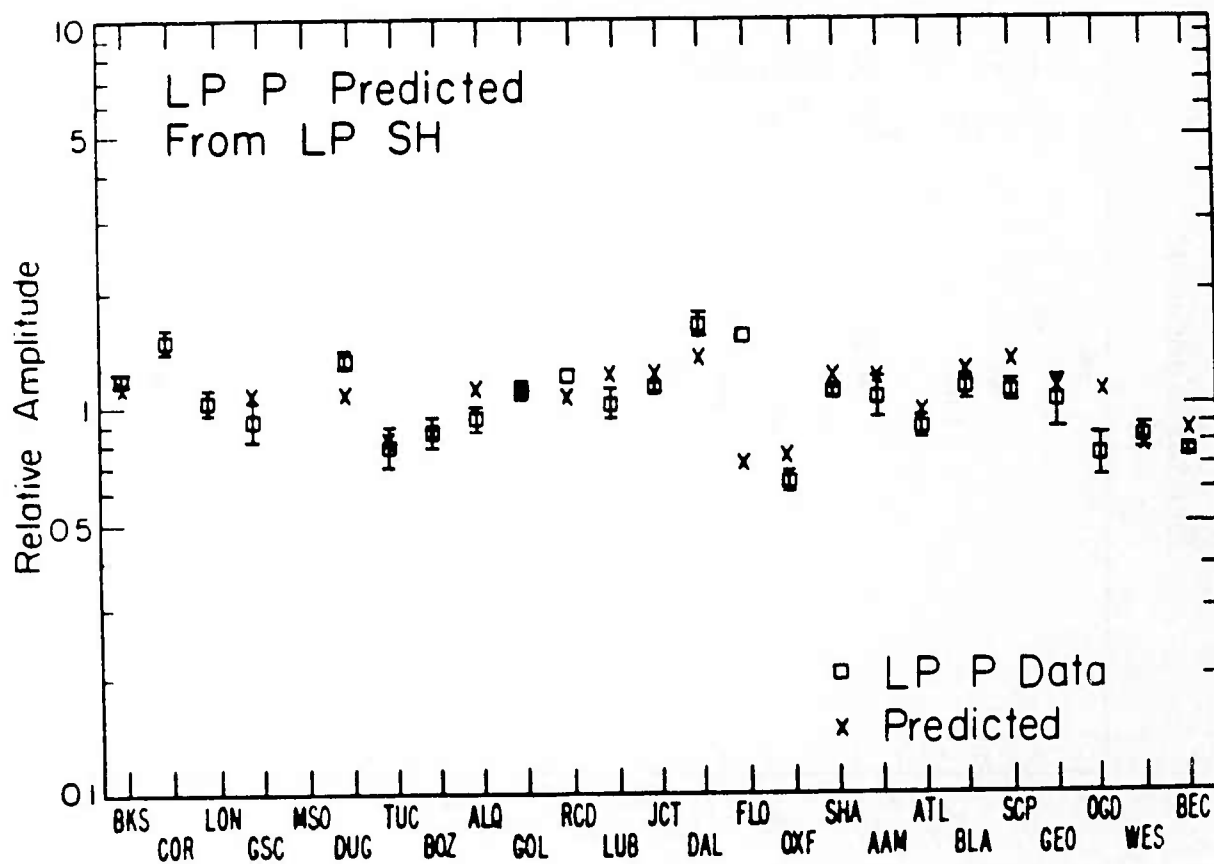


Fig. 24

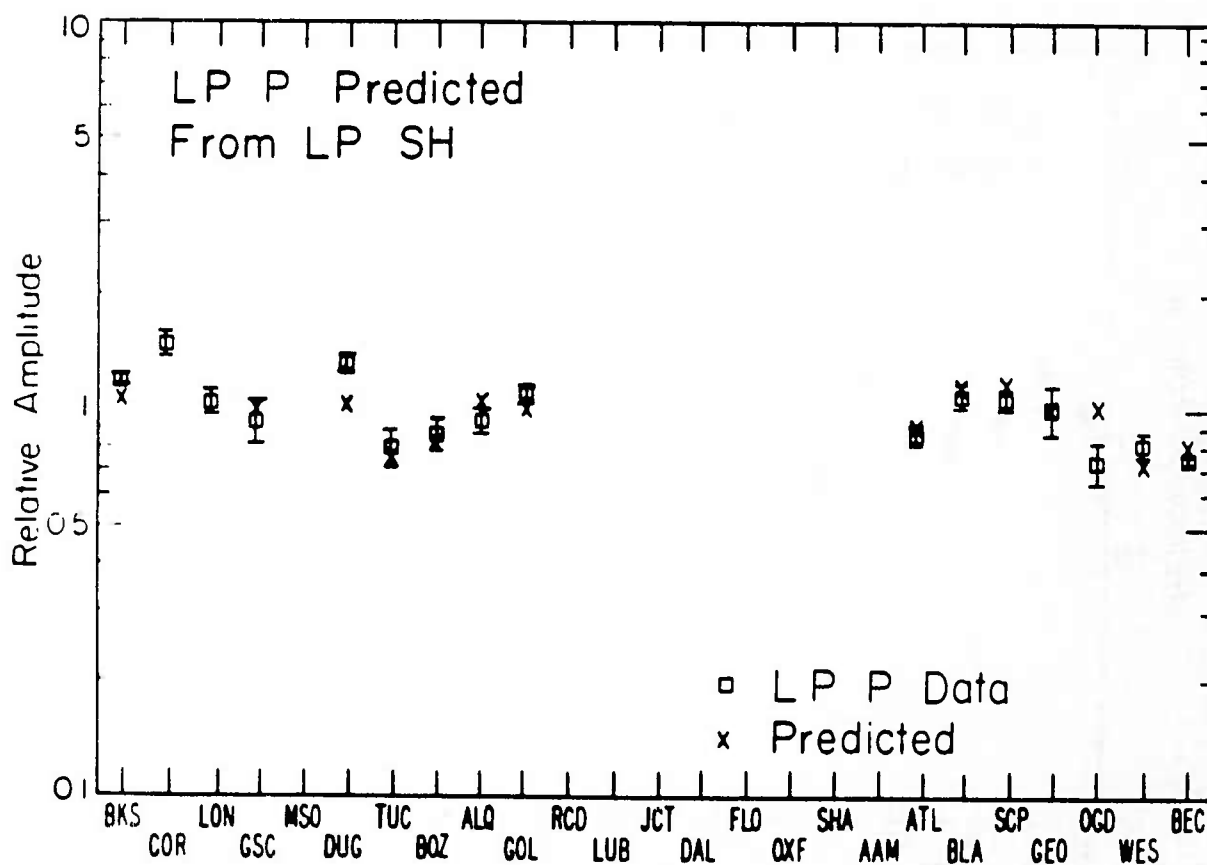


Fig. 25

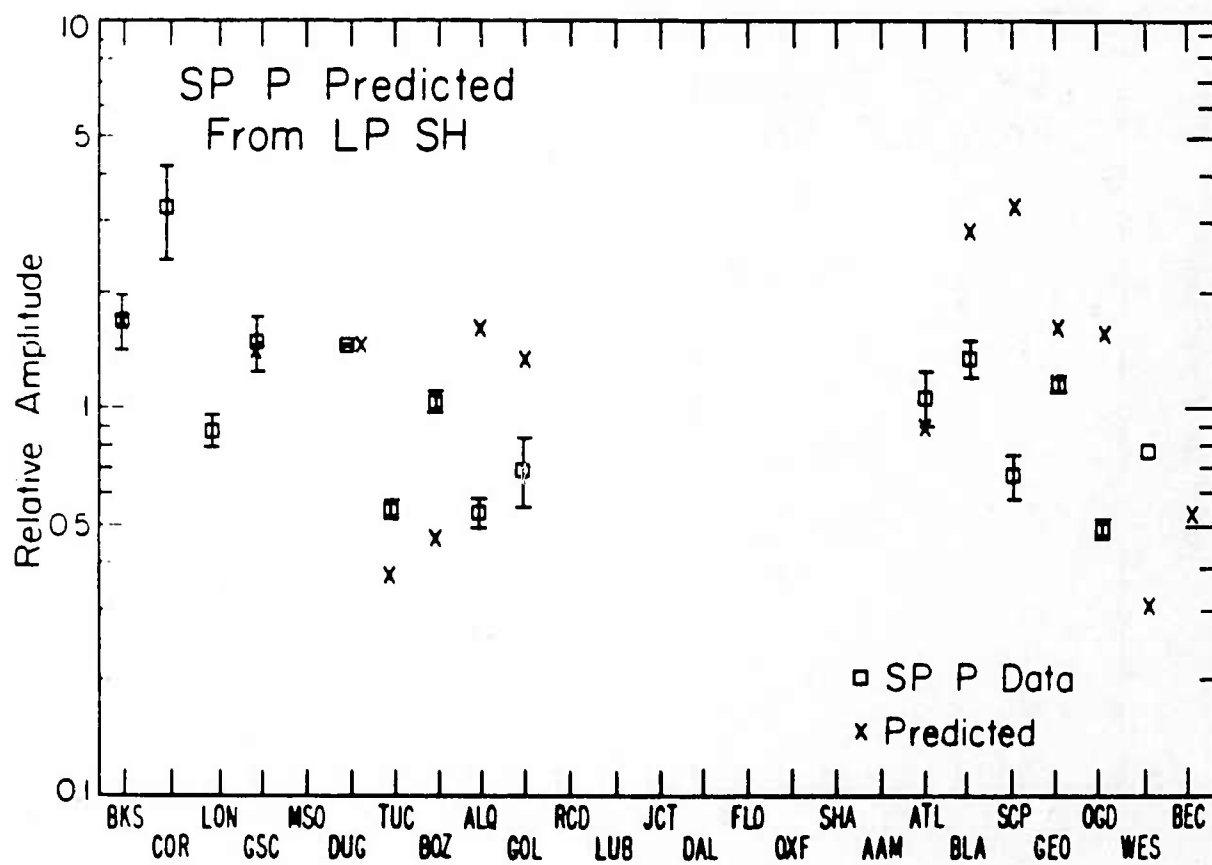


Fig. 26

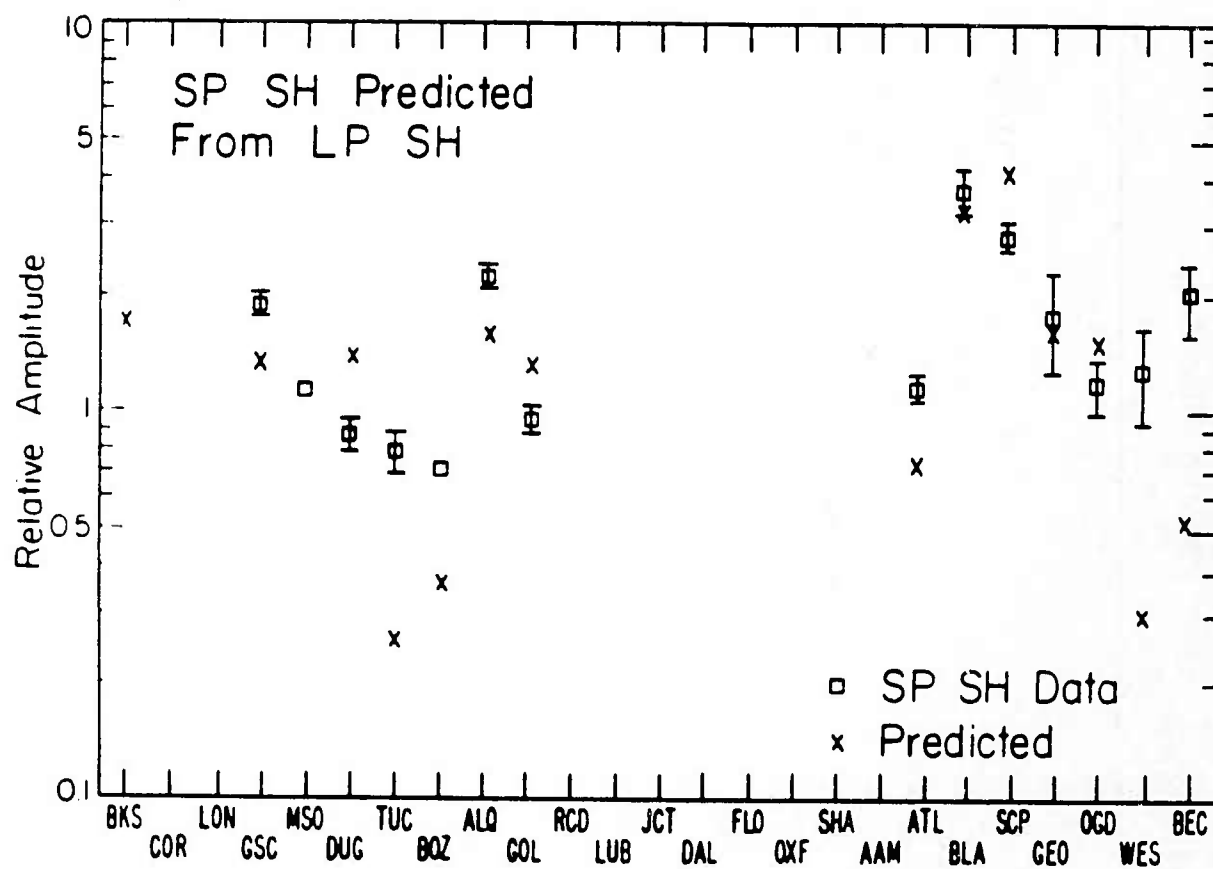


Fig. 27

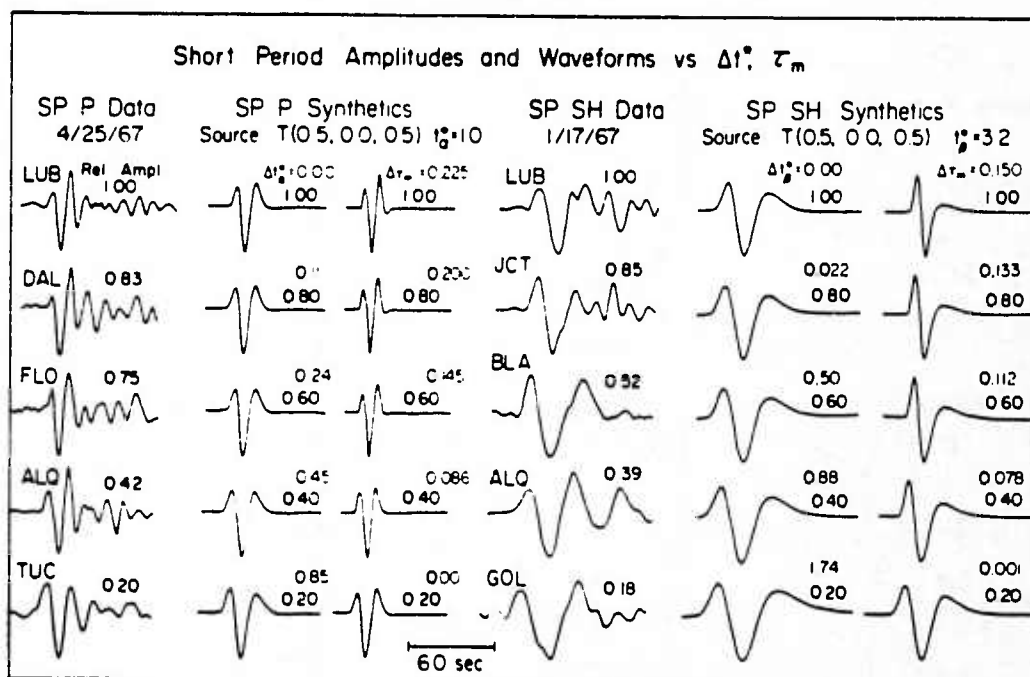


Fig. 28

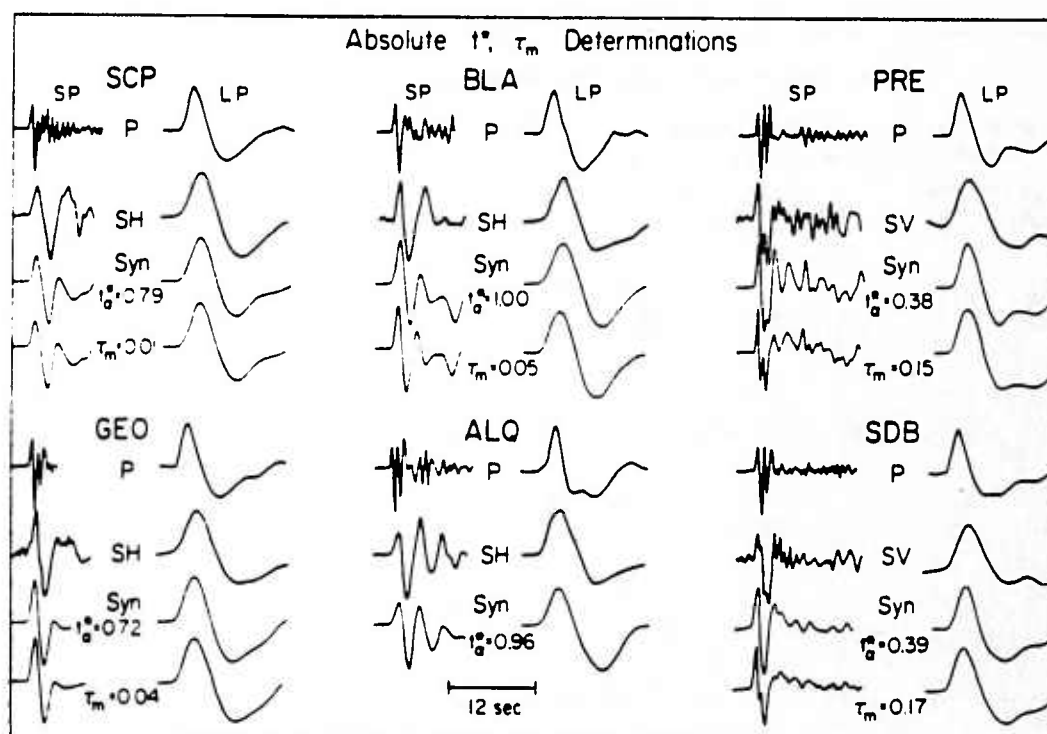


Fig. 29

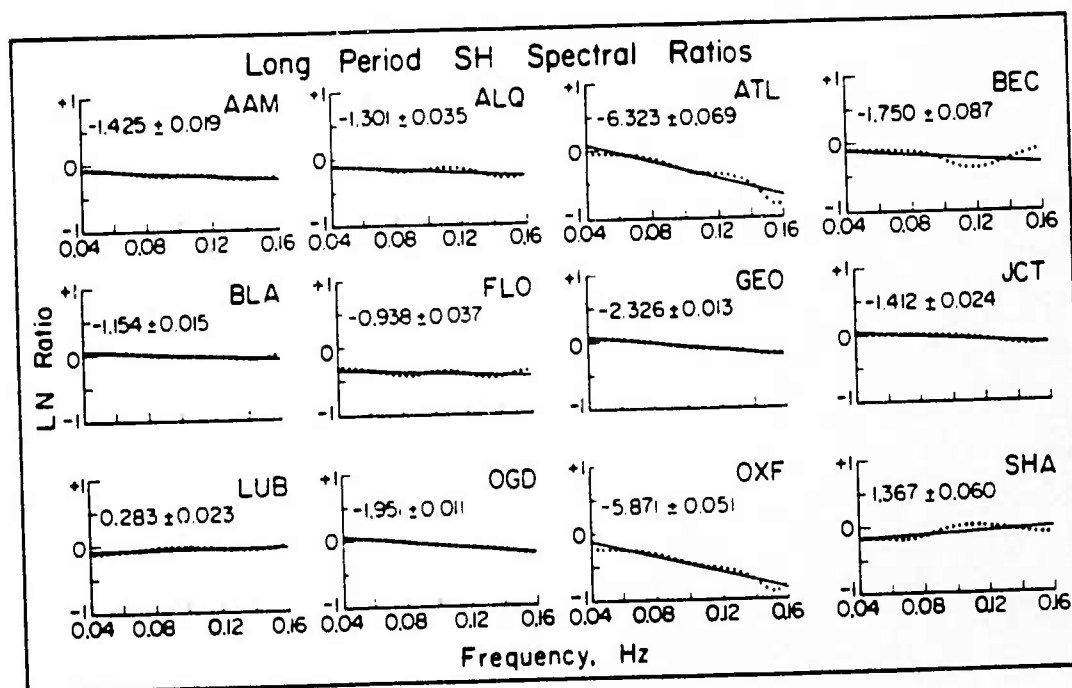


Fig. 30

Multimode analysis of Rayleigh-type Lg I: Theory and
applicability of the method

Michel Cara

J. Bernard Minster

MULTIMODE ANALYSIS OF RAYLEIGH-TYPE Lg I: THEORY
AND APPLICABILITY OF THE METHOD

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Abstract

Rayleigh-type Lg propagating in a laterally homogeneous continental crust can be synthesized by adding only a few overtones at periods greater than 2 seconds. Under minimal assumptions, we show that wavenumber analysis of Lg recorded on a several hundred kilometers long linear array of 10 stations should allow us to isolate the different overtones, providing a tool to study both crustal structures and excitation of the overtones at the source. In this first paper, we use synthetic Lg seismograms to investigate the applicability of a time-frequency-wavenumber analysis technique (UC diagram algorithm) to realistic arrays of stations. The behavior of the algorithm in the presence of lateral heterogeneities is studied numerically by introducing either random or coherent phase perturbations. We find that 1) the method is tractable if random phase fluctuations from station to station are spread over less than half a cycle, and 2) coherent velocity changes between two halves of a profile are spatially averaged if they are too small to be resolved by the array.

I. Introduction

The Lg phase has been first described by Press and Ewing (1952) as a transverse impulsive phase propagating over several thousand kilometers across North America at crustal S-velocities. For paths greater than 1000 km, Lg is a distinct phase while it merges with Sg at shorter distances. Significant radial and vertical components of Lg were also recognized soon thereafter (Lehmann, 1953; Gutenberg, 1955). Båth (1954, 1956) called Lg₁ and Lg₂ groups propagating respectively at 3.5 and 3.3 km/s, and pointed out the complexity of this phase.

Early explanations involving rays guided in crustal low velocity channels (Gutenberg, 1955) are no longer necessary since normal mode theory has shown that Love and Rayleigh modes propagating in any realistic continental structure exhibit group velocity extrema at periods smaller than 10 seconds, which can be associated with Lg arrivals (Oliver and Ewing, 1957, 1958; Kovach and Anderson, 1964). It has been demonstrated more recently that interpretation of Lg in terms of higher modes is valid not only from the point of view of group velocities but also from the point of view of amplitude excitation at the source, mainly for mid-crustal earthquakes (for Love waves, see Knopoff et al., 1974, and for Rayleigh waves, Panza and Calcagnile, 1975).

On the other hand, little progress has been made in observational analysis. Observations are usually very rough: existence or non-existence of Lg, arrival time, peak to trough amplitude (Ruzaikin et al., 1977; Chinn et al., 1980). Explanation of the full observed Lg record is far from being achieved. For example, the Lg wave train does not terminate at group velocities of 3.1-3.2 km/sec, as predicted from normal mode theory for realistic models of continental crust. As pointed out by Ruzaikin et al. (1977) this probably represents direct evidence of scattering, which is a source of obvious difficulty in modeling the signal.

In this paper, our approach is to analyze Lg by using a phase velocity criterion and wavenumber analysis. We shall show that stacking methods applied to a linear array of stations several hundred kilometers long should allow us to isolate the first few higher Rayleigh modes in

the period range 2-5 sec provided that the crust is sufficiently homogeneous laterally. In a companion paper (Part II) we will apply such methods to Lg recorded on the SCARLET array through the CEDAR system in Southern California, and on the CALNET array in Central California.

II. Preliminary Conditions for Multimode Analysis

The Fourier transform of a continuous signal spatially sampled by a linear array of stations can be expressed as the convolution product of the spatial Fourier transform of the signal with the "array response"

$$r(k) = \sum_{n=1}^N \exp i k x_n ,$$

where k is the wavenumber and x_n the spatial coordinate of station n . The basic wavenumber resolution of a linear array, δk , depends on the total length L of the array ($\delta k \sim 2\pi/L$). For equally spaced stations, $r(k)$ is a periodic function of period Δk which is inversely proportional to the spacing e between stations ($\Delta k = 2\pi/e$). Such a periodicity, or pseudo-periodicity in case of nonequally spaced stations, is responsible for spatial aliasing.

Spatial aliasing.

Spatial aliasing occurs if the wavenumber spectrum is spread over an interval wider than Δk . Successful retrieval of the actual wavenumber spectrum of a continuous signal from Fourier analysis of observations made at discrete points in space thus requires some a priori knowledge of the wavenumber content of the signal. For a time-signal

this problem is solved simply by use of analog anti-aliasing filters but no such device is of course possible for arbitrary space-signals.

In the case of a pure seismic signal propagating in a laterally homogeneous structure, it is easy to determine the wavenumber content of the signal from normal mode theory. In particular, at a fixed period and within a narrow group-velocity window only a few modes have to be considered if the period is long enough. This is the concept applied by Cara (1976) and Nolet and Panza (1976) to the analysis of Sa at periods longer than 20 sec. The Lg situation is illustrated on Figure 1 which displays the phase and group velocity curves for Rayleigh modes propagating in a flat realistic model of the crust in Southern California (Kanamori and Hadley, 1975). For periods greater than 2 sec, fewer than five overtones are expected to arrive in the group velocity window 3.2-3.6 km/sec. In the case of such pure multimode Lg, the wavenumber spectrum is restricted to a narrow band. For example, in the group velocity window 3.2-3.6 km/sec and at a period of 2.5 sec, the slowest and fastest phase velocities in Figure 1 are found for the first and fourth overtones, respectively. To avoid aliasing, the wavenumber periodicity Δk must then be greater than $k_4 - k_1$ and the critical spacing between station is $e_c = 2\pi(k_4 - k_1) \sim 55$ km, where $k_n = 2\pi/C_n T$ and C_n is the phase velocity of the n^{th} overtone at period T . Note that we do not take the fundamental mode (Rg phase) into account here, because its group velocity at periods of a few seconds is very low in the presence of low velocity superficial layers. Furthermore, the great sensitivity of that mode to low Q , highly heterogeneous shallow structure causes it to be strongly attenuated at epicentral distances of a few hundred kilometers.

The value of e_c thus estimated is slightly model dependent, but crust and upper mantle structure is known well enough that e_c cannot differ greatly from the above value. e_c is also period dependent but it is easy to show from Figure 1 that 55 km is an approximate lower bound in the group velocity window 3.2-3.6 km/sec and at periods between 2 and 5 seconds.

Now, if some signal or noise with a different wavenumber spectrum arrives in the above group-velocity window -- for example, any signal arriving from a different azimuth in case of non-cylindrical symmetry -- the above conclusions are no longer valid and spatial aliasing can be a source of problems. The applicability of wavenumber analysis of Lg to arrays with such wide spacing between stations is thus a priori limited to circumstances where the signal-to-noise ratio is high, particularly where signal-generated noise due, e.g., to multipathing is concerned.

Resolution

In order to isolate the different modes interfering in Lg through wavenumber analysis, we have to use a long enough array: δk must be smaller than the wavenumber spacing between modes. For the model used in Figure 1, the array length must be at least 225 km at 2.5 sec and 500 km at 1.5 sec but in contrast to the critical spacing e_c , this estimate is more strongly model-dependent. For instance, if a low velocity layer exists in the crust, the phase velocity curves can present kinks (Andrianova et al., 1967; Panza et al., 1972) and come so close to each other at some periods that an unrealistic array length would be required to isolate the modes.

A low velocity channel might be present near 15 km in southern California as suggested for example by fundamental Rayleigh mode data (Hadley and Kanamori, 1979). However, numerical experiments have shown that such phase velocity kinks are not likely to occur between 2 and 5 seconds but might be found at periods smaller than 2 seconds.

We can thus state the following preliminary set of conditions for multimode analysis of Rayleigh-type Lg in the period range 2-5 seconds using linear receiver arrays: The total length of the array must be 200 to 500 km and the spacing between the stations smaller than 50 km. A 300 km long array of 10 stations thus appears to be adequate for this purpose.

III. UC Diagrams

Several wavenumber analysis techniques have been developed for the study of multimode surface waves. To our knowledge, they have only been applied successfully to the Sa phase at periods greater than 20-30 sec (Nolet, 1975; Cara, 1978). A simple stacking algorithm, the UC-diagram technique (Cara, 1976, 1978) seems appropriate for Lg in a first study because it needs only little a priori information about the signal. In fact, one needs only an approximate idea of the wavenumber content of the signal (total bandwidth, spacing between modes) to fix the geometry of the array as discussed previously, and an estimate for the central group velocity of the multimode wavepacket which is used in this algorithm to stack the records (Cara, 1976). Other techniques, such as band-pass wavenumber filtering to isolate a mode once its "sub-array" phase velocity is known (Cara, 1978), require more information about the signal and could

be used for amplitude and source excitation studies, but only after it has been recognized on UC diagrams that clear pure mode signals exist in the records.

The basic idea underlying the UC diagram algorithm is to stack N multimode records after narrow-band filtering and phase shifting in order to insure constructive interference of individual modes. Let $F(x_n, \omega)$ be the time Fourier transform of a record $f(x_n, t)$ observed at epicentral distance x_n , and let

$$G(k, \omega) = \sum_{n=1}^N F(x_n, \omega) \exp[-\alpha(\omega - \omega_0)^2] \exp[-i K(\omega) x_n]$$

where

$$K(\omega) = k + (\omega - \omega_0) / U_c.$$

In these expressions, ω_0 is the circular frequency at which analysis is to be performed and α controls the width of the Gaussian frequency filter. U_c is an input parameter, chosen equal to the central group velocity of the multimode wave packet and U_c^{-1} is approximately the first derivative of $K(\omega)$ at ω_0 -- in this paper we choose $U_c \approx 3.3$ km/sec. The modulus of the inverse Fourier transform $g(k, t)$ of $G(k, \omega)$ exhibits peaks at $k_p = \omega_0 / C_p$ and $t_p = \bar{x}(U_p^{-1} - U_c^{-1})$ where C_p and U_p are respectively the phase and group velocity of mode p and where $\bar{x} = \sum_{n=1}^N x_n / N$ (Cara, 1976). The UC diagram technique consists in producing a set of contour plots of $|g(k, t)|$, conveniently plotted in terms of the hyperbolic coordinates $c = \omega_0 / k$, and $U = \bar{x} / (t + \bar{x} / U_c)$, for a sequence of values of ω_0 . Should dispersion properties vary slightly with epicentral distance along the

profile, then it is important to note that U represents the group velocity between the source and the center of the array, and C is the phase velocity along the array (Cara, 1977). Being computed for a given, fixed period, a UC diagram contains information about the average dispersion properties of the wave train in a narrow frequency interval. In particular, peak shapes in the U direction depend mainly on the frequency filter parameter α . In this paper, we chose α such that the total bandwidth of the Gaussian filter at -30 db be $\omega_0/2\pi$.

For purposes of illustration we computed a set of synthetic L_g trains (Figure 2) along a 351 km long array of non-equally spaced stations. We used the dispersion curves shown on Figure 1 and, based on our previous discussion, only included the fundamental Rayleigh mode and five overtones in the synthetics. These modes have been summed assuming a very simple constant unit amplitude for periods longer than 2 sec for each mode (0.5 for the fundamental) and a linear amplitude variation between 0.5 and 2 seconds. The signals have also been convolved with a standard seismometer-galvanometer response ($T_s = T_g = 1$ sec, $\alpha_s = \alpha_g = 1$).

Figure 3 depicts a typical UC diagram calculated from this set of synthetic records at a period of 2.5 sec. The first three overtones are clearly identifiable, but the fourth one is already strongly dispersed at this period and does not generate a peak. Residual interference between the second and third overtones is probably responsible for the biased location of the third peak. Mode amplitudes should be computed for realistic source models but this is beyond the scope of this paper: our intent here is to show that modes can be isolated from L_g in a situation where interference effects between modes play a major role. Choosing the

same amplitude for all overtones can be considered as a modelization of the worst situation we might have to face for this purpose. If one mode is more strongly excited than the others, it will be of course easier to isolate it from Lg records.

Mode identification from actual data

Processing records through the UC diagram technique yields a set of peaks in the UC plane at different periods. If for some reason one, or several modes are weak or absent in the signal (due to weak mode excitation at the source, attenuation along the path, etc.), the identification of an observed peak as a mode of known rank is not obvious.

Figure 4 displays, for example, the variation of overtone phase velocity for two models. The model shown in Figure 1 (primed numbers in Figure 4) exhibits a "normal" sub-Moho S velocity (4.5 km/sec) and the depth of the Moho is 32 km. The second model (unprimed numbers in Figure 4) has a homogeneous 3.5 km/sec S velocity in the crust, a Moho at 32 km and a low S velocity upper mantle ($V_g = 4.2$ km/sec). If the total sequence of modes is not present on an actual UC diagram, it is clear from Figure 4 that identification of a peak with a mode of given rank can be a problem in cases where a priori estimates of model parameters are uncertain. On the other hand, if crustal parameters are known a priori with sufficient accuracy, mode identification of observed peaks should not be a problem for periods greater than 2 or 3 seconds.

IV. The Effect of Phase Perturbations

In order to isolate Rayleigh overtones near 2.5 sec period from Lg records, we require arrays over several hundred kilometers long. We have assumed implicitly so far that lateral heterogeneities are weak enough that they do not invalidate the multimode representation of Lg. This is a rather strong assumption for these short period crustal waves, and it is desirable to fix quantitative limits on permissible departures from the ideal, laterally homogeneous case.

For our purposes, the main effect of inhomogeneities is to partially or totally alter the constructive interference sought in UC diagrams. In what follows, we define phase perturbation as any change in the phase of a mode due to departure from a laterally homogeneous structure. Modeling overtone propagation for realistic laterally varying structures, including topography, is a very difficult task which lies well beyond the scope of the present paper. However, order of magnitude effects can be estimated in two especially simple cases. They are 1) random fluctuations (statistically independent from station to station) and 2) coherent phase perturbations within a subgroup of stations. We attempt to quantify those effects below.

Let us first note that, irrespective of the cause of phase perturbations, for a fixed overtone n , at a fixed period T , the apparent phase at an epicentral distance x_1 is of the form $\phi_n(x_1, T) + 2\pi\delta t_n(x_1, T)/T$. Here, ϕ_n is the phase calculated for the unperturbed model (e.g., a spatially averaged structure for the area), and δt_n is a time term which accounts for the local phase perturbation. In a multimode signal, δt_n

depends a priori on n , since different modes sample the structure differently and thus are not sensitive to the same heterogeneities. A more precise statement would necessitate considerable analysis, but is fortunately not required for our purposes. We have perturbed synthetic Lg records under the simplifying assumptions that 1) δt_n is independent of n and 2) the mode amplitudes are not affected, and have submitted the resulting profiles to UC analysis. In that case, perturbations reduce to simple time shifts of the multimode wavetrain if one looks at a single period.

The two types of perturbations discussed above (random and coherent) are then modelled respectively through the two relations

$$\delta t_n(x_i, T) = \Delta t \text{ rand}(i)$$

or

$$\delta t_n(x_i, T) = (\epsilon/V)(x_i - x_p), \quad x_i \geq x_p$$

where $\text{rand}(i)$ is a random number with uniform probability on $[-0.5, 0.5]$, Δt a scaling parameter, V a reference velocity, ϵ a relative velocity offset for the second portion of the profile.

Random fluctuations

Figures 5 and 6 correspond to two sets of random numbers at $\Delta t = 1$ sec. The UC coordinates of overtone peaks remain very stable in this case, but spurious peaks are sometimes generated (Figure 6). Further numerical experiments with Δt ranging from 0.5 sec to 1.5 sec and other sets of random time shifts show that apparent phase velocities for modes 1 and 2 were perturbed by less than 0.02 km/sec but that the peak amplitudes

could be strongly affected, particularly for the larger values of Δt . Stronger perturbations were obtained for mode 3 (± 0.06 km/sec). It is interesting that the unperturbed profile already yields a slightly biased estimate of this peak due to a lack of wavenumber resolution, as mentioned earlier.

Figure 7 exhibits a much higher perturbation level for $\Delta t = 2.5$ sec, which coincides with the period of the UC diagram. In this particular example, peaks are still recognizable at nearly correct locations, but many spurious peaks have been generated too. In fact, other sets of random numbers lead to such changes in the topography of the diagram that overtones are not recognizable.

Although this numerical experiment is based on an oversimplification, it clearly indicates that the UC diagram technique remains stable under phase fluctuations with a random spatial distribution, and up to the order of half a cycle in magnitude. However, these fluctuations do seem to affect peak amplitudes by a large amount, so that even in cases where sub-array dispersion measurements can be achieved successfully, great caution must be exercised before interpreting peak amplitudes (in terms of overtone excitation at the source, for example).

Coherent fluctuations

In a second numerical experiment, the same set of synthetic L_g records were left unperturbed for epicentral distances shorter than 412 km, and a time shift proportional to distance was applied between 412 and 580 km. This mimics the effect of a relative perturbation in the phase velocity of the form

$$\Delta C/C = -\epsilon C/V .$$

Figures 8 and 9 show UC diagrams for $\epsilon = 2.5 \times 10^{-2}$ and 5×10^{-2} , respectively, and $V = 3$ km/sec (phase velocity variations of -3.25% and -6.5% for $C \approx 3.9$ km/sec, the central phase velocity for these UC diagrams).

On Figure 8, longitudinal variations of about 3% in phase velocity (depending on the mode) both broaden the peaks and shift them toward lower phase velocities, as expected. Stronger variations of about 6% (Figure 9) lead to clear splitting of the peaks for modes 1 and 2. For these two modes we obtain therefore two phase velocity estimates which correspond each to a portion of the profile. Thus it seems that UC peak splitting results when coherent phase velocity changes reach about 5% along half the profile. Note that this corresponds to the typical phase velocity spacing of modes near 2.5 sec period. This is also the intrinsic resolution of the array, because of the way it was constructed a priori for overtone separation.

These two numerical experiments permit us to place quantitative limits on the applicability of the method. It yields stable results, averaging possible phase fluctuations along the station array, provided that 1) statistically independent random phase fluctuations from station to station have a spread of less than half a cycle, and 2) coherent velocity variations between the two halves of the profile are less than the resolution of the array. When these conditions are violated, spurious peaks can be generated and peak splitting can occur. Of course, in this case, because the wavenumber resolution of the array is about the same

as the wavenumber spacing between modes, a representation of the Lg wave-field in terms of overtones with coherent propagation over several hundred kilometers loses all significance.

Finally, we must add that the critical conditions described in this section touch on problems of a different nature than the problems discussed in detail by Nolet and Panza (1976). These conditions are more directly related to the physics of wave propagation on actual structures than to the geometrical details of the array. It is also interesting to point out that for the Sa phase, another type of multimode wave of longer period, the maximum observed lateral variations for the first Rayleigh overtone reach 0.2 km/sec at 50 sec period (Cara et al., 1980) while the phase velocity spacing between first, second and third overtones is greater than 0.6 km/sec for realistic earth models. This is probably the reason why stacking methods such as the UC diagram technique work well for long period overtones even in the case of strong lateral upper mantle heterogeneities, as is the case across the whole continental United States WSSN network for example (Cara, 1978).

V. Conclusion

This short study shows that although severe conditions are required for retrieving modes from Lg records, this is theoretically possible in the period range 2-5 sec, even at a few hundred kilometers from the source, by using linear 300 to 500 km long arrays of about 10 stations. The more severe limitation by far seems to be that the crust be rather homogeneous laterally in the area under study.

Lateral arrivals of any kind can modify the wavenumber content of the signal. In this case, the low station density proposed here becomes a source of serious difficulties because of spatial aliasing. Another source of difficulty might be the mode identification from observed UC diagrams unless the crust is known with sufficient accuracy in the area under study to predict overtone phase velocities close enough to actual ones.

In the presence of lateral variation of structure under the array, the apparent phase of Lg signals can differ locally from the values predicted by a standard multimode representation of the wavefield. A numerical experiment involving random as well as coherent phase perturbations shows that the UC diagram technique yields stable estimates of spatially averaged phase velocities for the different overtones, provided that 1) random phase fluctuations from station to station remain smaller than half a cycle, or 2) apparent velocity variations between two halves of a profile remain less than the resolution of the array and thus less than the phase velocity spacing between modes (e.g., ~5% at 2.5 sec period).

If such conditions are satisfied, multimode analysis of Lg on arrays of stations should provide a powerful tool to study both details in the crustal structure and crustal overtone excitation at the source. Actual applications to data recorded on the SCARLET Array through the CEDAR system and on the CALNET array in Central California are described in a companion paper.

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Figure Captions

- Figure 1. Dispersion curves of Rayleigh overtones 1 to 5 computed for a realistic model of Southern California. Dashed line: group velocity; solid line, phase velocity. The S-velocity model in the inset is derived from the P velocity model of Kanamori and Hadley (1975) by setting Poisson's ratio to 0.25.
- Figure 2. Synthetic Lg and Rg phases computed by summing the first six Rayleigh modes with dispersion curves as shown in Figure 1, along a realistic array of stations. (Station distribution corresponds to a quasi-linear sub-array of SCARLET.)
- Figure 3. UC diagram after stacking the synthetic records of Figure 2 (period 2.5 sec; contour spacing: 10% of maximum amplitude). The dots show the theoretical values for phase and group velocities. The vertical coordinate varies linearly with $1/U$ (time scale) and the horizontal coordinate with $1/C$ (wavenumber scale). The modulus of the array response $r(k)$ is shown here centered on the fundamental mode peak.
- Figure 4. Phase velocity curves of the first five Rayleigh overtones computed for two models. Solid lines (primed numbers) correspond to the model shown in Figure 1 and dashed lines (unprimed numbers) to a one layer crust -- low upper mantle velocity model (see text).

Figure 5. UC diagram obtained for synthetic records simulating the propagation of Lg on the profile shown in Figure 2, with randomly distributed time shifts spread over a 1 sec interval.

Figure 6. Same as Figure 5 for a different distribution of random time shifts.

Figure 7. Same as Figure 5 for time shifts spread over a 2.5 sec interval.

Figure 8. Same as Figure 5 for coherent time shifts simulating a -3.25% decrease of a 3.9 km/sec phase velocity on the second half of the profile.

Figure 9. Same as Figure 8 for a -6.5% decrease of velocity.

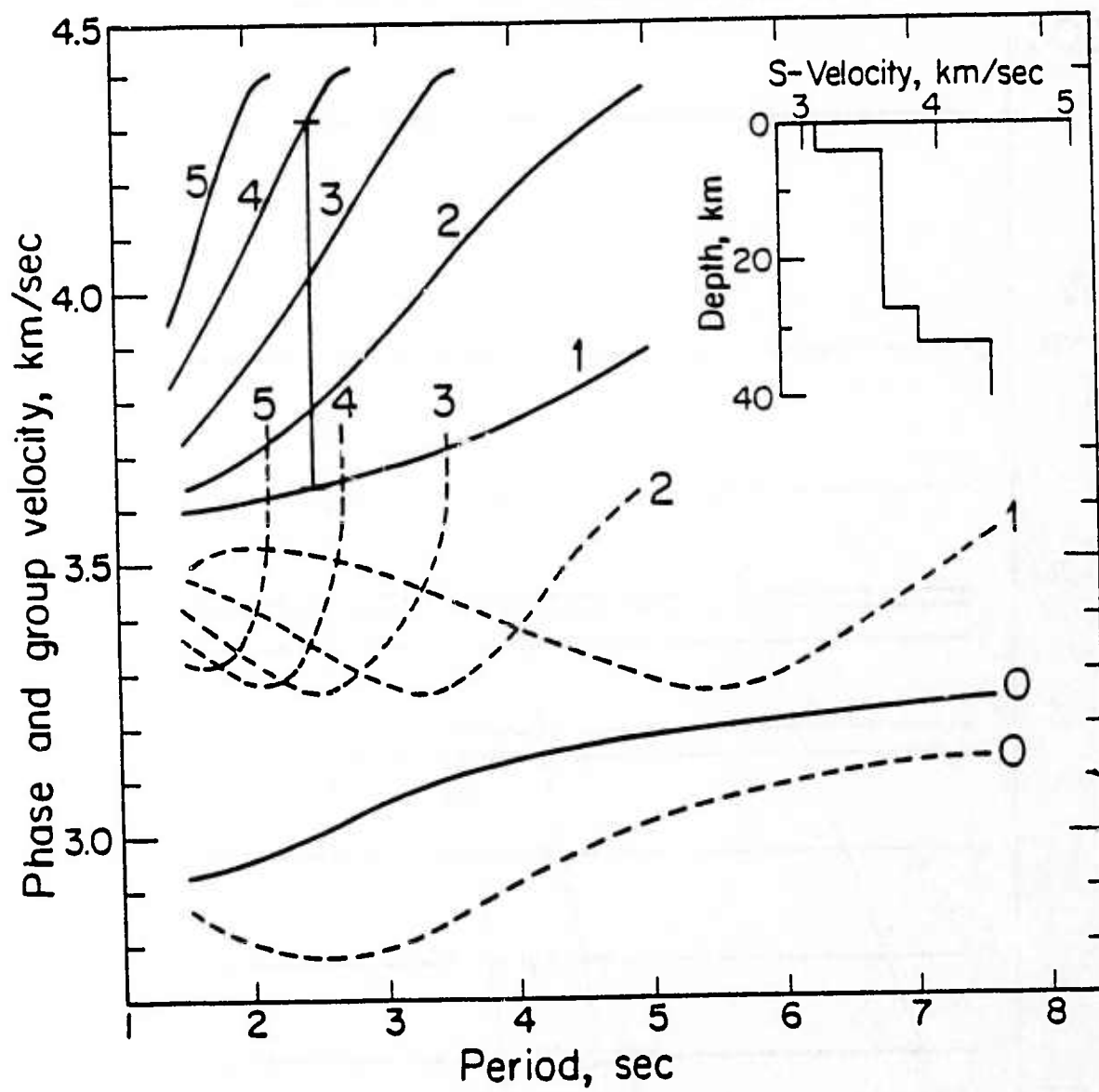


Fig. 1

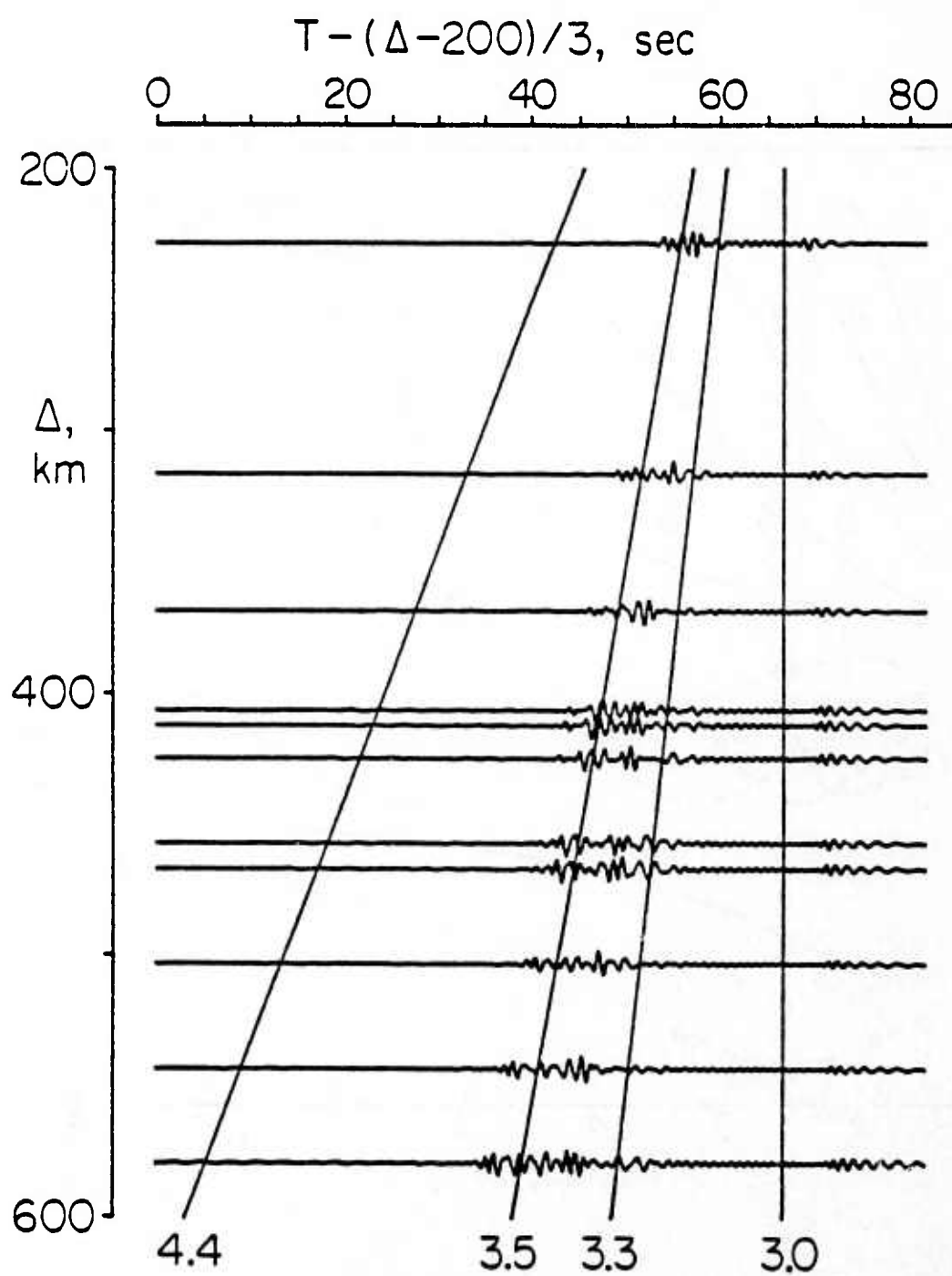
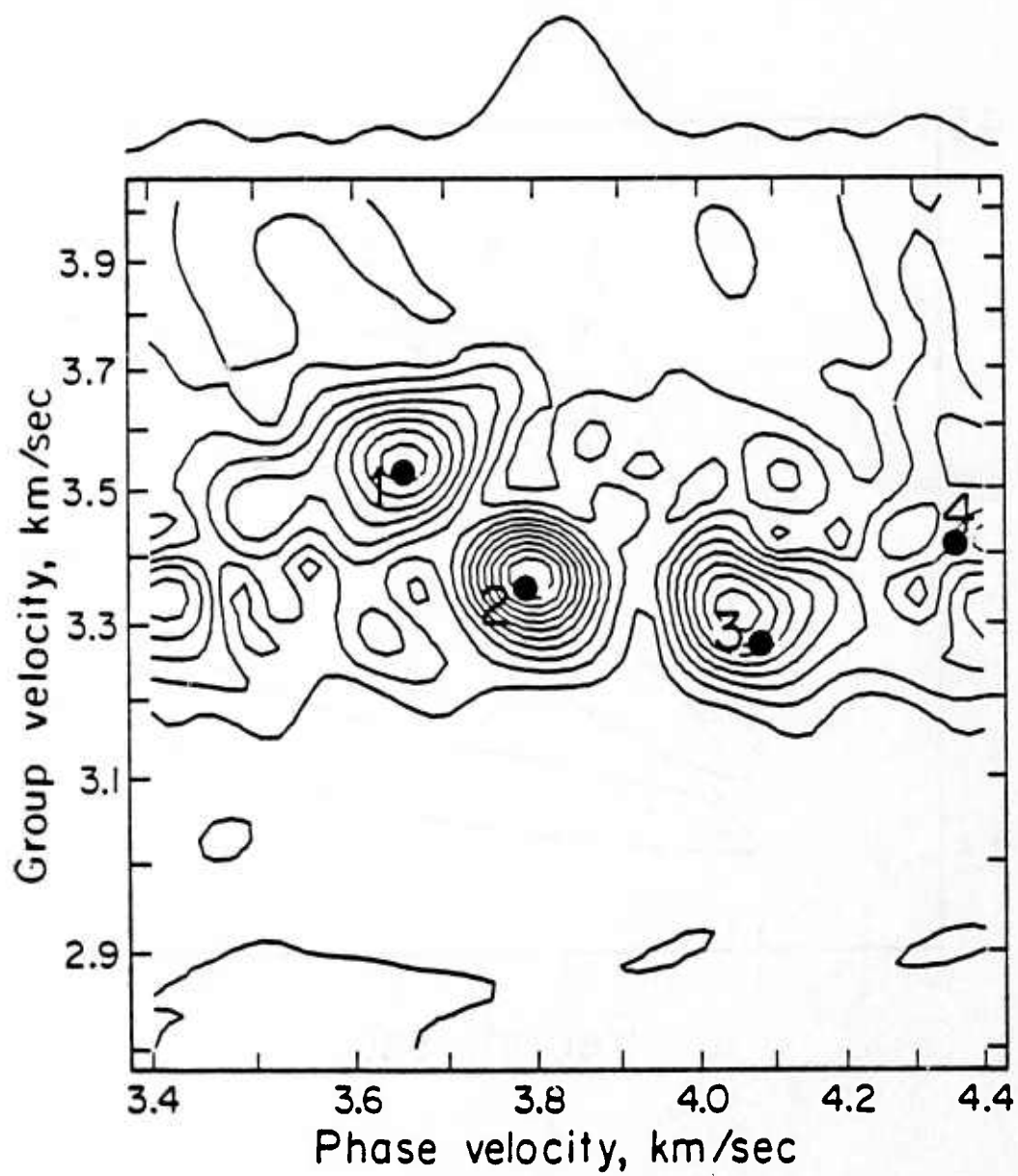


Fig. 2

*Fig. 3*

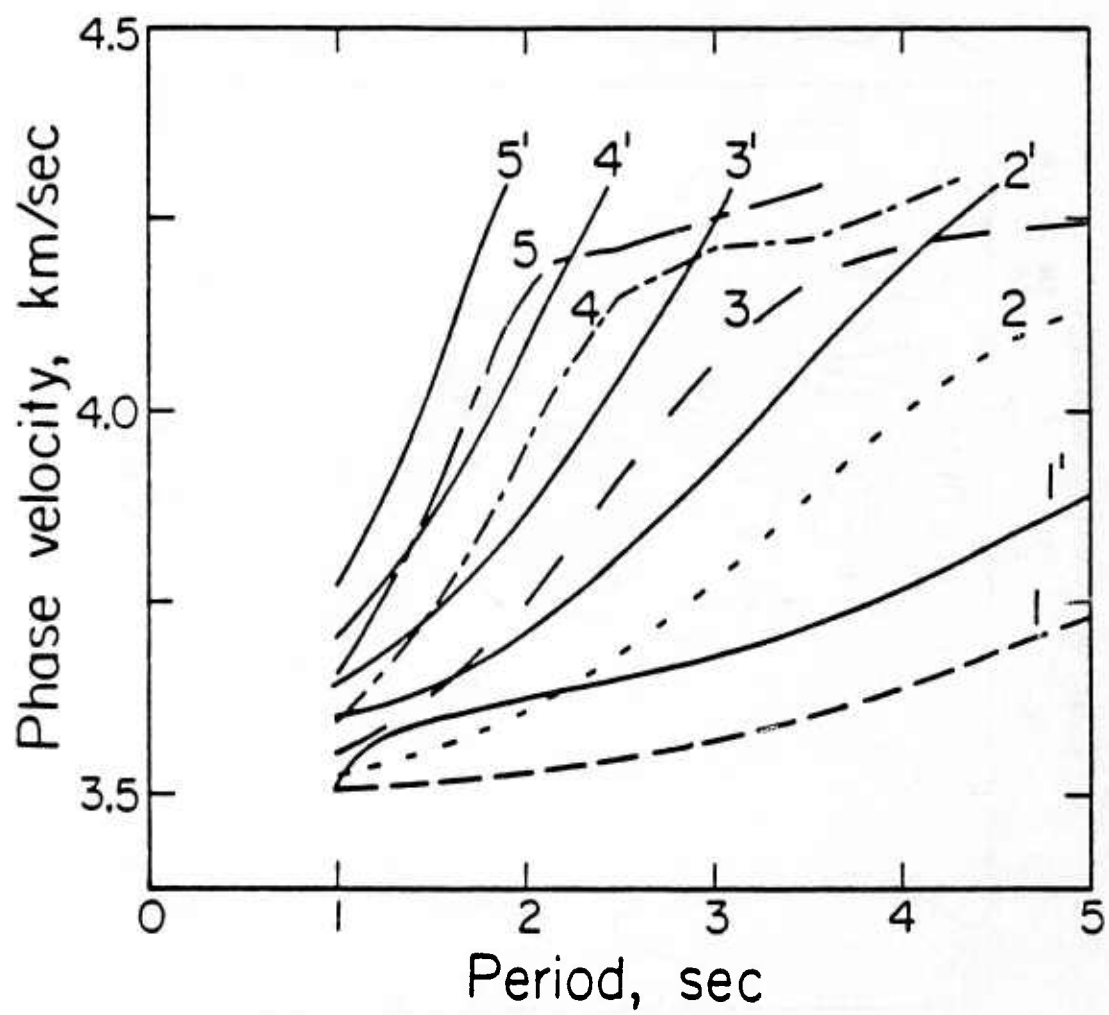
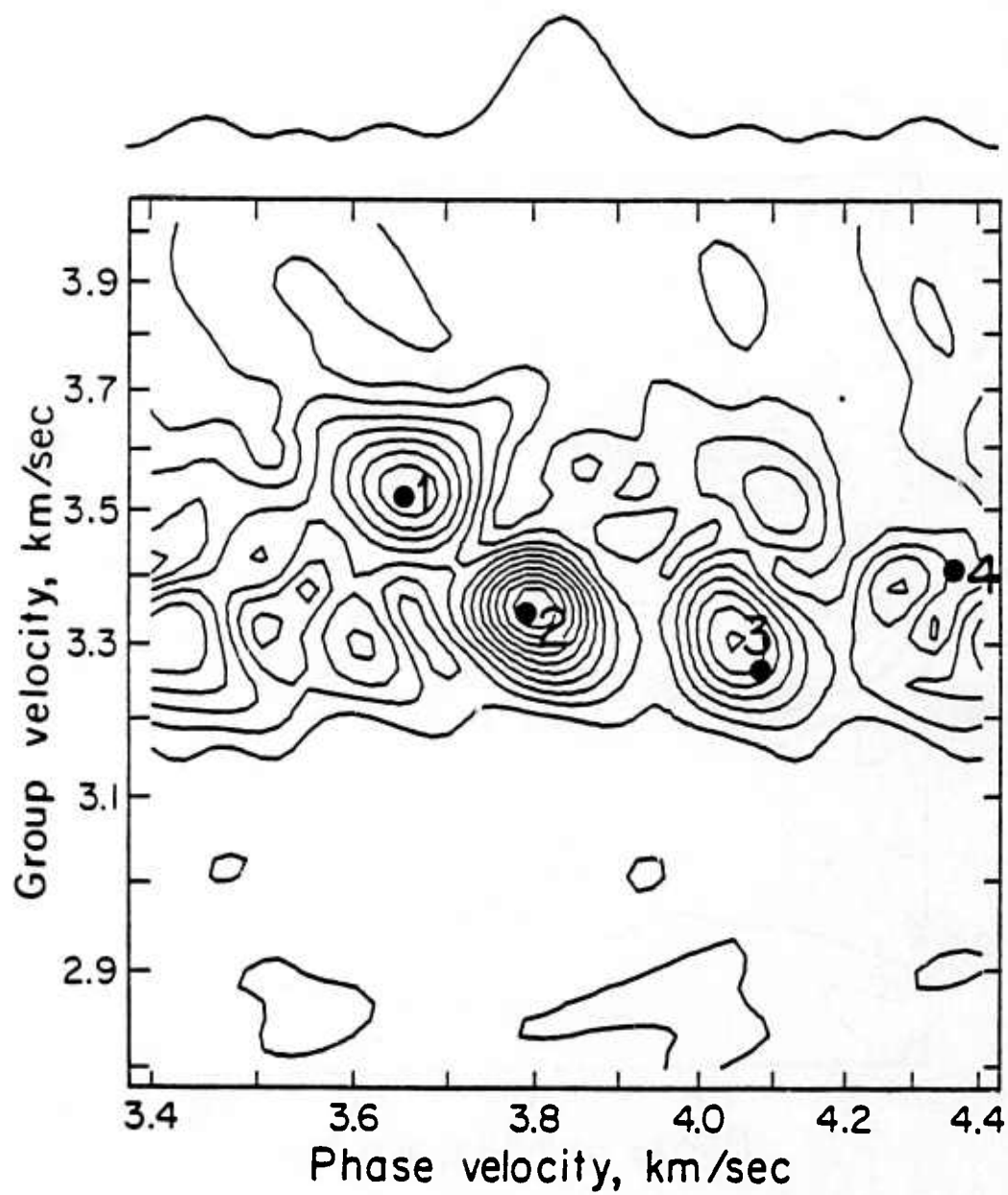
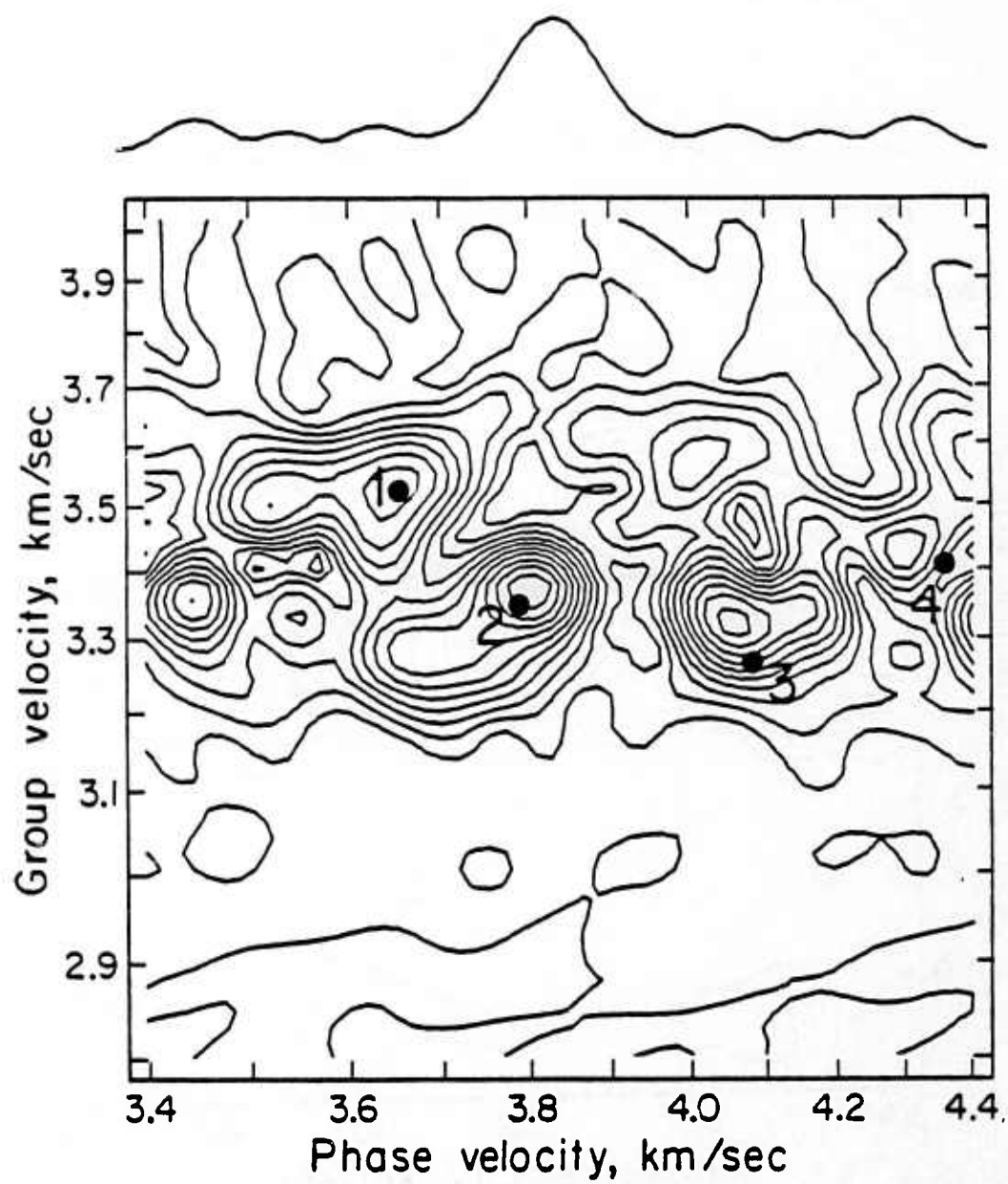
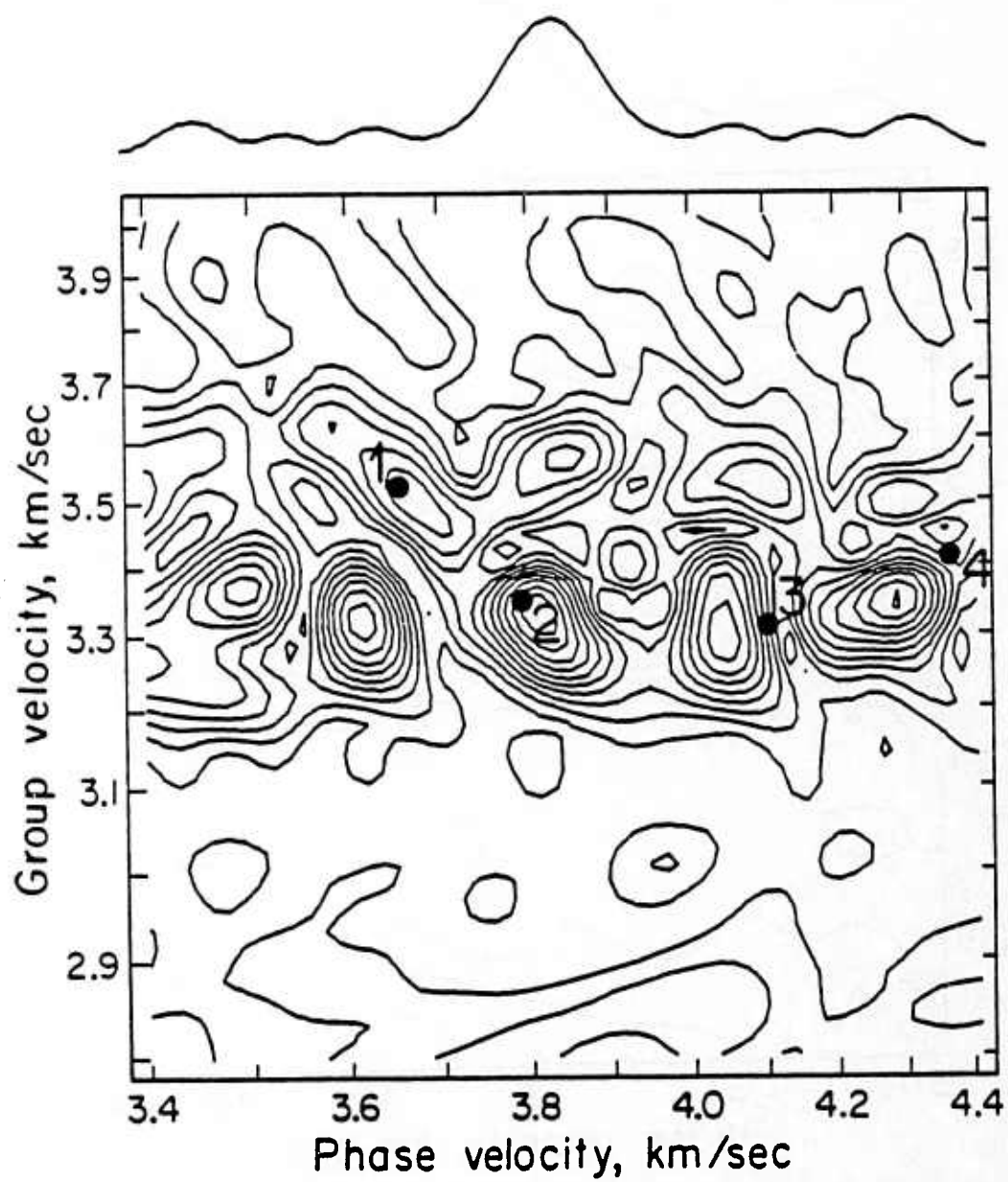
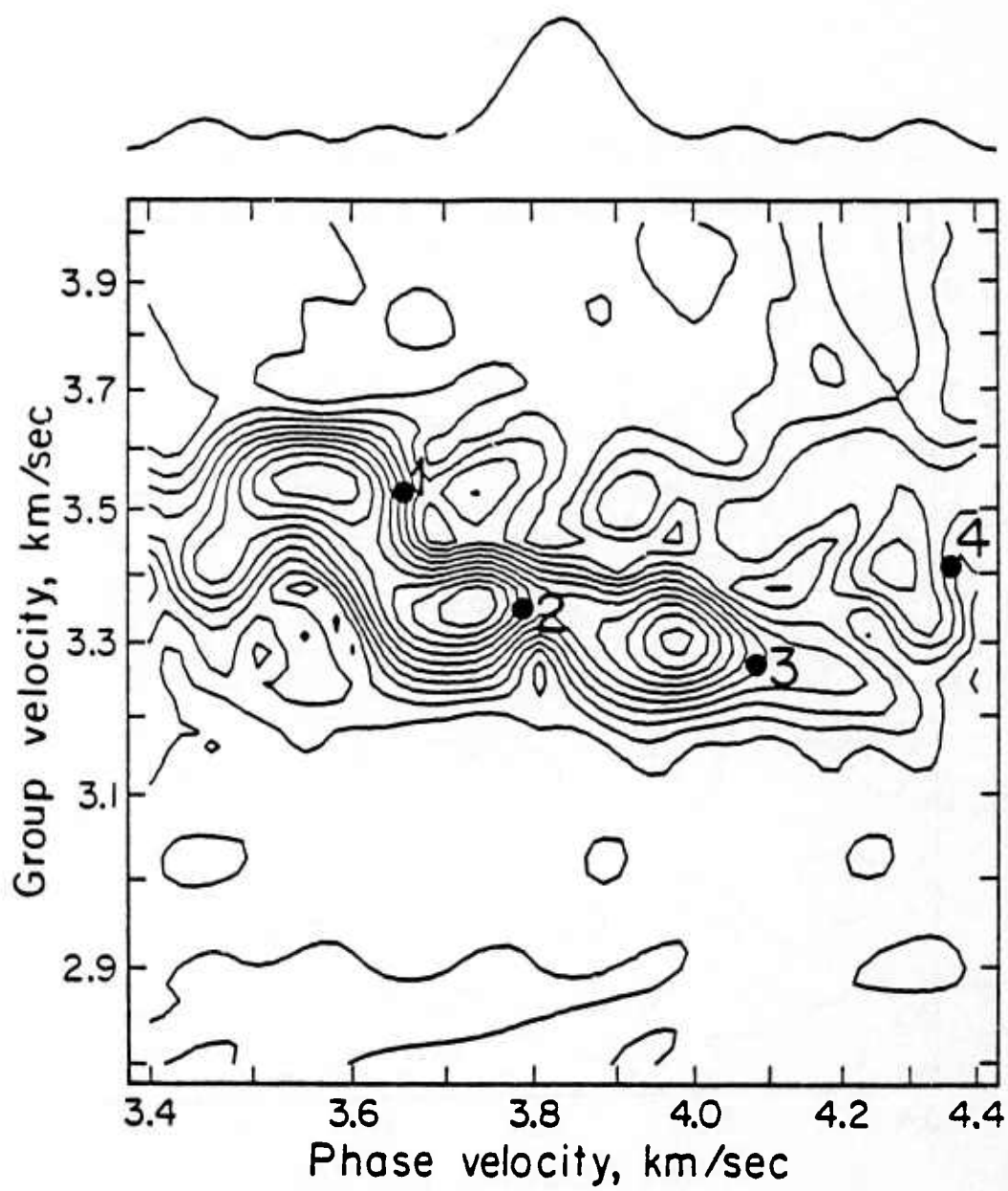


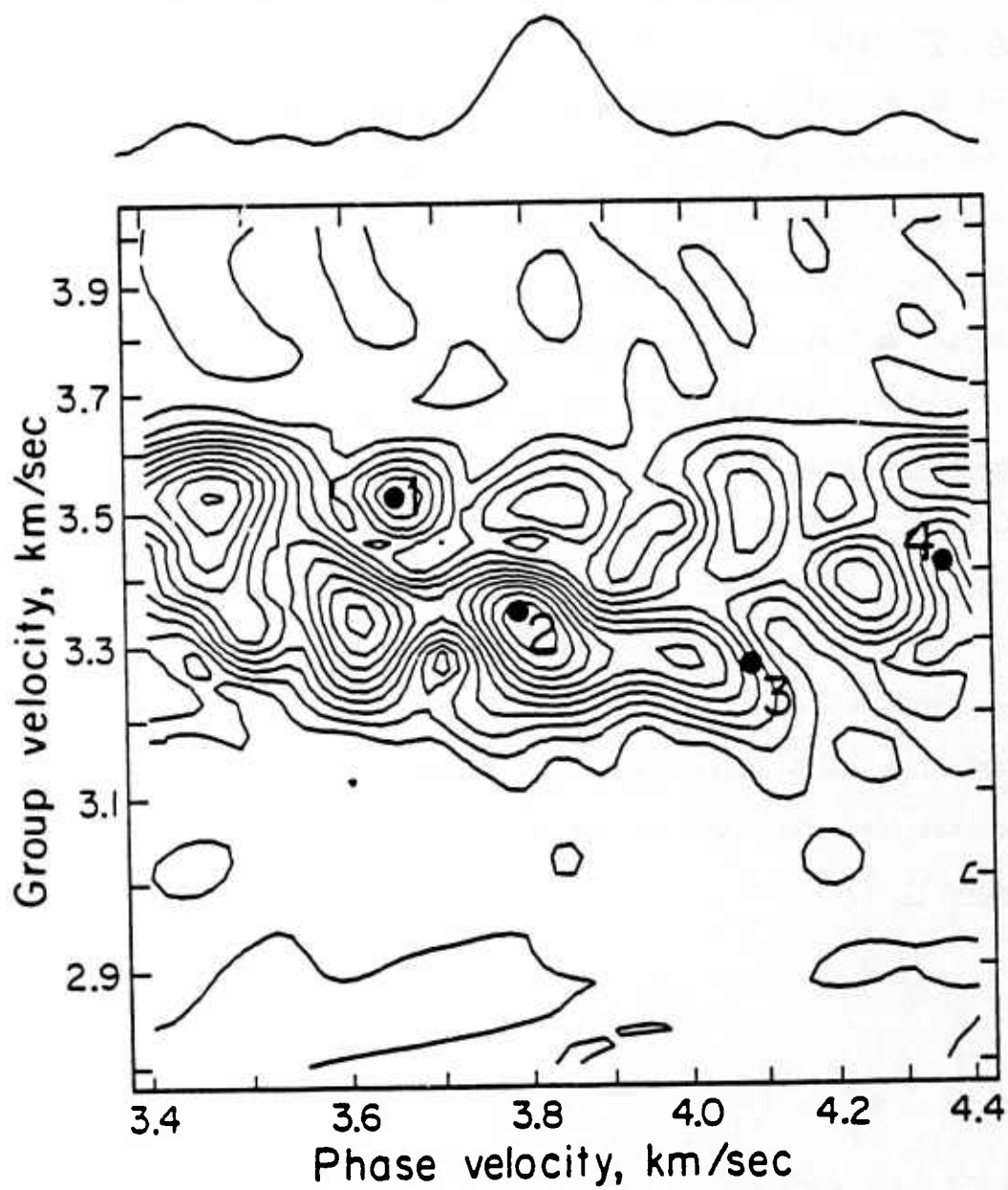
Fig. 4

*Fig. 5*

*Fig. 6*

*Fig. 7*

*Fig. 8*

*Fig. 9*

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Telesceismic short period amplitudes: source
and receiver variations

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TELESEISMIC SHORT PERIOD AMPLITUDES: SOURCE
AND RECEIVER VARIATIONS

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ABSTRACT

Short period P wave amplitude data from nuclear explosions in the Soviet Union recorded by WSSN stations in the United States are presented. Thirty-four events in five test sites are analyzed. The consistency and similarity of the initial P waveforms allow a stable amplitude measure. A well-defined amplitude pattern is obtained for each source region. The test sites at northern and southern Novaya Zemlya show a relative amplitude trend of a factor of three across the United States in their respective amplitude patterns. This is in contrast to two sites at Semipalatinsk which are in good relative agreement. A pattern of lateral variation of amplitude in the United States is obtained for a northern azimuth of approach. Stations situated on sediments are corrected for amplification effects. Stations in the western U.S. do not have systematically lower amplitudes than eastern U.S. stations, in contrast to previous studies. Lowest amplitudes are found in Golden, Colorado (Gol) and Albuquerque, New Mexico (ALQ), a factor of four lower than high amplitude stations. Preliminary amplitude data are presented from earthquakes in the Kuriles and South America. Events are chosen for consistency of waveforms across the U.S. to minimize earthquake source and directivity effects. These earthquake data indicate that amplitude variations in the United States are azimuthally dependent.

INTRODUCTION

Since the establishment of the World Wide Standard Seismograph Network (WSSN) and during the deployment of the Long Range Seismic

Measurement (LRSM) stations in the early 1960's a number of studies have undertaken the measurement of the variation of teleseismic body wave amplitudes in the conterminous United States. Cleary (1967) measured first peak amplitudes of P waves from short period LRSM recordings of 22 earthquakes, each having common first motions across the United States, and noted that the signal strength tended to be lower in the west than elsewhere, although exceptions did occur. Evernden and Clark (1970) and Booth, Marshall, and Young (1974) determined magnitude anomalies for stations of the LRSM and found short period (1 sec) magnitudes in the western United States - WUS, approximately west of the eastern front of the Rocky Mountains - are about 0.5 magnitude units (a factor of 3) smaller than the eastern United States - EUS, east of the Rocky Mountains. Long period magnitude (16 sec) anomalies for the LRSM show large amplitudes along the Gulf coast but no other distinct pattern (Booth et al., 1974). Solomon and Toksoz (1970) used a spectral ratio technique on long period WWSSN recordings of P and S waves from two deep focus South American earthquakes to determine apparent attenuation variation across the United States, and concluded that attenuation was higher between the Rockies and Sierra Nevada - Cascades, and the northeastern United States. Der, Masse, and Gurski (1975) using short period LRSM stations measured maximum amplitudes of P and S waves and the dominant period of S waves for 3 deep focus South American and two deep focus Kurile earthquakes and concluded that the data indicated higher attenuation in WUS than EUS. In a study using four deep focus South American earthquakes recorded by the WWSSN in the the United States, Burdick (1978) utilized

the ratios of short period to long period amplitudes of P and S waves to minimize the effect of source radiation in determining amplitude variations at the receiver. No lateral amplitude variation across the United States was seen in this data set.

The purpose of this paper is to report on the variation of P wave amplitudes in the United States from nuclear explosions in the Soviet Union. Amplitude data from 34 nuclear explosions at 5 test sites are presented in 24 WWSSN stations in the conterminous United States plus BEC, Bermuda. The results are interpreted in terms of source, path and receiver effects. Source effects are deduced by intercomparing amplitude patterns among the different sites. Data which have been diffracted at the core mantle boundary are presented, but will be discussed and interpreted in a forthcoming paper. The apparent amplitude variation in the United States for a northern azimuth of approach is determined from an average of all non-diffracted data. The effect of signal amplification for a receiver situated on low velocity sediments is considered. A qualitative measure is made of the importance of scattering in the amplitude variation utilizing the horizontal components of the P waves. An explanation of the amplitude variation in terms of attenuation differences is considered. Finally, to provide an indication of the azimuthally varying characteristics of the amplitude pattern, preliminary results are presented in a study of amplitude patterns in the United States using earthquakes in the Kuriles and South America.

Nature of the Study

The overall geometry of the stations of the WWSSN in the United States and the source regions of the nuclear explosions in the Soviet

Union are shown in Figure 1. The Russian events used in this study - listed in Table 1. - have been naturally divided into several groups based on test sites. All events between 1965 and early 1976 having a yield of 70 kt. or greater, are analyzed in the collection of data. The following abbreviations are used for the test sites: NNZ, northern Novaya Zemlya; SNZ, southern Novaya Zemlya; SEMI E, Semipalatinsk east; SEMI W, Semipalatinsk west; KAZ, Kazakh. NNZ and SNZ are situated roughly 300 km apart; SEMI E and SEMI W lie about 70 km apart. The approximate distances in degrees delta between the United States stations and the source regions are listed in Table 2. The United States is situated in the range 60° to 80° delta from the Novaya Zemlya sites, NNZ and SNZ. P waves at these ranges bottom in the smooth lower mantle and should be relatively free of propagational path effects. The KAZ and SEMI source regions lie in the 80° to 100° range, and thus the southern United States stations will experience amplitude effects due to diffraction at the core-mantle boundary.

There are several advantages in using nuclear explosions rather than earthquakes in an amplitude study of this nature. Earthquake amplitudes contain source radiation and directivity effects which must be corrected before amplitude effects due to the path or receiver can be ascertained. In contrast the explosion source is theoretically spherically symmetric, and thus is naturally free of this complication. Event locations within the Soviet test sites vary on the order of tens of kilometers. The several events in each site allow the amplitude measurement to be repeated, and afford us an opportunity to view the stability of the measurement. Short period vertical P waves are

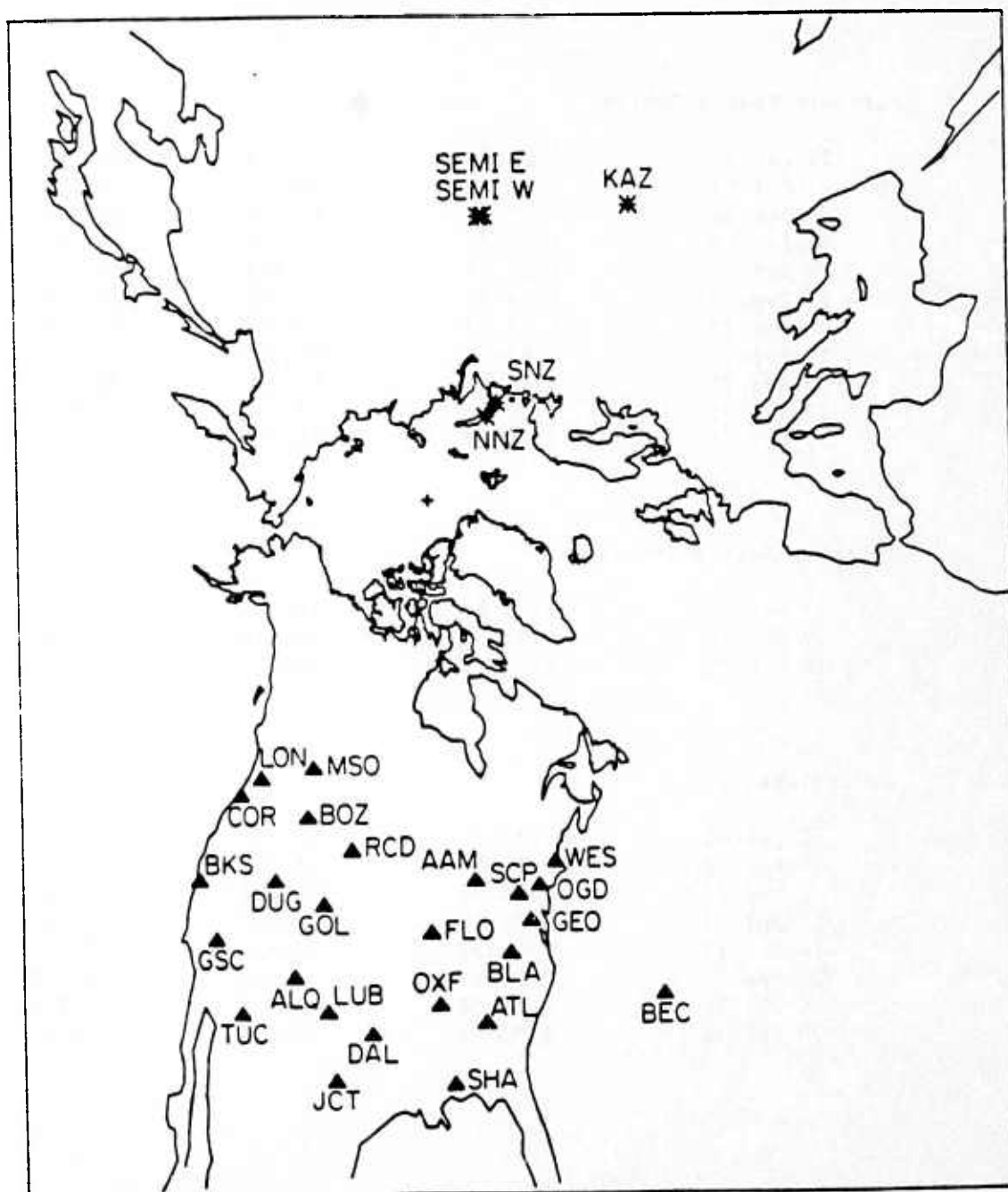


Figure 1 A gnomonic projection (all great circles are straight lines) showing the source regions in the Soviet Union and the WWSSN stations in the United States.

TABLE 1
EXPLOSION DATA SET*

Northern Novaya Zemlya

27 Oct 66	5:57:58	73.44N	54.75E
21 Oct 67	4:59:58	73.37N	54.81E
7 Nov 68	10:02:05	73.40N	54.86E
14 Oct 69	7:00:06	73.40N	54.81E
14 Oct 70	6:02:57	73.31N	55.15E
27 Sept 71	5:59:55	73.39N	55.10E
28 Aug 72	5:59:57	73.34N	55.08E
12 Sept 73	6:59:54	73.30N	55.16E
29 Aug 74	9:59:56	73.37N	55.09E
23 Aug 75	8:59:58	73.37N	54.64E
21 Oct 75	11:59:57	73.35N	55.08E

Southern Novaya Zemlya

27 Sept 73	6:59:58	70.76N	53.87E
2 Nov 74	4:59:57	70.82N	54.06E
18 Oct 75	8:59:56	70.84N	53.69E

Seripalatinsk East

15 Jan 65	5:59:59	49.89N	78.97E
30 Nov 69	3:32:57	49.92N	79.00E
2 Nov 72	1:26:58	49.91N	78.84E
23 Jul 73	1:22:58	49.99N	78.85E
14 Dec 73	7:46:57	50.04N	79.01E
31 May 74	3:26:57	49.95N	78.84E
4 Jul 76	2:56:58	49.91N	78.95E
23 Nov 76	5:03:00	50.00N	79.00E

TABLE 1

(Continued)

Semipalatinsk West

19 Oct 66	3:57:58	49.75N	78.03E
20 Apr 67	4:07:58	49.74N	78.12E
17 Oct 67	5:03:58	49.82N	78.10E
29 Sept 68	3:42:58	49.77N	78.19E
28 Jun 70	1:57:58	49.83N	78.25E
22 Mar 71	4:32:58	49.74N	78.18E
25 Apr 71	3:32:58	49.82N	78.09E
30 Dec 71	6:20:58	49.75N	78.13E
20 Feb 75	5:32:58	49.82N	78.08E

Yazakh

6 Dec 79	7:02:57	43.83N	54.78E
12 Dec 70	7:00:57	43.85N	54.77E
23 Dec 70	7:00:57	43.83N	54.85E

* Locations and origin times from Dahlman and Israelson (1977)

TABLE 2

APPROXIMATE DISTANCES BETWEEN U.S. STATIONS
AND RUSSIAN TEST SITES
(degrees delta)

	<u>NNZ</u>	<u>SNZ</u>	<u>SEMI</u>	<u>KAZ</u>
AAM	61.2	63.2	86.9	86.5
ALQ	71.2	73.6	95.4	99.8
ATL	69.9	71.9	95.7	94.5
BEC	67.0	68.4	91.9	86.2
PKS	69.0	71.6	90.6	98.6
BLA	65.6	67.5	91.3	89.7
BOZ	60.9	63.5	84.4	90.1
COP	62.3	64.9	83.9	91.9
DAL	72.3	75.6	97.5	99.4
DUG	66.4	68.9	89.6	95.6
FLO	65.6	67.8	91.1	91.9
GFO	63.3	65.2	89.1	86.8
COL	66.4	68.8	90.7	94.9
CSC	71.4	74.0	94.0	100.9
JCT	75.0	77.3	99.9	102.5
LON	60.1	62.7	82.0	89.7
LUB	72.1	74.5	96.8	100.1
MSO	59.8	62.4	83.0	89.2
OCD	69.7	62.5	86.5	84.0
OXF	70.0	71.8	95.3	95.5
RCD	61.8	64.3	86.3	90.9
SCP	61.6	63.5	87.4	85.6
SHA	73.2	75.3	98.9	98.4
TUC	74.1	76.6	97.6	103.1
WES	58.8	60.6	84.5	81.5

shown in Figure 2 from four event source regions. The overall consistency and similarity of the initial waveforms from station to station and event to event represent another natural advantage of the explosion data set. The initial upswing and following larger downswing are characteristic of the explosion waveform. The coherency of waveforms diminishes after this initial portion due to different crustal effects at different stations. The relative amplitude of each station within each event is indicated. There is no apparent correlation between frequency content and amplitude in this data set. Synthetic waveforms computed for the von Seggern and Blandford(1972) theoretical explosion source are shown for comparison. The A and B swings are primarily controlled by the direct P wave, whereas the C peak is controlled by pP. The somewhat longer period waveform of the Kazakh event indicates a deeper depth of burial and slower rise time of the source.

The amplitude measure used throughout this study is indicated by the synthetic waveforms: the amplitude of first peak to first trough, or AB amplitude. This particular measure was chosen for several reasons. It may be consistently and unambiguously read at each station for all events. It is the portion of the waveform least contaminated by source effects due to variations in pP and source structure, and from the receiver effects of different station crustal structures. The AB amplitude measure also minimizes baseline problems in the presence of noise.

The quality of the explosion amplitude data is quite good. The amplitudes may read with a precision of better than 10%, except at low gain stations for smaller events where the error may be up to 30%.

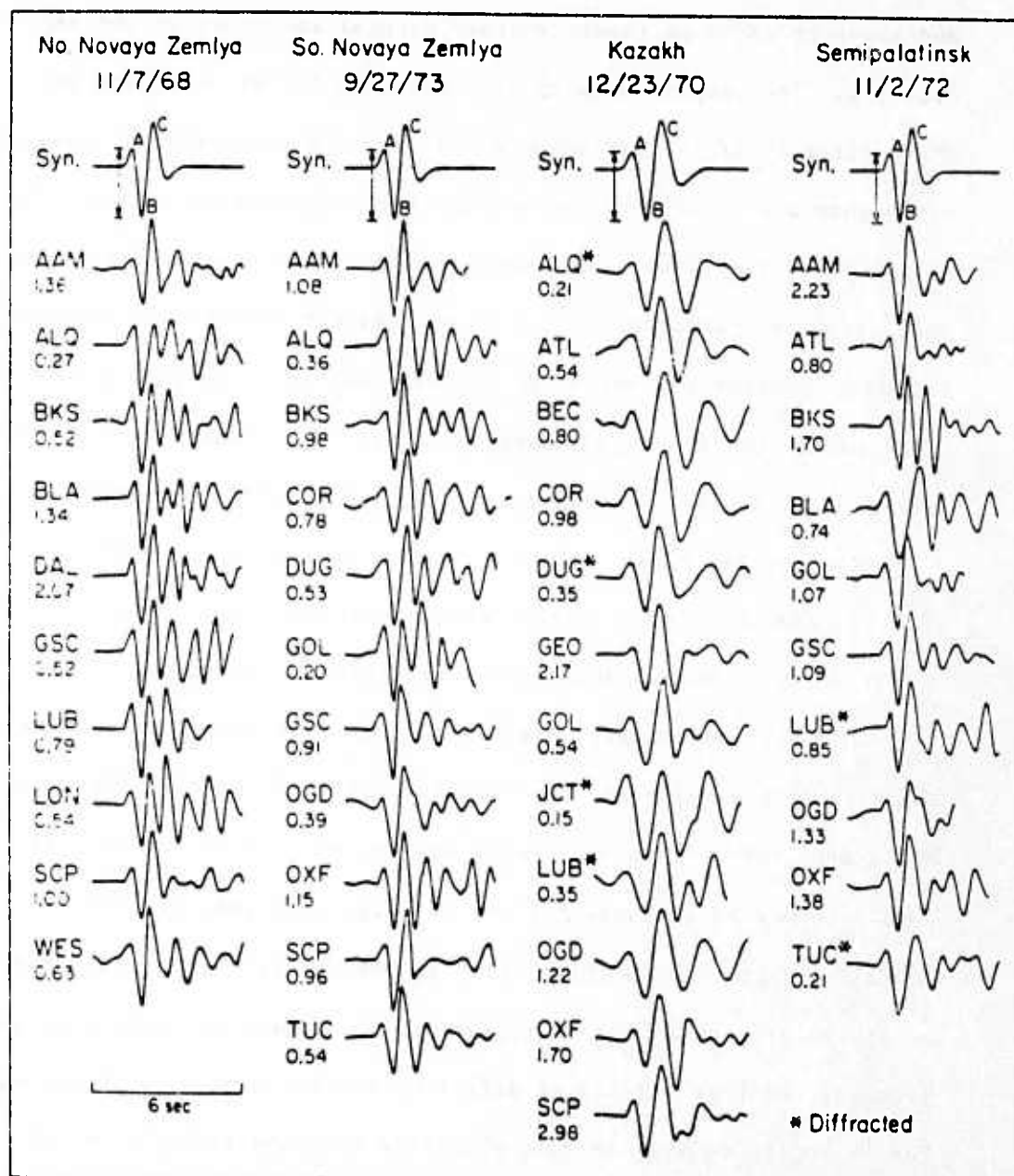


Figure 2 Short period P waves recorded at WWSSN stations across the United States from four Russian nuclear events. The numbers beneath the station codes indicate the true relative amplitudes as defined by the AB measure indicated on the comparison synthetic waveform. Stations beyond 95.5° are noted as diffracted.

The data are measured, corrected to a common station gain of 100 k, and grouped by source region. In each source region we have amplitude observations o_{ij} for i events and j stations. While the relative amplitude relationship from station to station is similar from event to event, the absolute amplitudes of each event are different. A simple normalization procedure would be to choose one station as a reference, and then for each event divide all the station amplitude measurements by the amplitude at the reference station. In this way the events are normalized to a common scale but the reference station is always unity. However, this method forces whatever scatter that is characteristic of the reference station upon the rest of the stations. To improve this situation the following normalization procedure was used. From the i events choose a reference event k , to which the other events are to be scaled in a least squares sense. Scale factors α_i are determined such that the least squares error is minimized for each event $i \neq k$:

$$\min \sum_j | \alpha_i o_{ij} - o_{kj} |^2 \quad (1)$$

Let $\bar{o}_k$ be the average amplitude of the master event k :

$$\bar{o}_k = \frac{1}{j} \sum_j o_{kj} \quad (2)$$

The total error for all events i in the source region is then

$$\frac{1}{j} \sum_i \sum_j | \alpha_i o_{ij} - \bar{o}_k |^2 \quad (3)$$

We then iterate on the process, letting each event i in a source region be the master event k . The scale factors α_i in each source are chosen for the event k which minimizes the total error in (3) the best.

Data

The amplitude data for the five source regions are plotted in Figures 3 through 6. The ordinate of each plot is a log scale indicating relative amplitude. The stations of the WSSN used in this study are listed in a west to east arrangement with respect to station location in the United States viewed from the Semipalatinsk test sites. Detailed intercomparisons of the source regions will be considered in the next section but some observations may be noted at this time. The results for NNZ are shown in Figure 3. The overall scatter at each individual station is quite low; less than a factor of 1.5 about the mean. The lowest stations are ALQ and GOL, followed by DUG and BEC. The amplitudes in the west are somewhat lower than those in the east. There is an amplitude differential of nearly a factor of 5 between GOL and ALQ and the higher amplitude stations, AAM, DAL and SHA. The data for SNZ plotted in Figure 4 are similar in overall pattern to NNZ, but different in detail. ALQ and GOL are still low, but OGD and WES are now among the lower values. The stations in the far west for SNZ have amplitudes comparable to the high values in the central United States. The relationship of BLA with other eastern stations is different for NNZ and SNZ. Figure 5 plots the data for both Semipalatinsk test sites. There appear to be no systematic differences between the SEMI E and SEMI W sites, in contrast to NNZ and SNZ. Figure 5 is considerably more complicated in appearance

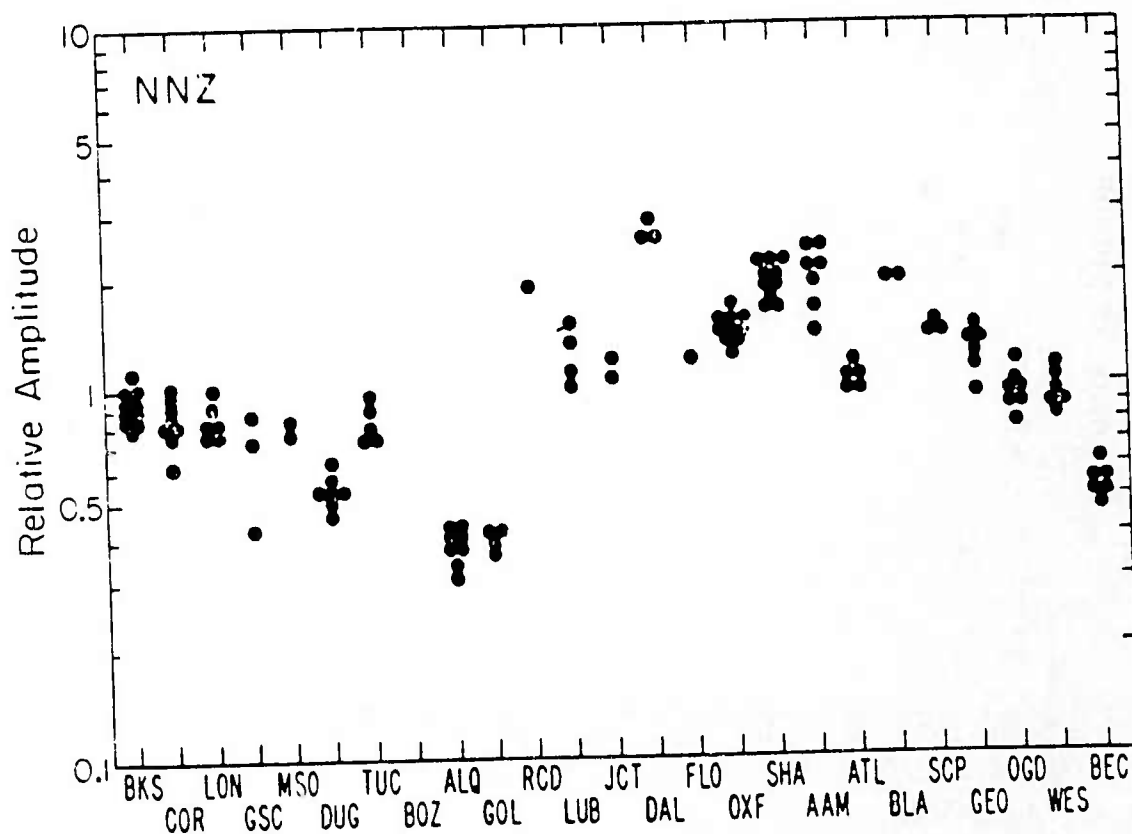


Figure 3 Relative amplitudes of short period P waves observed at WWSSN stations across the United States from eleven nuclear explosions at the northern Novaya Zemlya site.

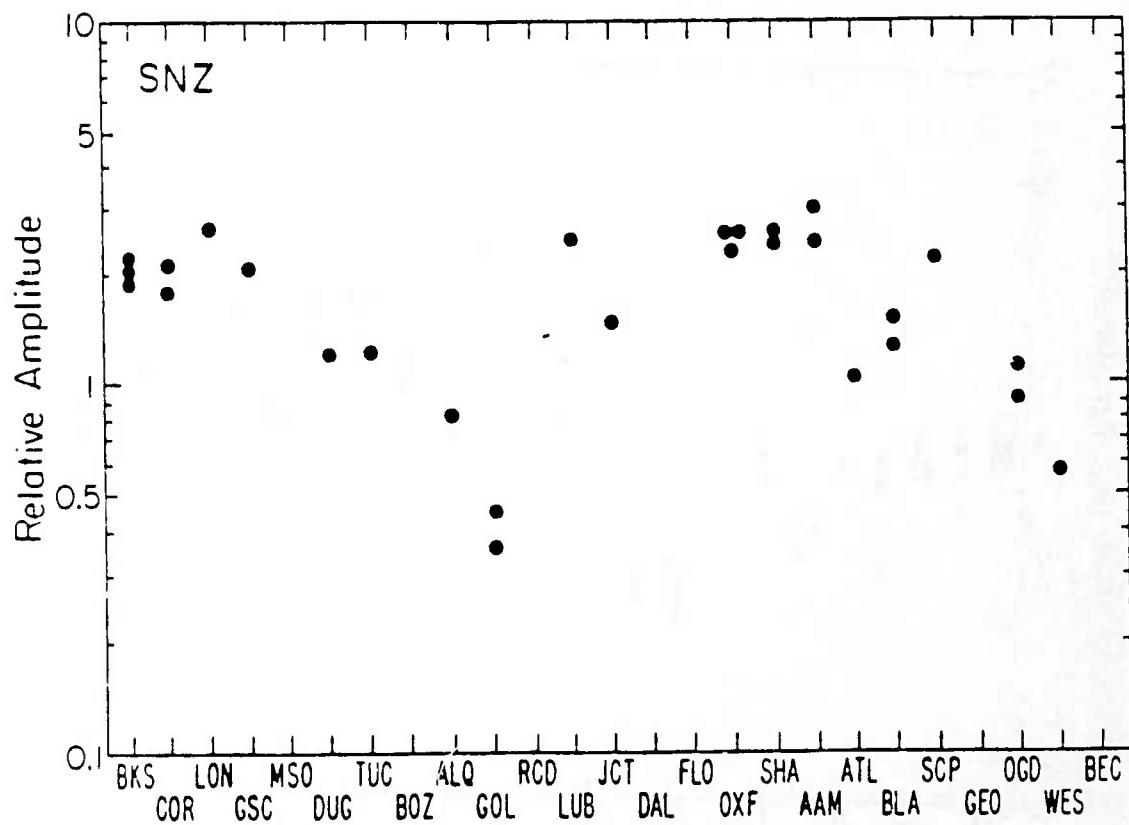


Figure 4 Relative amplitudes of short period P waves observed at WSSN stations across the United States from three nuclear explosions at the southern Novaya Zemlya site.

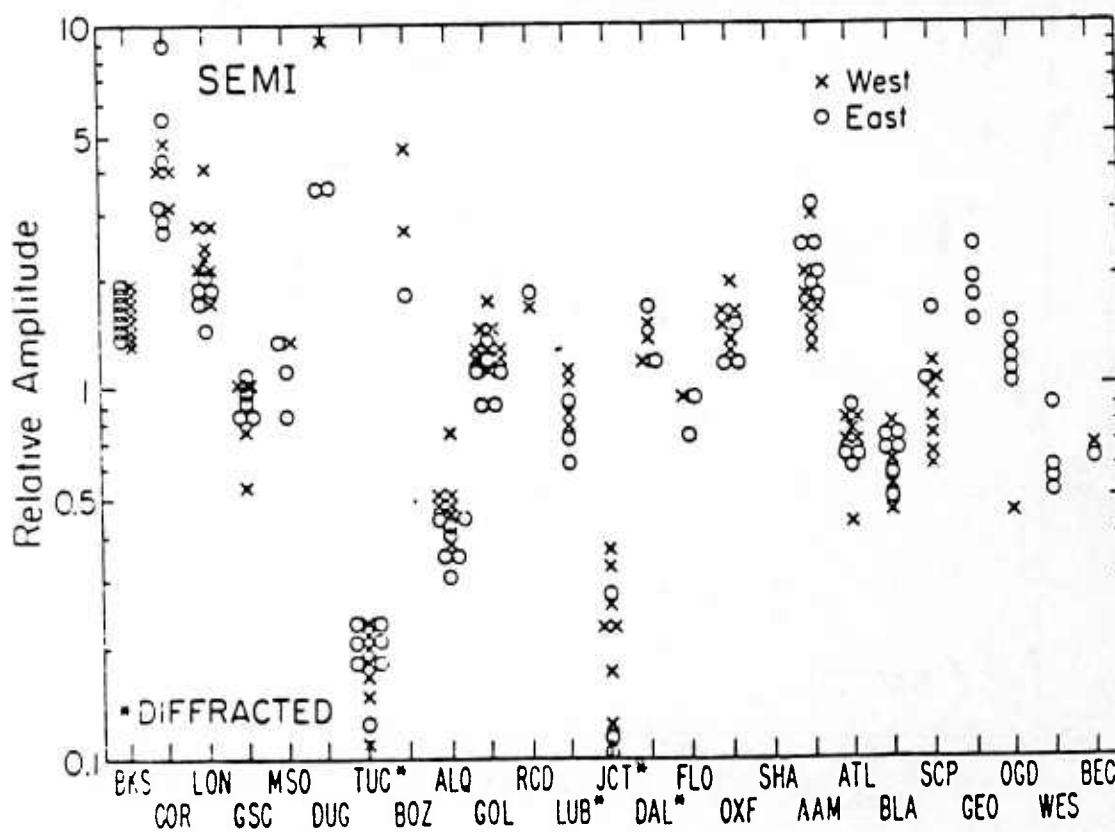


Figure 5 Relative amplitudes of short period P waves observed at WWSSN stations across the United States from seventeen nuclear explosions at the east and west Semipalatinsk test sites. Stations beyond 95.5° are noted as diffracted.

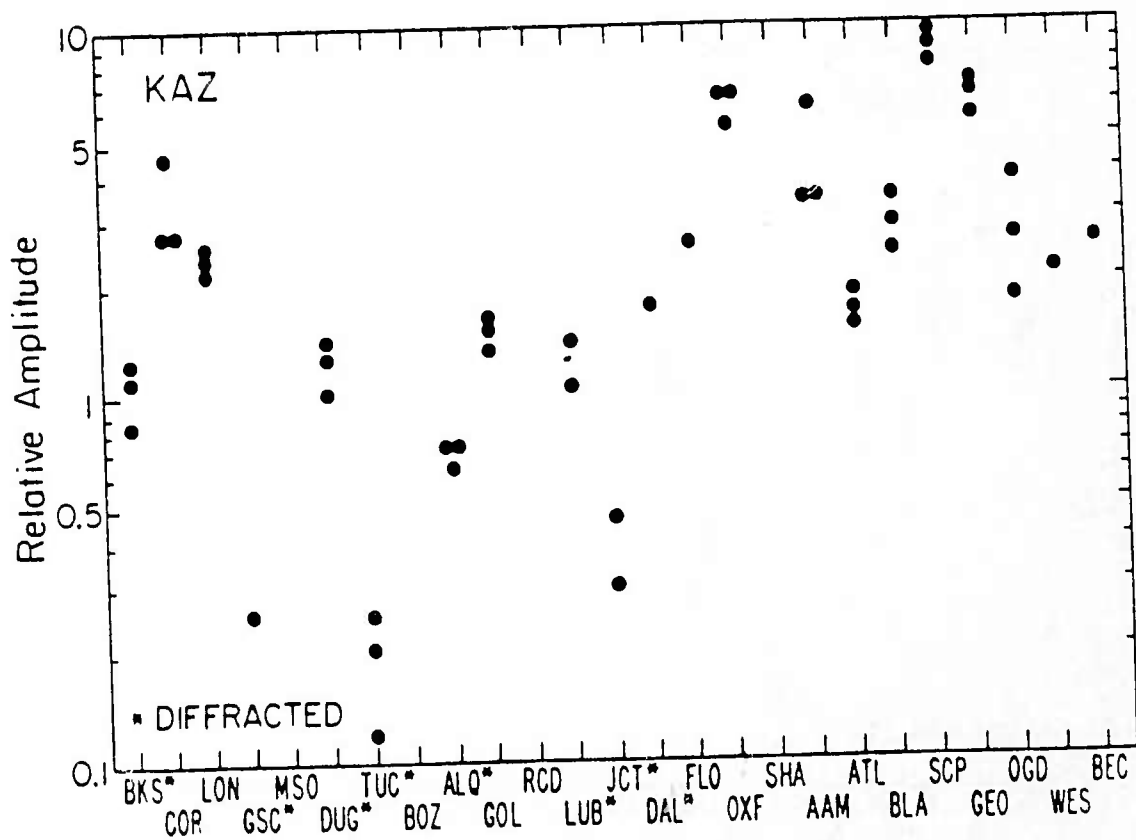


Figure 6 Relative amplitudes of short period P waves observed at WWSSN stations across the United States from three nuclear explosions at the Kazakh site. Stations beyond 95.5° are noted as diffracted.

than the plots for the Novaya Zemlya source regions. The greater scatter at the individual stations may come in part from including stations showing amplitude diffraction effects - TUC, DAL, LUB, JCT - in the data normalization procedure, and in part from larger reading errors in measuring the amplitudes of the smaller yield Semipalatinsk explosions. Stations COR, DUG, GOL show greater relative amplitudes from the Semipalatinsk sources than were observed for NNZ and SNZ. The three events for the Kazakh site are plotted in Figure 6. Low individual station scatter is observed for this site. A number of stations - BKS, GSC, DUG, TUC, ALQ, LUB, JCT, and DAL - show relative deamplification due to diffraction at the core-mantle boundary.

Source Comparisons

The amplitude patterns in Figures 3 through 6 represent a combination of source, path, and receiver effects. We may remove the effect of the receiver to look at the effects due to the source and the path if we assume the P waves emerging at a station in the United States from the different test sites experience a common receiver effect. As the azimuths and angles of incidence from the five source regions vary by only several degrees, this is a plausible assumption. The mean and standard error of the mean are calculated from the event data at each station for the different source regions and are used in further manipulation of the data. Dividing the station amplitude values for one test site by those of another cancels the common receiver and smaller geometric spreading effects. The resultant ratios may be plotted in the same fashion as Figures 3 through 6 and represent the relative source-path effects between the two source

regions. In the case of the two Novaya Zemlya sites, NNZ and SNZ, as the ray paths are similar and bottom in the rather smooth lower mantle, the ratio pattern represents primarily relative source effects. This ratio pattern is plotted in Figure 7 from the data in Figures 3 and 4. The figure shows an unmistakable trend of high in the west to low in the east, roughly a factor of 3 variation. As the ratios only indicate a relative pattern, the amplitude variation could originate at NNZ or SNZ, or a combination of both. To attempt to clarify this point, intercomparisons with the SEMI and KAZ sites are tried.

To utilize SEMI and KAZ for source and receiver consideration, stations suffering amplitude diminution due to diffraction at the core-mantle boundary must be separated. The diffraction effect is considered in detail in a separate paper (Ruff and Butler, in preparation) but may be briefly outlined. The SEMI and KAZ data are ratioed to the NNZ and SNZ data. The ratios are then plotted amplitude versus distance. The ratios hover about unity until the distances from SEMI and KAZ reach about 95.5° delta. Past this distance amplitudes systematically decline due to diffraction. Stations showing this amplitude diminution have been indicated in Figures 5 and 6 with an asterisk. These stations are largely deleted from further consideration in this paper. Stations less than 95.5° shall be assumed to be free of diffraction amplitude effects.

An interesting contrast is observed between the two sites at Semipalatinsk and the two sites on Novaya Zemlya. The scatter observed at a station for the different events at the Novaya Zemlya sites in Figures 3 and 4 is low, and is somewhat greater for Semipalatinsk. This

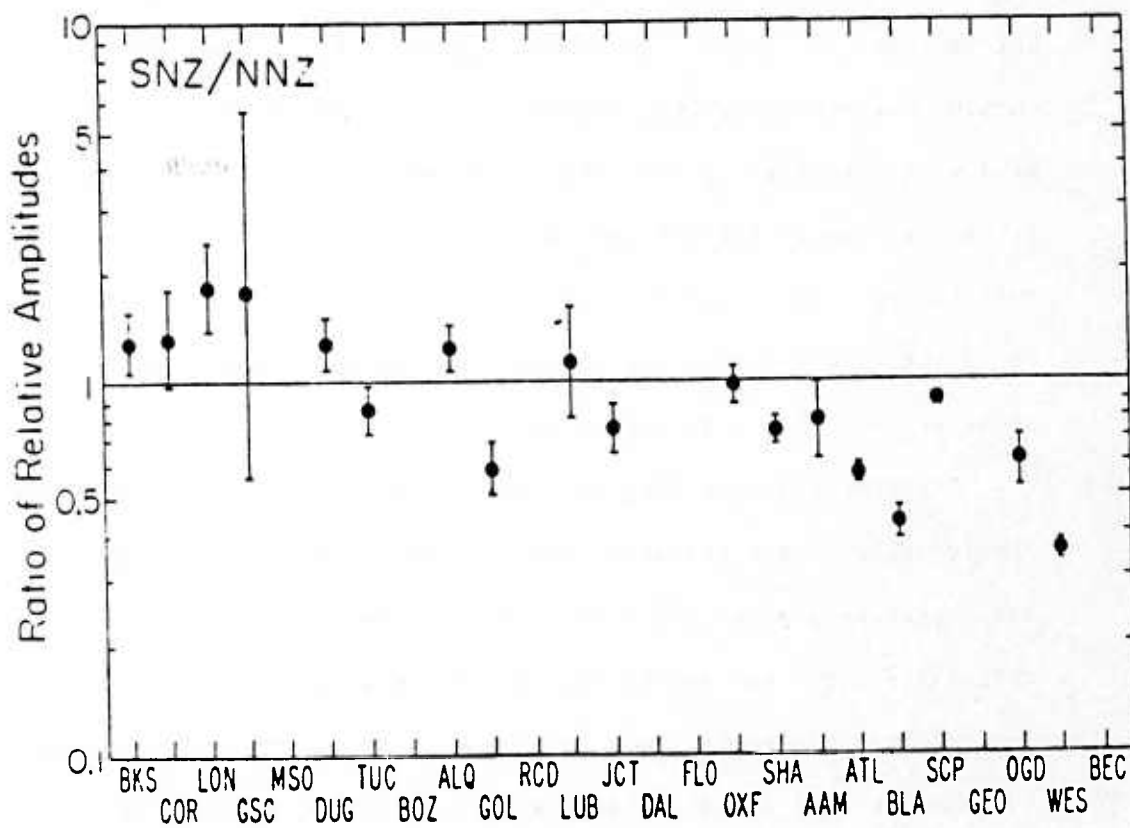


Figure 7 Ratio of the amplitude pattern for the southern Novaya Zemlya site to the amplitude pattern for the northern Novaya Zemlya site.

increased scatter in the SEMI values may be partly due to including the diffraction data in the normalization. More interesting, however, is the overall agreement of SEMI E with SEMI W, in marked contrast to the differences observed in the ratio plot of the SNZ to NNZ. The two SEMI sites lie about 70 km apart, while SNZ and NNZ are roughly 300 km apart. A ratio comparison among the combined SEMI data and NNZ and SNZ is shown in Figure 8. If we ignore COR, DUG, and GOL momentarily, the remaining SEMI stations fit the NNZ and SNZ data equally well with a scatter of less than a factor of two. This also appears to be the case in the KAZ data in Figure 9, although the control is more sparse. The scatter of the KAZ values in relation to NNZ and SNZ is of the same order as was observed in Figure 8 for Semipalatinsk, somewhat less than a factor of two.

The SEMI stations COR, DUG, GOL and the NNZ value for BLA show a considerably larger deviation than the other stations in Figure 8 and impetune a momentary digression to consider these anomalous values further. For additional comparison a ratio plot of the KAZ data to NNZ and SNZ is shown in Figure 9. In this plot neither COR nor GOL show large deviations, and quite the contrary agree well the NNZ and SNZ values. DUG is included, though it may be somewhat diminished due to diffraction, to show that it does not stand greatly different than the NNZ or SNZ data. This suggests that the apparent amplification of COR, DUG, and GOL is a property of the SEMI data set. As this amplification is observed for only these three stations in both the SEMI E and SEMI W data, and as the two sites are situated 70 km apart, and ruling out coincidence as low probability, we may conclude that the apparent amplification is not due to the source regions,

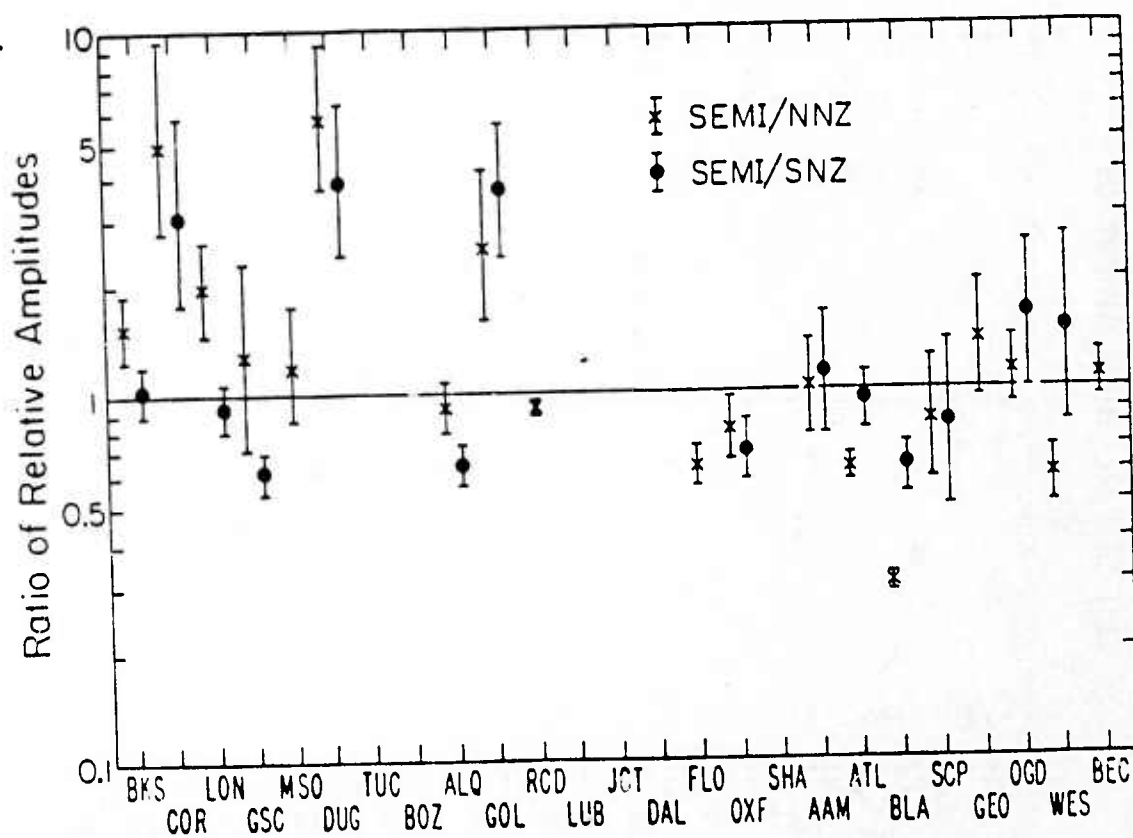


Figure 8 Ratios of the amplitude pattern for the combined east and west Semipaltinsk sites to the amplitude patterns for the northern and southern Novaya Zemlya sites. Diffracted data have been deleted.

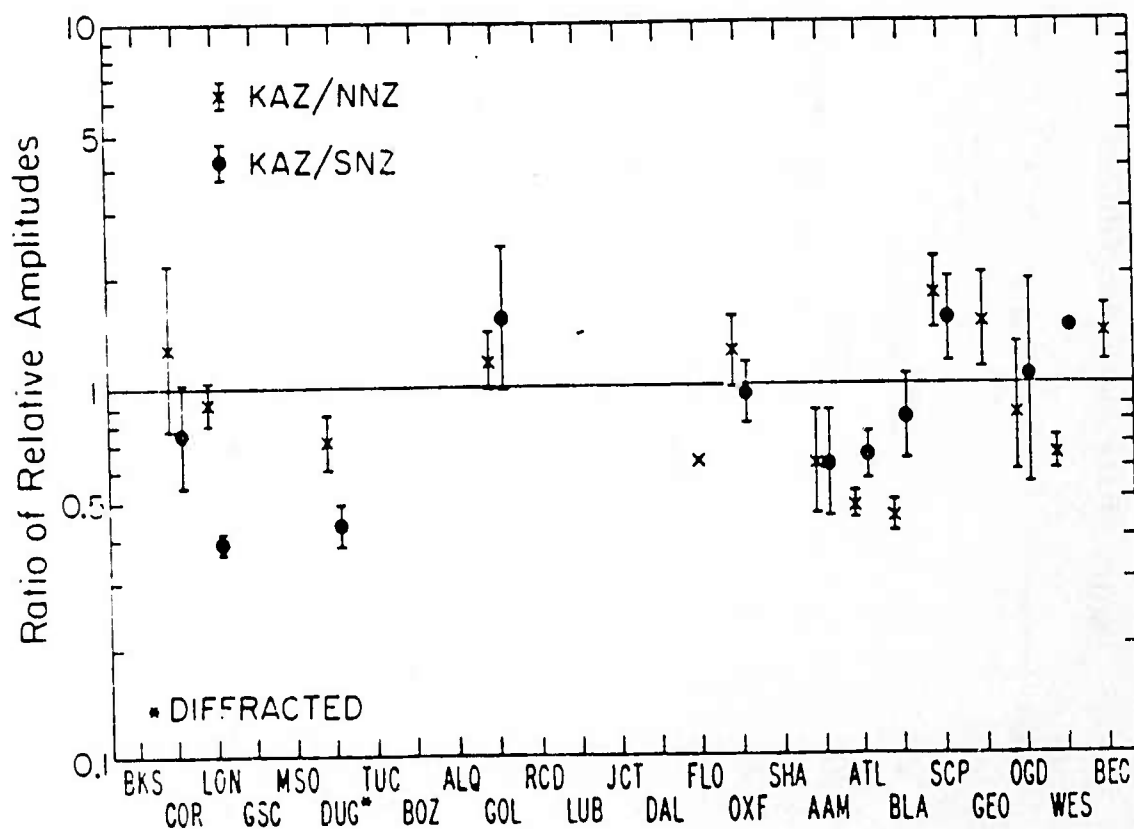


Figure 9 Ratios of the amplitude pattern for the Kazakh site to the amplitude patterns of the northern and southern Novaya Zemlya sites. Diffracted data - except DUG - have been deleted.

but rather rests in path or receiver effects. Other stations to be noted in Figure 9 are LON and BLA. LON has shifted from favoring SNZ at SEMI to favoring NNZ for the KAZ data. The amplitude value for BLA at NNZ is a factor of 2 or more greater than the values for BLA at the other test sites and most likely represents an anomaly in the NNZ source region.

Station COR at Corvallis, Oregon together with LON in Longview, Washington have been studied by Burdick and Langston (1977), Langston and Blum (1977), and Langston (1977, 1978) who analyzed long period body waveforms in the determination of receiver structure. Both COR and LON have been found to show receiver anomalies which may be partly attributable to dipping structure. A comparison of the P waveforms of COR and LON with the other WSSN stations in the United States for this explosion data set qualitatively reveals greater complexity for the vertical and horizontal components for both long and short period P waves at COR and LON. It seems likely then that the amplification at COR for the SEMI source regions, and possibly some of the scatter at LON, may be due to lateral heterogeneity at the receiver.

As DUC and COL are adjacent stations such that their raypaths occur close together, the apparent amplification at these two stations may be caused by a common velocity anomaly affecting both paths. Otherwise, we would have to appeal to separate receiver anomalies that affected both stations in a similar fashion from the same approach azimuth. The discussion of a common velocity anomaly is deferred to the companion study.

On the basis of the intercomparisons of the test sites, we may summarize our qualitative results. The two sites at Novaya Zemla show a relative amplitude trend of a factor of three across the United States in their respective amplitude patterns. This is in contrast to the two test sites at Semipalatinsk, where they are in good relative agreement. This amplitude anomaly associated with Novaya Zemlya does not appear to be a local function of either NNZ or SNZ, as the amplitude patterns for KAZ and SEM1 are intermediate between NNZ and SNZ. No information is available in the geophysical literature to help to resolve this issue. Geologically, the islands of Novaya Zemlya are a continuation of the Ural Mountains, and as such are part of a suture zone between ancient European and Asian plates. The complex geologic structure observed on geology maps of the Urals and Novaya Zemlya are circumstantially supportive, but offer no additional information to aid our interpretation.

Amplitude Variation at Receiver

In the previous section, amplitude data for the five test sites were inter-compared using a ratio technique which essentially eliminated common receiver effects. In this section the amplitude data from the five test sites are plotted together to look for systematics of the common receiver effect. Stations showing amplitude diffraction effects and the amplitude values of three anomalous stations for the Semipalatinsk data - COR, DUC and GOL - have not been included in the common plotting. Although some systematic differences among the various source regions were found in the previous sections we include the data from each site without any corrections, understanding that the source

region differences represent an inherent uncertainty in determining the effects common to the receiver. To plot the data from the different sites, the amplitude patterns in Figures 3 through 6 are characterized by the mean value of each station. The resultant amplitude patterns for the five sites are normalized in the same manner as was the event data at each site. Geometric spreading corrections have not been included as the effect is minor (about 15%) compared to the variations in the data.

The amplitude data from the five Russian nuclear test sites -- NNZ, SNZ, SEMI E, SEMI W, KAZ -- are plotted in Figure 10. The pattern represents apparent amplitude variations for the United States stations of the WWSSN for a northern azimuth of approach. This distinction--the northern azimuth of approach--is important as preliminary results (to be discussed in the next section) using earthquakes to view other azimuths indicate that the pattern is azimuthally dependent. However, in accepting this narrowing distinction of a northern azimuth, several observations are important. The amplitudes in the western United States from BKS to GOL are not a consistent factor of three lower than amplitudes in the eastern United States. Stations ALQ and GOL are a factor of 4 to 5 lower than the higher stations, but the other low stations in the west, DUG and TUC, have amplitude values not dissimilar from JCT, FLO, ATL, BLA, OGD, WES. Station BEC in Bermuda also has a value comparable to DUG and TUC. The values of BOZ are quite high, albeit they represent only three events from SEMI. The remaining stations in the west, BKS, COR, LON, GSC, and MSO, share a range of amplitude values with the higher stations in the east -- RCD, LUB, OXF, SHA, AAM, SCP, and GEO.

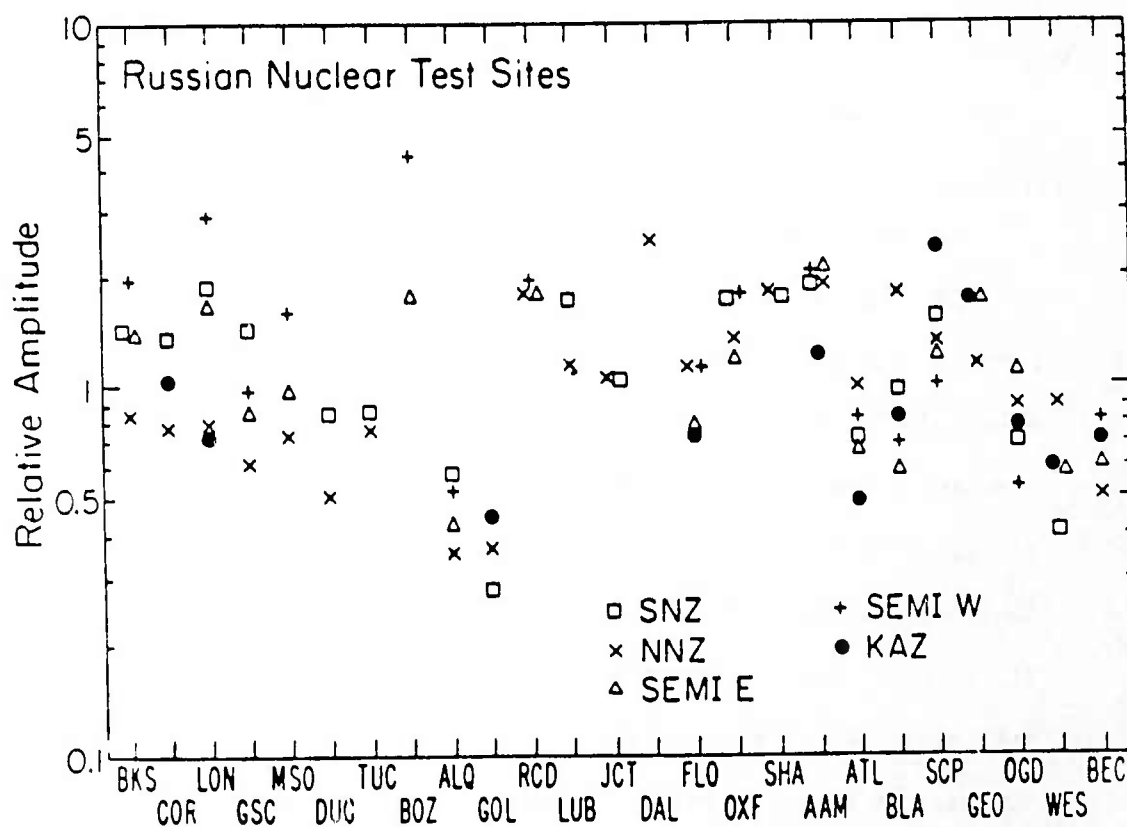


Figure 10 Combined plot of the amplitude data from the southern and northern Novaya Zemlya, east and west Semipalatinsk, and Kazakh sites for WWSSN stations across the United States. The mean of the amplitude data at each station for each site is plotted. Diffracted and anomalous (see text) data have been deleted.

Before comparisons of a more quantitative nature can be considered, effects due to differences in the geologic siting of the stations must be evaluated. Gutenberg (1956, 1957) and Borchardt (1970) have noted that stations situated on low velocity sediments are amplified relative to nearby stations situated on bedrock. Booth et al., (1974) invoked sediment amplification to explain high magnitude residuals for long period LRSM stations along the Gulf coast. This sediment amplification effect is illustrated in Figure 11. The receiver effect of a variety of different sediment structures overlying bedrock are compared to simple bedrock site (upper left) using Thomson-Haskell propagator matrices and a synthetic explosion waveform assuming Poisson's ratio for the shear velocity and a 0.2 g/cm^3 bedrock-sediment density contrast. Sediment thicknesses of 300 meters show appreciable amplification. Short period P waves propagating through near surface P wave velocities of 2 km/sec are amplified a factor 1.6 to 1.8 relative to bedrock; surface velocities of 3 km/sec amplify by 1.4 to 1.5. As stations in the central United States (from RCD to AAM) are situated on slower sedimentary materials in contrast to the hard rock siting of the western and eastern coastal stations, some amplitude corrections must be made. The handbook of the WSSN contains a short description of the local geology at each station: RCD, 2700 feet of shale, sandstone and some limestone; LUB, Pleistocene terrestrial deposits; JCT, Cretaceous Edwards limestone; FLO, 50 to 60 feet of recent clay overlying Missippian bedrock; OXF, 2100 feet of Cenozoic and Mesozoic sediments; SHA, sands and gravels of Plio-Pleistocene age underlain by clays and sand of Miocene age; AAM, 200 feet gravel, 800 feet of shale, 4800 feet of limestone. Composing a velocity section from the geologic descriptions is

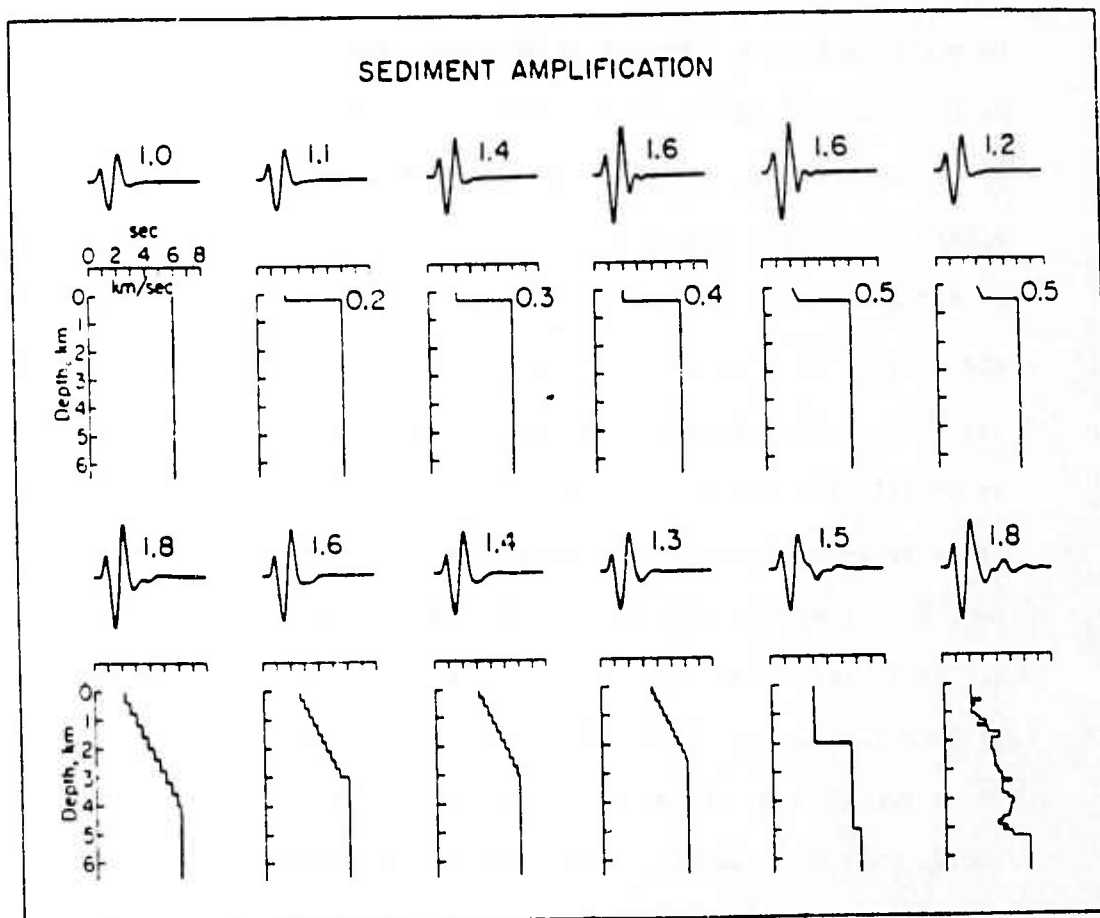


Figure 11 Amplification effects of shallow low velocity sediments upon crustal models. Amplitudes noted are relative to the bedrock model in upper left. Sediment thickness in kilometers is indicated in upper models.

somewhat arbitrary. Shallow lying sandstones and shales have compressional velocities from 1.4 to 3.3 km/sec. Limestone velocities are sensitive to the extent of crystallization and range between 1.7 and 6.1 km/sec. (Press, 1966). Seismic refraction surveys provide only general control as sediments are grouped in a single layer characterized by the highest velocity arrival. Well log velocity depth data provide the only accurate control, but such information was obtainable only from SHA. The lower right velocity-depth section in Figure 11 was simplified from a well log 25 km from SHA, and shows a factor of 1.8 amplification relative to bedrock. Given the lack of velocity control in the surface layers at RCD, LUB, JCT, DAL, and AAM, only an approximate sediment amplification correction of a factor of 1.4 to 1.8 may be estimated.

Returning to Figure 10 and noting the stations in the central United States to be corrected for sediment amplification, we observe that no pervasive factor-of-three-difference in amplitude is apparent between WSSN stations in west (left of RCD) and WSSN stations in the central and east (right of GCI). This factor of three variation between the east and west is based upon studies of earthquake magnitude variations for LRSM stations in the United States by Evernden and Clark (1970) and Booth et al. (1974).

The amplitude variations for receivers in the United States observed in Figure 10, whether or not the pattern is azimuthally dependent, may be attributed to a number of causes. Among the more likely are variations in attenuation properties, variations in scattering properties, and focusing or defocusing due to non-planar structure in the crust or uppermost mantle. While the focusing possibility is station specific

and cannot be modeled without more information, variations in attenuative and scattering properties may be tested for internal consistency with the data. A crude but simple test for variations of shallow crustal scattering can be made in the following manner. Let us consider the amplitude variations in the United States for the NNZ test site in Figure 3. For a simple scattering model we may assume the low amplitude stations are caused by energy scattered from the vertical component of motion. A large part of this scattered energy may manifest itself as large secondary arrivals on the horizontal components on the P wave train. This hypothesis was tested for the events at NNZ. Amplitudes of large secondary arrivals were measured from the north-south and east-west components of the P wave. These components are approximately radial and tangential, respectively, for geometry of source and receiver. The amplitudes are normalized by the values at OXF and plotted in Figure 12 in a fashion similar to the vertical data in Figure 3. If this simple scattering hypothesis was correct, we would expect the pattern of low and high amplitudes for the vertical arrivals in Figure 3 to be reversed for the horizontal secondary arrivals; that is, a high amplitude on the vertical would imply low amplitude on the horizontal and vice versa. This is clearly not the case in Figure 3. Thus, the amplitude variations observed for the vertical component are inconsistent with this simple scattering test. A more sophisticated test for scattering might be to compute the total power in the horizontal coda and compare this to a maximum amplitude, but this detailed effort is beyond the scope of the present investigation.

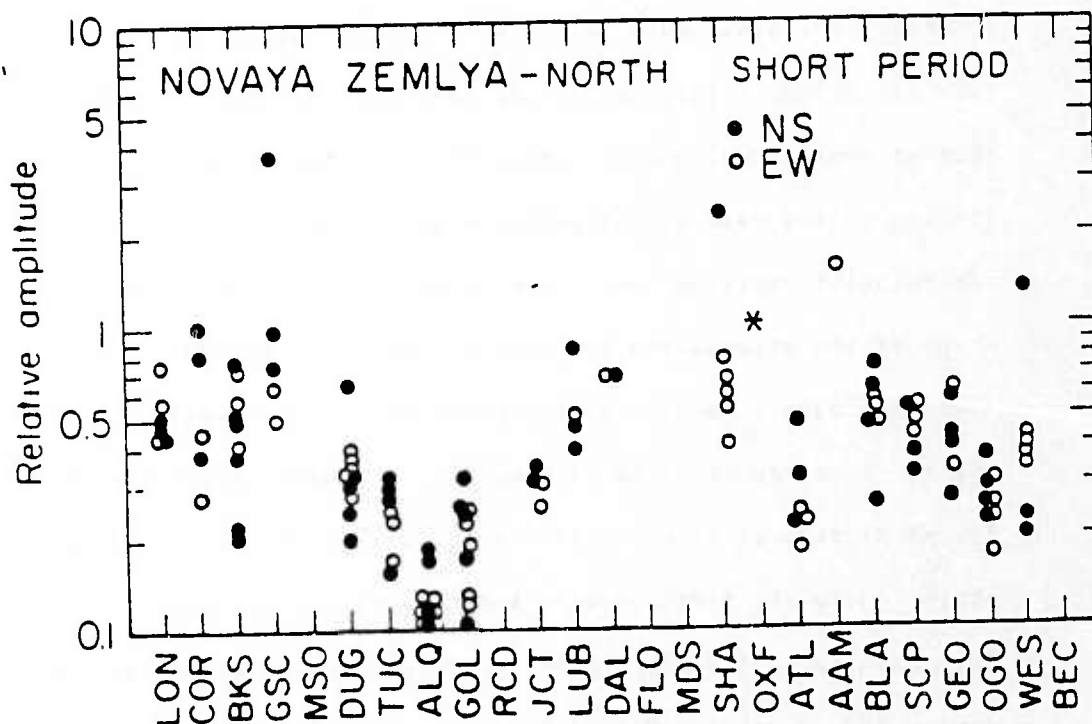


Figure 12 Relative amplitudes of horizontal P wave secondary arrivals (normalized to OXF) observed by WWSSN stations across the United States from eleven nuclear explosions at the northern Novaya Zemlya site.

A second possible explanation for the amplitude variations in Figure 10 is from differences in the attenuative properties along the ray paths to the stations. If we assume that attenuation is independent of frequency in the band of the WSSN short period instrument, we may characterize variations in attenuative properties by variations in $t^* = T/Q$, the ratio of travel time to path average Q . To verify the consistency of an attenuation explanation for the amplitude variation, a large amplitude waveform was carefully digitized and then attenuated using a Futterman (1962) attenuation operator. The results of the test are illustrated in Figure 13. The waveform and its relative amplitude are shown in the left column and the t^* and shape of the attenuation operator in the right column. A factor of five variation in amplitude is equivalent to a t^* variation of 0.5 units. It is somewhat surprising that the waveforms suffering different degrees of attenuation are quite similar. Restricting our attention to the first upswing and downswing - the part of the waveform used in this study and most free of crustal complications at the source and receiver - differences due to variation in attenuation are not appreciable resolvable to be of use as a criterion in accepting or rejecting attenuation as an explanation for the amplitude variations in Figure 10. Thus, the variations of amplitudes and lack of variation in the waveforms in the Russian explosion data set are consistent with an interpretation in terms of varying attenuative properties along the ray path to the receiver.

Preliminary Results for Earthquake Sources

The study of Russian nuclear explosions has provided an interesting pattern of amplitude variations in the United States ascribable to

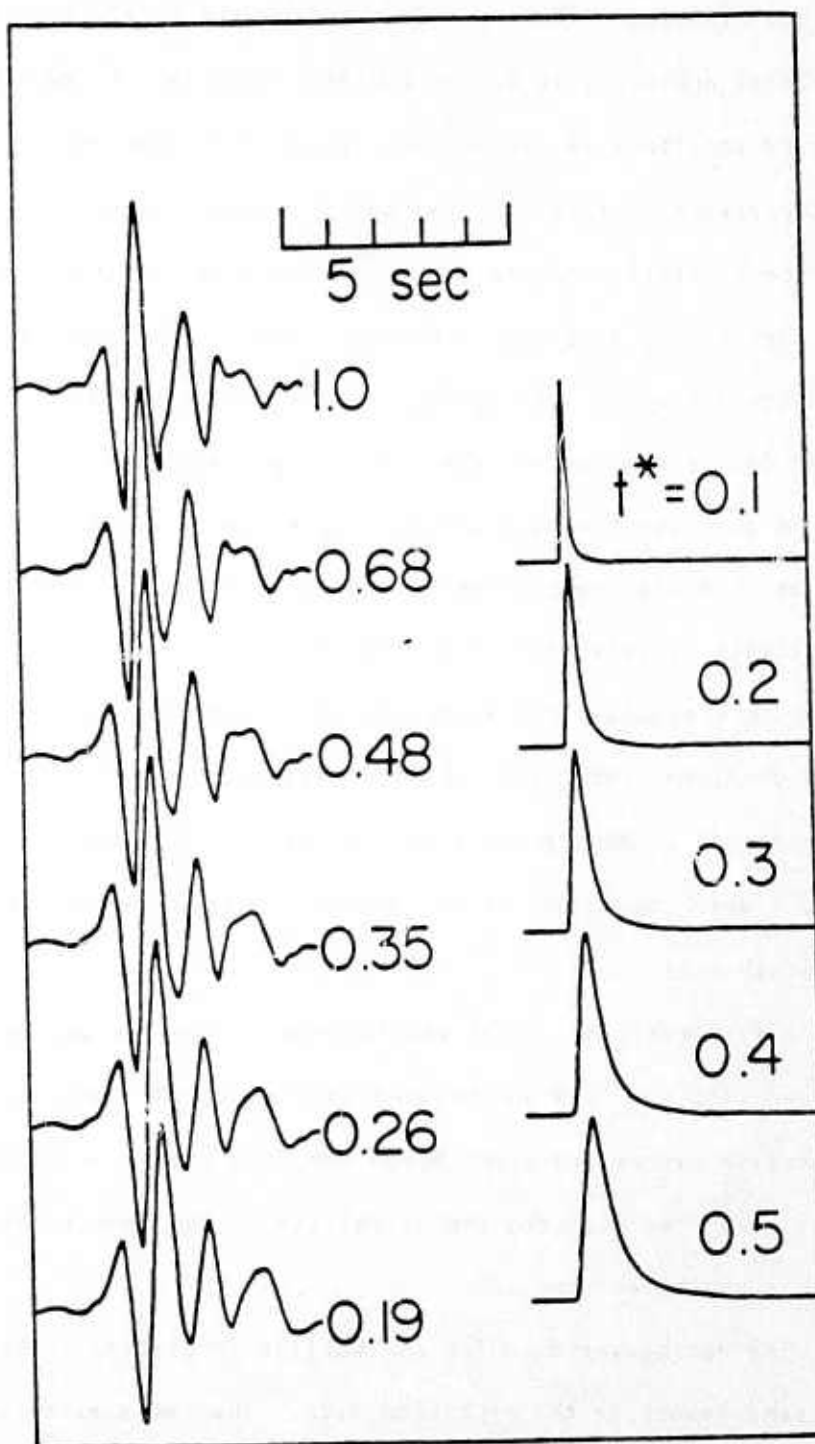


Figure 13 Result of numerical experiment in which a high amplitude short period P wave from a nuclear explosion is successively attenuated. The numbers following the traces indicate the signal amplitude normalized to the top trace at the left. The right-hand column shows the attenuation operators used to obtain each corresponding left-hand trace from the original record at top left.

receiver effects. However, as these data are solely from a northern azimuth of approach, it is particularly important to determine if the observed amplitude variations are azimuthally dependent. In view of the importance of this question and to temper premature speculation upon the interpretation of the explosion data, we shall present preliminary amplitude data from earthquakes in the Kuriles and South America which suggest that the amplitude variations in the United States are indeed azimuthally dependent.

To simulate the study of the explosion data set and to overcome as best as possible the problems of earthquake source radiation and directivity, a selection of earthquakes were sought to satisfy the following criteria: the short period P waveforms must be simple, short duration, explosion-like in character, and must be similar at all stations recording the event in the United States. Two source regions were considered in the search: events in the Kurile Islands and South America.

A systematic search of earthquakes in the two regions occurring between 1965 and 1968 and between September, 1976 and May 1977 yielded six Kurile events and eight South American events of acceptable nature. Amplitudes were measured and normalized in the same manner as the Russian explosion study.

The earthquake data for the Kuriles is plotted in Figure 14 in the same manner as the explosion data. The low scatter of the event data at the individual stations suggests that source radiation and directivity effects are minimal. The azimuth to the Kurile earthquakes roughly lies N 45° W of the Russian test sites, and a good deal

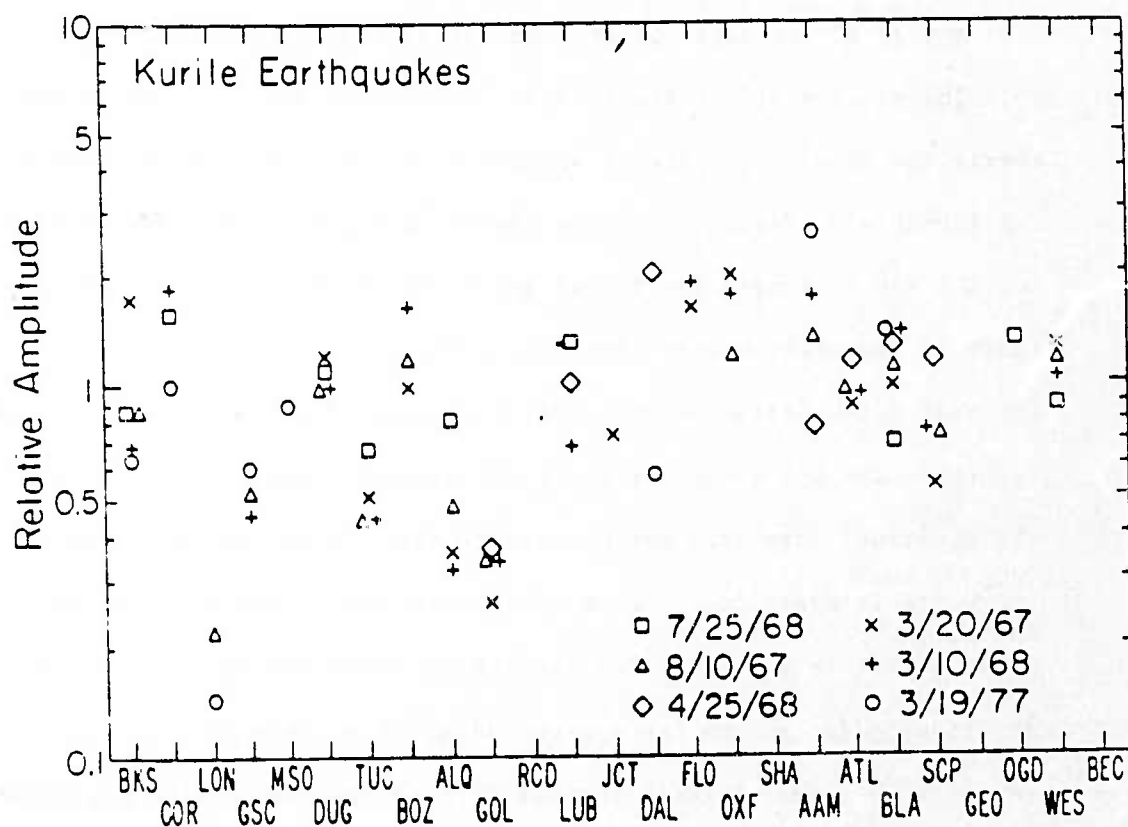


Figure 14 Relative amplitudes of short period P waves observed by WSSN stations in the United States for six earthquakes in Kurile Islands. The events were selected to minimize earthquake source radiation and directivity effects.

of consistency is seen comparing Figure 10 and Figure 14.

Figure 15 plots the average values of Russian explosion data together with the Kurile earthquake data for better comparison. Except for LON which was noted earlier to be associated with anomalous receiver effects, the Kurile data lie within a factor of two of the Russian explosion data.

A plot of the eight South American earthquakes is shown in Figure 16. The scatter at the stations is again quite low. The South American events lie nearly opposite in azimuth to the Russian explosion data, and considerable differences between the two regions may be observed. ALQ and GOL are no longer the lowest amplitude stations, but now fall in the range of the other western and east coast stations. Amplitudes in the central United States are considerably larger than values east or west, but there are not enough data yet to preclude anomalous source effects. The amplitude data from earthquakes in the Kuriles and South America are quite interesting. The waveform criteria in selecting events yield amplitude data with low internal scatter and appear to minimize amplitude effects from the source radiation and directivity. The relative agreement between the Kurile data and the Russian explosion data indicates that the amplitude variation pattern in the United States is stable with moderate changes of azimuth. However, differences between the South American earthquakes and the Russian explosion data suggest that the total azimuthal variation is quite pronounced at some stations. Further discussion of azimuthal differences in the amplitude variations of WWSSN stations in the United States must await a more detailed study. However, the preliminary results presented herein suggest considerable promise in the ability to approach the azimuthally dependent amplitude variation problem.

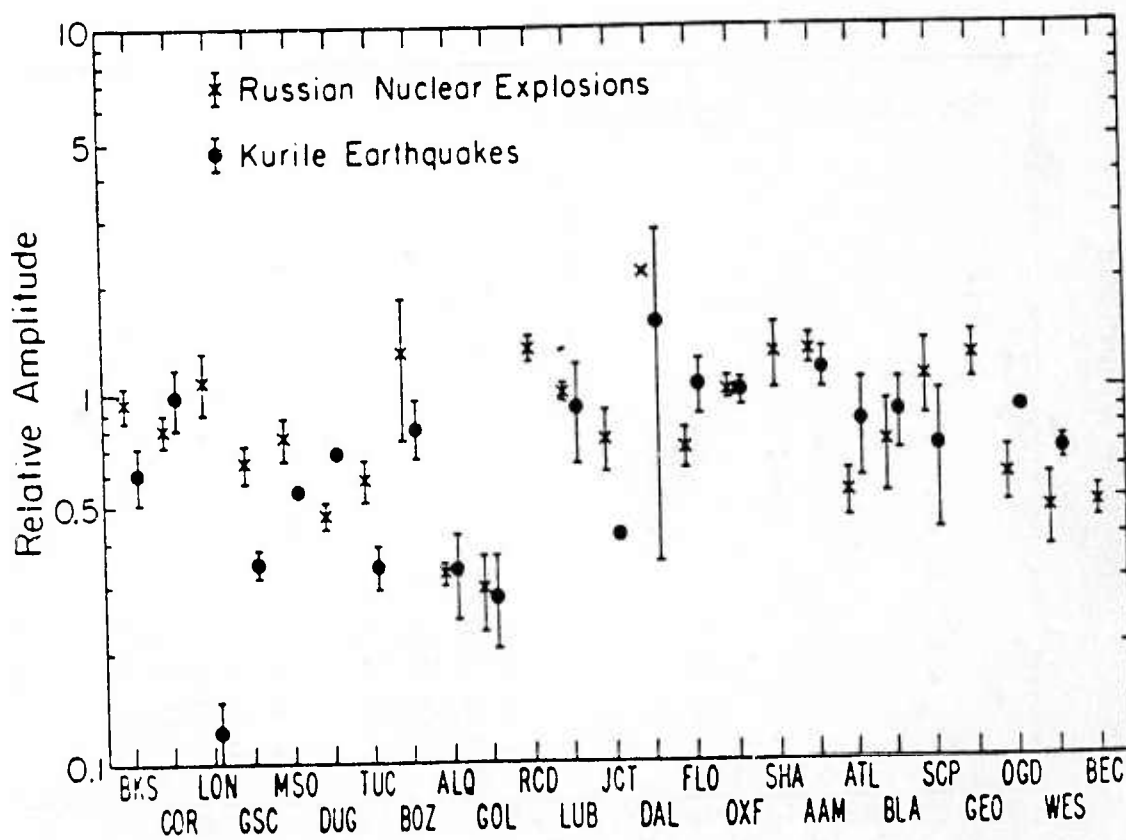


Figure 15 Comparison of the amplitude pattern from the combined Russian explosion data with the amplitude pattern from the Kurile Island earthquakes.

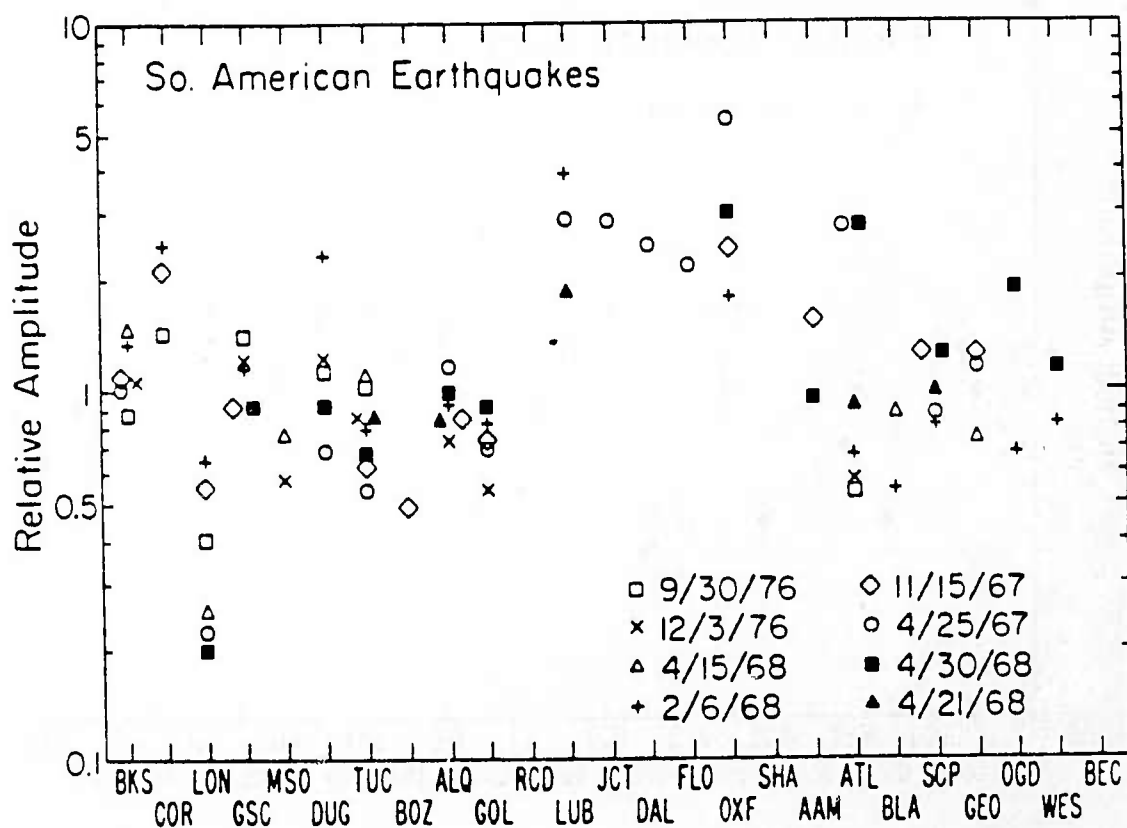


Figure 16 Relative amplitudes of short period P waves observed by WWSSN stations in the United States for eight earthquakes in South America. The events were selected to minimize earthquake source radiation and directivity effects.

SUMMARY

The study presented is an analysis of the short period amplitude behavior of WWSSN stations in the United States recording underground nuclear explosions in the Soviet Union. We have found that there is a well-defined relative amplitude pattern for any one Russian test site, and additionally, that the relative amplitude patterns observed at the U.S. stations are quite similar for the various test sites. This is an indication that the relative differences among the U.S. stations are primarily due to effects near the receiver. These initial results concerning the coherence of short period amplitudes and waveforms encourage a similar analysis be undertaken for all WWSSN stations.

This study may be considered a first step toward understanding the amplitudes and waveforms of short period seismograms. The observed short period seismograms are coherent from station to station for approximately 2 seconds after the initial onset. While the relationship between relative peak heights is maintained, the absolute amplitudes can vary by nearly an order of magnitude. Stability with regard to source region seems well demonstrated. There is an excellent agreement between the amplitude patterns of Semipalatinsk east and west, indicating a source coherence over 70 km. However, a systematic source azimuthal difference exists between southern and northern Novaya Zemlya, a distance of 300 km. For the purpose of defining an overall amplitude pattern from the northern approach, it is encouraging to note that the values for Semipalatinsk and Kazakhlie between the values for southern and northern Novaya Zemlya.

With regard to receiver effects, although a relative amplitude pattern is well-defined for the Russian test sites, this pattern cannot be simply interpreted in terms of attenuation, scattering, or other mechanisms until the azimuthally varying characteristics of the relative amplitude pattern can be determined. Preliminary amplitude data presented herein from earthquakes in the Kurile islands and South America suggest that significant azimuthal variations occur both at individual stations and for regional groupings of stations. Further work in determining the azimuthally dependent amplitude characteristics of the WWSSN stations in the United States may yield estimates of the localization of the areas of low and high amplitudes.

In conclusion, the principle results of this study are: (1) a demonstration of amplitude and waveform coherence of short period data from Russian nuclear explosions; (2) a preliminary result concerning the azimuthal variation in the effective radiation pattern from the Russian test sites; (3) the determination of a relative amplitude pattern across the United States, though noting the azimuthal dependence of this pattern. Additionally, we have sought possible mechanisms for the amplitude variations and find that amplification due to low velocity surface sediments and diminution caused by intrinsic attenuation are viable processes in explaining the variations.

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