Modeling Near-Field and Teleseismic Observations From the Amchitka Test Site

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During the Amchitka testing program, recordings were made of the near-field ground velocity for events MILROW and CANNIKIN. These records exhibited a P arrival, a feature associated with near-field pP and a developing Rayleigh wave. In this investigation, simple models were found which match the complete seismograms. They simultaneously predict the amplitude, wave shape, timing, and periodicity of both the body waves and the surface waves. Models for both the Amchitka crustal structure and the source time functions were developed. The crustal model is similar to previous models in its average properties, so travel time constraints are still satisfied. Small modifications have been made to match the waveform data. The events on Amchitka can apparently be represented as isotropic point sources with the types of simple source time histories predicted by classical yield scaling laws. A source model for event LONGSHOT was developed on the basis of the yield scaling laws since no near-field records are available for it. The source models were tested by comparing the teleseismic body waveforms they predict with observations. Short-period P waves, long-period P waves, and long-period pS waves were considered. The short-period P waves provided the best constraint. To match them, it was necessary both to develop a model for teleseismic pP since it arrives later than the near-field model would predict and to estimate the average value of \( t^* \). The MILROW and CANNIKIN teleseismic short-period P waves indicate that the average effective \( t^* \) for teleseismic ray paths from Amchitka is 0.9 s. The scatter about this value, however, is very substantial. The accuracy of the scaling laws was tested by comparing synthetics for LONGSHOT to teleseismic observations. The predicted amplitudes and wave shapes fall within the scatter in the observations, though again this scatter is substantial.

INTRODUCTION

It has long been recognized that nuclear explosions provide a unique tool for studies in seismology. They have been used extensively to constrain travel time curves [Romney et al., 1962], to map variations in crustal structure [Pakiser, 1963], and to measure regional biases in \( m_0 \) [Butler and Ruff, 1980]. Because of the recent development of efficient numerical methods for computing near-field synthetic seismograms [Helmerger and Harkrider, 1978; Bouchon, 1981], explosions can now also be used to develop an understanding of near-field strong ground motions. The source, origin time, and location are known exactly for explosions, and unless the event is contaminated by induced faulting, the records are unaffected by rupture propagation. The observed near-field ground motions can thus be used to understand the phenomenon of near-field seismic wave propagation without having to consider the many unknowns associated with earthquake sources.

Some nuclear tests have been well recorded in both the near field and at teleseismic distances. The composite data sets from such events offer especially good constraints on theoretical models for explosive sources and wave propagation. Here we report the development of relatively successful models for the three Amchitka events LONGSHOT, MILROW, and CANNIKIN. The source models correctly predict the near-field velocity waveforms at ranges from 7 to 20 km.

We have made the assumption that these records are distant enough from the source so that the material response can be modeled as being linear. This assumption should certainly be accurate for the most distant records. The source models also correctly predict the behavior of a broad spectrum of teleseismic body wave observations. The successful linking of near-field and far-field data permits us to estimate the average absolute value of \( t^* \) for paths from Amchitka to the World-Wide Standard Seismograph Network (WWSSN) stations.

Attempts to model simultaneously near-field and far-field data from explosions have been made in several previous studies. Some of the more noteworthy were the investigations of Frazier and Filson [1972], von Seegern and Blandford [1972], and Helmerger and Hadley [1981]. The last of these was perhaps the most sophisticated. Synthetic seismogram techniques were used to model near-field and far-field P waves from the Nevada Test Site (NTS) events JORUM and HANDLEY. We will adopt many of the techniques introduced by Helmerger and Hadley [1981] in this investigation.

We begin with a discussion of the near-field data from Amchitka. Three component arrays of L7 velocity meters were deployed for events MILROW and CANNIKIN. The observed near-field velocity records display clear P arrivals, an interference feature associated with \( pP \), and developing near-field Rayleigh waves. Next, we discuss the methods used for computing synthetics and formulate our model for the effective source. The source model consists of a mathematical representation for the source time function and a set of scaling laws which describe how to adjust the function for event yield. Models which fit the complete (body wave and surface wave) near-field records are then presented. No near-field data set was collected for LONGSHOT, so we predict its source using the scaling laws after first verifying them with the MILROW and CANNIKIN data.

In the final sections, the predictions of the source models are compared to the observed teleseismic data, again through
the use of synthetic seismograms. The waveforms and amplitudes of the short-period $P$ waves, long-period $P$ waves, and long-period $pS$ waves are considered. A specific feature associated with $pP$ is identified in the short-period records. Finally, the average value of $t^*$ for teleseismic $P$ waves from Amchitka is estimated from the short-period amplitude data.

NEAR-FIELD OBSERVATIONS OF MILROW AND CANNIKIN

The Amchitka testing program included three nuclear events. MILROW was reported to have a yield of 1000 kt, CANNIKIN a yield of 5000 kt, and LONGSHOT a yield of 80 kt [Marshall et al., 1979]. Many auxiliary geophysical experiments were carried out in the test area which provide information of value in analyzing the seismic data. The most important of these were measurements of the crustal structure [Engdahl, 1972]. Previous analyses of the seismic data include the works of Bakun and Johnson [1973], Davies and Julian [1972], Jacob [1972], King et al. [1972], Perret [1972], Toksöz and Kehrer [1972], and Willis et al. [1972].

The first phase of this investigation was directed at the determination of the effective seismic source functions of the nuclear events using near-field data. The near-field data collection effort involved the deployment of both surface and downhole seismic instruments. We shall concentrate our efforts on the surface velocity gauges from ranges between 7 and 20 km. Records from these ranges should be far enough away so that they are outside the zone of nonlinear material response. At ranges greater than 20 km the seismic records start to be dominated by the effects of crustal structure, and information about initial source excitation becomes difficult to resolve.

The instrument used to record the near-field data, the L7 velocity meter, was designed to have a flat response to velocity over the frequencies of interest in the near field. The traces it produces are thus essentially the same as plots of single vector components of ground velocity as a function of time. Each recording site contained a vertical, a radial, and a tangential meter. The tangential meters all showed relatively small motions. The amplitudes were only about one third the size of the radials. Comparable records from high tectonic release events at NTS typically show tangential motions as large as the radial. The absence of significant tangential movement and low $F$ factors found from Love wave analysis [Toksöz and Kehrer, 1972] indicate that little tectonic release was triggered by these events. This simplifies modeling the near-field data.

The MILROW Data Set

The MILROW underground nuclear detonation occurred on October 2, 1969, at about 2206 UT. The working point was at a depth near 1.2 km. The event was successfully recorded at six sites on Amchitka [Orphal et al., 1970]. Figure 1 displays the recording sites with respect to the MILROW epicenter. The slant ranges from the working point to the sites are 7 km to M02 and M03, 8 km to M01, 9.8 km to M04, 11.5 km to M06, and 28.5 km to M07. The M05 vertical component did not record and the M07 site is too distant to be of interest here.

Figure 2 shows the records in a distance profile. Note the excellent reproducibility of the signals at M02 and M03. This indicates that very local instrument site characteristics are not strongly affecting the signals. The numbers beside each record indicate the amplitude of the first peak in centimeters per second. We shall concentrate on modeling the amplitude of this first peak, since we believe it is the clearest measure of the source strength. Later peaks in the record are more strongly affected by wave propagation. The first-peak amplitude begins at about 35 cm/s and decays by a factor of 2 over the distance range.

The shape of the vertical $P$ wave (the first pulse on the vertical records) exhibits a clear evolution with distance. The first overswing is a single peak at the shortest ranges and becomes a double peak at the largest range. We find that this phenomena is caused by the phase $pP$ developing just behind
direct P. The Rayleigh wave can be clearly seen as a long-period pulse traveling at a slower velocity than the first arrival. (It has a period of about 2 s.) Such well-behaved and stable surface wave pulses have seldom been observed previously over these distance ranges. They provide an important opportunity to learn what parameters control surface wave propagation in the near field. This information may have important future applications in the area of near-field seismology.

The CANNIKIN Data Set

The CANNIKIN event was detonated on November 6, 1971, at about 2000 UT. The working point was at a depth near 1.8 km in volcanic breccias. The L7 velocity meter array consisted of nine recording sites on Amchitka, which are again shown in Figure 1. The M04 instrument was moved to the site indicated as M04A for CANNIKIN. Otherwise, the recording locations were the same as for the MILROW experiment. The slant ranges to the recording sites were 10.4 km to M04A, 14.7 km to M01, 14.8 km to M02, 15.8 km to M05, and 18.7 to M06. In this instance, the M03 instrument apparently malfunctioned, and the M07 and M10 sites were too distant to be included in the data set.

The distance profile of the CANNIKIN data is shown in Figure 3. The second peak in the first overswing is the pP arrival. Because the recording stations are more distant than in the MILROW profile (Figure 2) and because the source is buried more deeply, the pP arrival stands out more clearly. It shows a clear evolution of relative amplitude and arrival time with distance. The developing Rayleigh wave can again be very clearly identified. The amplitude values given again refer to the peak velocity value of the first swing in the recording. They show a distance decay of about a factor of 3.5 between 10 and 20 km beginning with the largest observed value of 35.1 cm/s at M04. The M05 vertical record exhibits much higher amplitude than nearby stations. This one value does not strongly influence our average estimate of the strength of the seismic wave, however, so we will include it in our analysis. The M01 and M02 records are very comparable even though the recording sites are separated by about 0.5 km. This again shows that local station site characteristics are not strongly contaminating the records.

METHODS FOR COMPUTING NEAR-FIELD SYNTHETIC SEISMOGRAMS

Computer Algorithms for Computing Wave Propagation in the Near Field

In order to take full advantage of the data shown in the previous section, we needed computer algorithms for calculating complete (body wave and surface wave) synthetics in the distance range of 5–20 km. The instruments were broadband, so the calculations needed to be accurate up to about 10 Hz.
still much too cumbersome for routine use in a modeling study such as this one. Calculation of synthetics for realistic models out to frequencies of 10 Hz required more than 100 times as much computer time as the rays plus modes technique.

To resolve these difficulties, we took the same approach as Burdick and Orcutt [1979]. We computed synthetic seismograms for a representative crustal model with the wave number-frequency integration code and then used this exact result to gauge which rays and which modes to include in our rays plus modes calculations. We found that we obtained reasonably good agreement between the exact and approximate calculations by including all first multiples of the P type, all first multiples of the pP type, and the fundamental Rayleigh mode. The results of our comparison are shown in Figure 4. The crust model we used is model 1 in Table 1. It is actually very close to the final crustal models we arrived at in our study.

A close comparison of the synthetics in Figure 4 shows that our limited ray sum and the fundamental mode provide good approximations to the P wave and the Rayleigh pulse. The exact calculation merely adds some smaller reverberations between the two. Similar arrivals do appear in the data, but they appear to be unstable and strongly affected by lateral variations in structure. We have not made any attempt to interpret these intermediate arrivals in this report, so we will utilize the rays plus modes algorithm throughout the remainder of our study.

**Source Time Function Representation**

Helmberger and Hadley [1981] present a clear discussion of the standard formalism for representing the velocity of the

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**Model 1: Test Model**

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**Source Time Function Representation**

Helmberger and Hadley [1981] present a clear discussion of the standard formalism for representing the velocity of the
earth's surface $V(t)$ in terms of a series of convolutions of linear operators

$$V(t) = \frac{d}{dt} \left[ \frac{d^2 \psi(t)}{dt^2} \cdot S(t) \right] \cdot h(t) \quad (1)$$

$S(t)$ is the response of the crust model to an explosive source with a step time history, $\psi(t)$ is the reduced displacement potential of the source, and $h(t)$ is the delta function response of the L7 instrument.

There are a number of possible functional representations for the reduced displacement potential (RDP) which could be chosen for these calculations. Helmberger and Hadley [1981] suggest

$$\psi(t) = \psi_\infty \{1 - e^{-Kt} \left[1 + (Kt) + \frac{1}{2} (Kt)^2 - B(Kt)^3\right]\} \quad (2)$$

since it has several advantages over the other candidates. It is important to note, however, that Haskell [1967], von Seggern and Blandford [1972], and Mueller and Murphy [1971] have suggested RDP representations that have advantages of their own. In the Helmberger and Hadley [1981] representation for the RDP, the adjustable parameters are $\psi_\infty$, $K$, and $B$. The first determines the static value of the RDP and has units of volume. The second controls the source rise time or corner frequency and has units of inverse time, and the third is a unitless parameter which controls source overshoot.

There is one drawback to using the Helmberger and Hadley [1981] source potential. When they presented it, they did not develop scaling relations to describe how the free parameters should vary with yield. Von Seggern and Blandford [1972] and Mueller and Murphy [1971] did derive scaling laws for their source functions, and elements of their derivation can be easily adapted for the Helmberger and Hadley [1981] source representation. The fundamental scaling laws we propose for the Helmberger and Hadley [1981] source representation are

$$\psi_\infty \propto Y/h^{0.27} \quad (3)$$

$$K \propto h^{1.24}/Y^{1/3} \quad (4)$$

$$B = \text{const} = 1 \quad (5)$$

Because $\psi_\infty$ is a static parameter and the various RDP representations vary only in their dynamic aspects, the first of these scaling relations is appropriate for all of the representations. It was suggested by Murphy and Mueller [1971] and also recommended by T. Lay et al. (unpublished manuscript, 1984). Von Seggern and Blandford [1972] show how to derive a scaling law for $K$ by deriving the pressure function at the elastic radius from the RDP (equation (2)). A comparison of this pressure function with that of Mueller and Murphy [1971] leads to the identification of $K$ as being proportional to $r_E^{-1}$; $r_E$ is the elastic radius, and Mueller and Murphy [1971] specify the dependence of $r_E$ on $Y$ and $h$. Their relationship for $r_E$ is just the reciprocal of relation (4). The relationship for $B$ is approximate in that strict adherence to the derivation of von Seggern and Blandford [1972] predicts that $B$ should increase slightly with depth. However, this dependence is negligible for our purposes. Furthermore, in the course of our analysis we found that the parameter $B$ is not resolvable. It trades off very strongly with the value for $\psi_\infty$. The final results do not depend in an important way on the particular choice made for $B$. The value $B = 1$ corresponds approximately to the overshoot level that von Seggern and Blandford [1972] found to be appropriate for granite. (That is to say, $B = 1$ in the Helmberger and Hadley [1981] formalism corresponds approximately to $B = 2$ in the von Seggern and Blandford [1972] formalism.)

The reason that $B$ trades off with $\psi_\infty$ and the reason that the particular $(B, \psi_\infty)$ pair chosen does not alter our results can be easily understood by considering the spectrum of the time function. Von Seggern and Blandford [1972] show a suite of theoretical spectra for varying choices for $B$ in their Figure 5. The spectra of explosion time functions are level at long periods, but then they rise to a peak at intermediate periods. At high frequencies the spectra fall off rapidly just as for earthquakes. The wave shapes of the near-field velocity records and the teleseismic short-period records from MILROW and CANNIKIN are dominated by the high-frequency part of the spectrum. They are insensitive to its long-period character. The zero-frequency value of the spectrum is directly proportional to $\psi_\infty$. The value for $B$ controls the height of the spectral peak. If the value for $\psi_\infty$ is lowered, the value for $B$ can be increased to increase the height of the peak, leaving the high-frequency part of the spectrum unchanged. Higher values for $\psi_\infty$ can be balanced with lower values for $B$. Thus the values for $B$ and $\psi_\infty$ simply control the portions of the spectrum which have no effect on high-frequency waveforms. At several stages in our investigation we have tested the generality of the trade-off. We found in each case that the value for $B$ could be changed from 0 to about 5 and all significant changes could be immediately absorbed by a change in $\psi_\infty$. Our final estimate of $B$ is a high-frequency estimate, so our choice for $B$ does not affect it.

The Effects of the Instrument and Attenuation Operators

As noted previously, the L7 velocity meter has an essentially flat response to ground velocity from 10 s to 20 Hz. Thus the signals shown in Figures 2 and 3 are virtually the same as plots of ground velocity with time. We did apply an instrument correction to our synthetics in our modeling study ($\theta(t)$ in equation (2)), but we found that it had almost no effect, as expected.

There have been numerous time domain studies of strong motion records for earthquakes at distances comparable to those considered in this study [Hartzell and Brune, 1979; Heaton and Helmberger, 1979]. Liu and Helmberger [1980] have bracketed $t^*$ to be between 0.05 and 0.08 s for short paths in the Imperial Valley. It is reasonable to assume that $t^*$ would be as small or smaller for Amchitka Island. We tested the effect of an attenuation operator evaluated at $t^* = 0.05$ s and found the effect to be so small that we did not include an attenuation correction in further calculations.
MODELS FOR MILROW, CANNIKIN AND LONGSHOT

In the previous section it was shown that for calculation of near-field synthetics we needed to develop both a crustal model for computing the effects of wave propagation and a source time function or RDP model for computing the effects of initial source excitation. Unfortunately, there are some trade-offs between source and structure, but they are not too severe. The crustal model completely determines the relative arrival times of the various phases such as $P$, $pP$, and the Rayleigh pulse. The source time function primarily controls the frequency content of the $P$ wave. Both the crust model and the RDP control the relative amplitude and frequency content of the Rayleigh wave.

The MILROW Crustal Model

The crustal structure of Amchitka was studied in some detail by Engdahl [1972]. He presented a $P$ wave velocity model based on refraction and drill hole measurements. We used his structure as the starting model for our study and altered it as necessary to fit the data. We found that the near-field records are very sensitive to the top 3 km of the crust. Both the $P$ wave and the $S$ wave velocities were adjusted to fit the synthetics. The $P$ wave gradient at the top of the crust was modified from the starting model to fit the decay of the direct $P$ arrival with distance and the interference with $pP$. The perturbations from Engdahl's model were kept small enough so that the absolute travel time predictions would be very similar. In the initial calculations it was assumed that the ratio of the $P$ to $S$ velocity was about $3^{1/2}$. It was immediately clear that the shear velocity needed to be lowered to $V_p/V_s \sim 2$ in order to fit the Rayleigh wave arrival time. The details of the $S$ wave gradient also had to be adjusted separately from the $P$ wave gradient in order to fit the shape of the Rayleigh pulse.

The crustal model for MILROW is given in Table 1 and compared to Engdahl's [1972] model in Figure 5. The two models obviously share the same average velocity, but the MILROW crust model is more detailed. Note also how the $S$ wave and $P$ wave profiles differ. These differences may or may not, however, represent true differences in the earth. The top 3 km of the earth vary strongly in their properties, and the model in Figure 5 represents an average. The $P$ wave and Rayleigh wave sample the earth in different ways, which might easily result in the different $P$ and $S$ wave profiles.

The MILROW Source Function

The crustal model was derived using first approximations to the source parameters $K$, $B$, and $\psi_0$. After the crustal model was fixed, the source parameters were refined. The waveforms...
are most strongly affected by \( K \). It changes the frequency content of the \( P \) pulse and the ratio of the \( P \) pulse to the Rayleigh pulse. The best fitting \( K \) value for MILROW was found to be 9 s\(^{-1}\). We verified that the value chosen for \( B \) had little effect on our synthetic waveforms and then left it fixed at 1 as prescribed by our scaling laws. The value for \( \psi_0 \) was determined by matching the amplitude of the first peak on the vertical records. The final value, \( \psi_0 = 1.4 \times 10^{11} \) cm\(^3\), is the value indicated by the average ratio of the predicted/observed amplitudes from the five vertical velocity records. Helmberger and Hadley [1981] point out that it is better to rely on the vertical rather than the radial records, since they are less sensitive to crustal structure. Nonetheless, our estimate does not change substantially if we include the radial amplitudes in the calculation.

Figure 6 compares the observed to the synthetic vertical velocity records from MILROW. There are three key features in the data which have been matched. The first is the overall shape of the \( P \) waves and the double-peaked overswing. This double peak is caused in the synthetics by generalized rays of the \( pP \) type. The arrival is not completely stable in the data in that it appears to be reduced at M01, but we attempted to fit its average amplitude and timing as closely as possible. The second key feature is the relative amplitude and timing of the Rayleigh pulse. Note that its general shape and period are also matched. The final feature predicted is the \( P \) wave amplitude and its decay with distance. For this set of records the predicted amplitudes are all within 20% of the observed and in most cases within 10%.

Figure 7 compares the observed to the synthetic radial records. The general quality of the fit is not as good as for the verticals, though the same features are matched to some extent. The \( pP \) phase does not stand out clearly in the data, though the synthetics predict that it should. One reason that the fit to radial records is poorer is that these records are much more strongly affected by \( P \) to \( S \) converted energy than the verticals. This type of energy is not accounted for in our calculation. The timing and shape of the Rayleigh pulse is well matched, though the theoretical pulse is smaller than the observed at the closest two stations. Each trace in the figure is normalized independently. When the different amplitude scales are accounted for, it becomes clear that the observed and predicted Rayleigh wave amplitudes at the farthest stations match well. Likewise, the absolute amplitudes at the closest two stations do not match, though they do at the farther three. The first arrivals in the synthetic seismograms

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**Fig. 7.** Synthetic seismograms computed for our source and crust models are compared to observed radial velocity records from MILROW. The peak amplitude of the first swing is indicated in centimeters per second.
are diving rays in each case. The direct ray traveling upward from the source to the recording site has a small effect at these ranges.

The CANNIKIN Crustal Model

The crustal model for the MILROW calculations needed to be modified for the CANNIKIN calculations. The modifications were very slight, however, and involved only the top 3 km of the crust. Since the travel paths from the CANNIKIN shot point to the recording stations are different than for the MILROW shot point, lateral variations of this order are to be expected. The CANNIKIN crust model is given in Table 1 and compared to the other models in Figure 5. The figure shows that the only significant modification in the CANNIKIN model is that the large interface near 7 km depth is displaced downward by about 0.5 km. We found that the CANNIKIN data require that the S wave and P wave gradients have different shapes, as did the MILROW data.

The CANNIKIN Source Time Function

The K value for CANNIKIN was adjusted to fit the frequency content of the P wave and the ratio of the body wave to the surface wave just as for MILROW. The best fitting value was found to be $6 \text{s}^{-1}$. The value for B, the overshoot parameter, was left fixed at 1. The average value for $\psi_0$, again determined from the average ratio of the predicted/observed amplitudes of the vertical records, was found to be $\psi_0 = 4.5 \times 10^{11} \text{cm}^3$.

Figure 8 compares the observed to the synthetic vertical velocity records. The quality of the fit is perhaps even better than for the MILROW vertical records. The $pP$ arrival is clear in every observation. Its timing and amplitude is well predicted by every synthetic except M05. The absence of $pP$ in the synthetic is due to a delicate interference caused by the coarse layering. Slight perturbations in distance cause the $pP$ arrival to reappear with strength. The Rayleigh pulse shape and timing are well matched. The fit to the amplitude of the first peak is worse than for the MILROW verticals in that the error for M02 and M05 exceeds 20%. The mismatches have opposing senses, however, and thus do not strongly control the estimate of $\psi_0$.

Figure 9 compares the observed radial velocity records for CANNIKIN to the predictions of the model. Again the synthetics do not match the observations as closely as they did for the verticals. The $pP$ arrival is clearly present in several of the records, the Rayleigh pulse timing and amplitude is matched, and the correspondence between the observed and predicted amplitudes is no worse than for the verticals. We conclude from the basic success of our models in predicting the amplitude-distance decay of the near-field records that nonlinear material response is not having a resolvable effect.

The LONGSHOT Source Time Function

Because there was no near-field data collection program for LONGSHOT, we must infer its source time function from the scaling laws and from the CANNIKIN and MILROW measurements. The form of the scaling laws is given in equations (3), (4), and (5), but we need to fix the proportionality
 constants. We will select MILROW to fix the constants and compare the predictions to CANNIKIN. Our scaling laws thus become

\[ \psi_\infty = 9.5 \times 10^8 \frac{Y}{h^{0.27}} \]
\[ K = 4.7h^{1/2}Y^{1/3} \]
\[ B = 1 \]

where we have used the fact that MILROW's depth was 1220 m, its yield was 1000 kt, its \( K \) was 9 s\(^{-1}\), and its \( \psi_\infty \) was \( 1.4 \times 10^{11} \) cm\(^3\). The predictions of the scaling laws for CANNIKIN are \( K = 6.2 \) s\(^{-1}\) and \( \psi_\infty = 6.3 \times 10^{11} \) cm\(^3\) using \( Y = 5000 \) kt and \( h = 1800 \) m. The values from trial and error modeling of the data were \( K = 6 \) s\(^{-1}\) and \( \psi_\infty = 4.5 \times 10^{11} \) cm\(^3\), which is certainly an acceptable comparison. The scaling laws thus predict for LONGSHOT that \( K = 16.7 \) s\(^{-1}\), \( \psi_\infty = 1.3 \times 10^{10} \) cm\(^3\), and \( B = 1 \) where values of 80 kt and 700 m have been substituted for \( Y \) and \( h \). The scaling laws which we propose were subjected to further testing in the remainder of our study where we attempted to link the near field to the teleseismic Amchitka data.

**TELESEISMIC OBSERVATIONS OF THE AMCHITKA TESTS**

Amchitka Island is favorably situated with respect to the WWSSN. Station coverage is dense at all azimuths except for a small range to the southeast. The uniform instrumentation of the WWSSN makes it possible to intercompare waveforms easily, and the azimuthal coverage permits us to estimate the magnitude of the effects of lateral variations in the earth. Because of the high-frequency content of the signals from explosions and the slow WWSSN recording speeds, we believe that time domain analysis of this data is probably more reliable than spectral analysis. Therefore we shall use waveform modeling methods in this portion of our study just as in the near-field portion.

All available WWSSN short-period P wave records were collected for our analysis as well as long-period P and PS records from the larger two events. Since we wished to avoid complexities introduced by the earth's core and upper mantle, we limited our final data set to records from epicentral ranges between 30\(^0\) and 95\(^0\). As many of the signals as possible were digitized to facilitate the waveform comparisons. There are two features of the short-period P waves which are important in this analysis. The first is the interaction in the waveforms between P and pP, and the second is the amplitude. In this section we display the data both to illustrate these features and also to show how much they scatter. In our analysis section we will present an average value for \( t^* \) and an estimate of the average properties of pP. We will also, however, attempt to determine how much our estimates would have to be changed to explain the variations in the data.

**Short-Period Waveforms and Teleseismic pP**

Figure 10 shows the short-period P waveforms in the MILROW data set. The traces are normalized to a common
pP does not appear to vary with azimuth in a smooth way, having an effect. It is not possible to determine uniquely which of these is responsible from the data at hand. The stations in Figure 11 are ordered by increasing azimuth. The behavior of stations in the earth or azimuthal variations in the source are not apparent in several cases. Either lateral variations in the crustal source or mantle propagation effects which cannot be explained rigorously. The second upswing appears to be affected by the arrival of pP in the MILROW and CANNIKIN waveforms. This implies that pP arrives late, but it also implies that the first peak-trough or absolute amplitudes are essentially measurements of the direct P arrival. The LONSHOT absolute amplitudes are probably affected by pP, but we must use them for consistency in the data sets.

In order to estimate $t^*$ it is desirable to analyze the most stable part of the short-period waveform. The variations in P wave coda after the first few seconds reflect different receiver and mantle propagation effects which cannot be explained rigorously. The second upswing appears to be affected by the arrival of pP in the MILROW and CANNIKIN waveforms. This implies that pP arrives late, but it also implies that the first peak-trough or absolute amplitudes are essentially measurements of the direct P arrival. The LONSHOT absolute amplitudes are probably affected by pP, but we must use them for consistency in the data sets.

To observe the azimuthal scatter in the amplitude data, we must first correct for distance and instrument gain. To intercompare data from all three events, we must correct for differences in their yields. We used the curve given by Langston and Helmberger [1975] to correct the observations to a common distance of 30°, then normalized all values to unit instrument gain, and finally used the procedure outlined by Butler and Ruff [1980] to estimate event corrections. The Butler and Ruff [1980] event corrections are designed to minimize the average scatter at the observing stations after the corrections have been applied.

Figure 12 shows all of the relative short-period absolute amplitude data from the Amchitka events. The event corrections have been applied. A substantial amount of scatter still remains, with the data for each event ranging over more than a factor of 10 in amplitude. However, the scatter at any one station between LONSHOT, MILROW, and CANNIKIN is typically less than a factor of 2. This stability of the relative signal sizes for fixed source station paths implies that the mechanisms producing the large azimuthal variations are not

which is most easily explained in terms of lateral variations near the receivers. A pP phase which varies rapidly with azimuth from the source cannot, however, be ruled out.

**Short-Period Amplitudes and the Scatter in Them**

In order to estimate $t^*$ it is desirable to analyze the most stable part of the short-period waveform. The variations in P wave coda after the first few seconds reflect different receiver and mantle propagation effects which cannot be explained rigorously. The second upswing appears to be affected by the arrival of pP in the MILROW and CANNIKIN waveforms. This implies that pP arrives late, but it also implies that the first peak-trough or absolute amplitudes are essentially measurements of the direct P arrival. The LONSHOT absolute amplitudes are probably affected by pP, but we must use them for consistency in the data sets.

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height since we will consider the amplitude information separately. Most of the waveforms show good coherence over the first three swings but pronounced variations after this. These later differences are presumably caused by receiver structure. The explosive source process and hence the direct P pulse should be over after the first few seconds. Thus we shall concentrate on matching the average absolute (first small upswing to first trough) amplitude. We are also interested in the detailed structure of the second upswing. Note the clear double peak at AAM, WES, and GUA. Note also the small shoulder in MAL, STU, and COP. Burdick and Helmberger [1979] suggested that this feature is caused by the arrival of pP. We shall show that this interpretation is probably valid, though the pP arrival is late compared with the times predicted by the crustal model.

The CANNIKIN short-period P data set contained 38 records and the LONSHOT data set contained 44. They are similar to the MILROW data set in that the records are coherent over the first few swings and notably incoherent at later times. Figure 11 compares the P waves from the Amchitka events at those stations that successfully recorded all three. Since the sources were so close together, the ray paths to each station are essentially identical. The evolution of the waveform at each station must be due to the changes in the source function and surface interaction from LONSHOT to CANNIKIN. The increase in dominant period with yield as well as the development of the pP arrival in the second upswing are readily apparent. A double-peaked second upswing appears in over half the CANNIKIN records shown. Nonetheless, the feature is not apparent in several cases. Either lateral variations in the earth or azimuthal variations in the source are having an effect. It is not possible to determine uniquely which of these is responsible from the data at hand. The stations in Figure 11 are ordered by increasing azimuth. The behavior of pP does not appear to vary with azimuth in a smooth way,
associated with the source process or the near-source structure. A similar conclusion was reached by Heimbarger and Hadley [1981] for NTS events. In order to appreciate the significance of any global t* estimate based on this data, it is important to characterize at least qualitatively the cause of this amplitude scatter.

The azimuthal amplitude pattern in Figure 12 does exhibit some broad trends. The lowest amplitudes at azimuths near 340° were reported by the middle eastern stations. The observations at azimuths near to due north, which are somewhat low, are from European stations. The observations from other azimuths scatter about the mean. Many investigators have attempted to quantify the effects of the subducting Aleutian slab on the Amchitka data. Davies and Julian [1972] and Sleep [1973] in particular considered how amplitudes might be affected. Their work suggests that low amplitudes in Canada (not included in our data) and Europe might be due to the presence of the slab. In our calculations of t* we have elected to include all of the observations. The detailed velocity and Q structure of the descending lithosphere are not well enough known to permit making accurate corrections. The data from Europe and the middle east could have simply been excluded, but this seems arbitrary without a much firmer reason for doing so.

One important question we wish to address is how teleseismic amplitudes scale with yield since this question is important in treaty monitoring efforts. It is generally assumed that m0 is linear with the log of the yield. We find that in this case the log of the average absolute amplitude is given by

\[ \log(A_{ab}) = 0.504 \log(Y) + 1.39 \]  

The observed log (Aab) for LONGSHOT is 2.35 averaged over 44 observations. The prediction from the curve is 2.35. The observed for MILROW is 2.80 averaged over 44 observations, and the predicted is 2.80. The observed for CANNIKIN is 3.26, and the predicted is 3.25. The slope of this relation differs from those found for m0* or log (A/T) regressions on yield due to the nature of the Aab measurement.

**ANALYSIS OF THE TELESEISMIC DATA**

The goal of this phase of our modeling effort was not only to explain the average amplitude and waveform properties of the teleseismic observations but also to characterize the variations in them in terms of variations in t* and in properties of pP. Determining appropriate value for t* near 1 Hz is an important scientific goal. There could be some question, however, as to why one should consider variations in pP in detail. In principle, we could have assumed that pP has the properties predicted by elastic wave propagation theory. There is, however, strong evidence from the literature that this assumption is not appropriate. Cohen [1970], King et al. [1972], Bakun and Johnson [1973], and Burdick and Heimbarger [1979] have pointed this out for the Amchitka tests, but it also seems to be true in general. Changes in pP relative amplitude and timing have an effect on short-period waveforms similar to the effect of changes in t*. It is important to characterize this trade-off in order to understand fully the uncertainty in t*.

Several different techniques are available for computing synthetic teleseismic waveforms for a given source time function and a given earth structure. We elected to use the procedure outlined by Langston and Heimbarger [1975] because arbitrarily changing the properties of pP is simple when using their approach. The same procedure was used for both P and PS calculations. The source velocity structures were the same as for the near-field calculations. An average crustal structure with P velocity of 5 km/s and S velocity of 2.9 km/s was assumed for the WWSSN receivers. We follow Langston and Heimbarger [1975] in using a frequency independent attenuation operator. We will discuss this choice in more detail in a following section.

**Trade-Offs Between pP and t**

To demonstrate the trade-offs between pP and t*, we show the effects of varying pP-P time, t*, and the \( \frac{|P||P|}{|P|} \) amplitude.
Fig. 14. Synthetic short- and long-period WWSSN P waves showing the effect of $t^*$ on the teleseismic waveforms. The synthetics are for the CANNIKIN source model ($K = 6, B = 1$) with $pP/P = 0.9$. The variability in the first 3 s of the observed waveforms and amplitudes can be explained by a range in $t^*$ from 0.7 to 1.5 s for a $pP$-P delay of 1.15 s. While varying the $pP$ delay produces similar waveform variations (Figure 13), it does not account for the order of magnitude variations observed.

Fig. 15. Synthetic short- and long-period WWSSN P waves showing the effect of $pP$ amplitude on the teleseismic waveforms. The synthetics are for the CANNIKIN source model ($K = 6, B = 1$) with $t^* = 0.9$ s. For the delay time indicated by the data, 1.15 s, the observations require that the $pP/P$ ratio is not less than 0.7, with no clear overall inconsistency with the elastic prediction of $pP/P = 0.9$.

In summary, the variations in short-period waveforms indicated in Figure 11 could be explained in two ways. One would be azimuthal variations in the properties of $pP$, particularly the $pP$-P lag time. The other would be $t^*$ variations from 0.5 to 1.5 s. Only the latter explanation also predicts the large scatter in amplitudes shown in Figure 12. We will show in the following section that there is a modest correlation between high amplitude (low $t^*$) and the emergence of the $pP$ arrival. Because the correlation is not complete, it is possible that both $pP$ and $t^*$ are varying azimuthally. There is undoubtedly a contribution to the waveform variations due to scattering effects.

Comparisons of Observed and Predicted Teleseismic Signals

In Figure 16 we compare synthetics computed from our final model to a representative suite of the observed CANNIKIN synthetics in Figures 13–15. In Figure 13 the effect of increasing the $pP$-P lag time is shown for two values of $t^*$. The source time function and source crust models are the ones developed from the near-field data. The short-period synthetics can be compared to the observations shown in Figure 11. The low $t^*$ value predicts a well-developed $pP$ arrival in the second upswing for all lag times, and for lag times longer than 1 s the waveforms are unlike any observations. The time separation of the first peak and first trough is too short to match the observations as well. In order to bring these waveforms into agreement with the data a much lower-frequency time function would have to be used. For $t^* = 0.9$ s the frequency content of the short-period signals is in much better agreement with the majority of the data. Slowing $pP$ down by 0.3 s produces the interference seen in many of the CANNIKIN short-period waveforms. There are relatively minor changes in the long-period synthetics for any of the $t^*$ or $pP$-P time values considered. With $t^*$ set at 0.9 s and a range in $pP$-P time from 0.9 to 1.2 s most of the observed waveforms can be matched over the first three swings. It is important to emphasize that $pP$-P time has very little effect on amplitude. We must find some other explanation for the order of magnitude variation shown in Figure 12.

Synthetics computed for two different lag times and a range of $t^*$ values are shown in Figure 14. For the elastic case ($pP$-P = 0.92 s), $t^*$ values less than 0.5 s are needed to produce the interference in the second upswing of $pP$. However, the amplitude and frequency content of the synthetics do not match the observations. To match both the average amplitude and the timing of the short-period signals, we found that we needed a $pP$-P lag time near 1.15 s. Synthetics for this $pP$-P value are shown on the right of Figure 14. This value compares favorably with the estimate of Burdick and Helmberger [1979] (1.1 s) and Bakun and Johnson [1973] (1.1–1.2 s). Fixing the $pP$-P value and varying only $t^*$ allows one to explain the observed variations in both amplitude and waveform to a large degree. Some variations in the long-period synthetics are visible in Figure 14 as $t^*$ increases to a large value. Corresponding variations do, in fact, occur in the observations.

Figure 15 shows the effect of diminishing $pP$ amplitude. Synthetics for values of the $|pP|/|P|$ ratio less than 0.7 do not match any of the observed waveforms. The effect of relative $pP$ amplitude on absolute amplitude is not strong, so it probably does not contribute to the scatter in these data in a substantial way. It is not possible to rule out diminished $pP$ amplitude, but there is no clear indication of it in our data, and the best fitting models are generated using the elastic reflection coefficient $|pP|/|P| = 0.9$.

In summary, the variations in short-period waveforms indicated in Figure 11 could be explained in two ways. One would be azimuthal variations in the properties of $pP$, particularly the $pP$-P lag time. The other would be $t^*$ variations from 0.5 to 1.5 s. Only the latter explanation also predicts the large scatter in amplitudes shown in Figure 12. We will show in the following section that there is a modest correlation between high amplitude (low $t^*$) and the emergence of the $pP$ arrival. Because the correlation is not complete, it is possible that both $pP$ and $t^*$ are varying azimuthally. There is undoubtedly a contribution to the waveform variations due to scattering effects.
period waveforms is matched by the predictions. The fit de-
near-field data. The pulse width of the first peak of the long-
This is again probably due to receiver structure.

The pP lag time used for the calculations was
These *t variations can apparently explain the variation in the
bottom we show synthetics for *t values of 0.8, 1.1, and 1.3 s.

The average t* in the earth is 0.92 s. However, it must be
emphasized that the scatter in the amplitude data, and hence
the uncertainty in this determination, is substantial. CANNI-

KIN waveforms. Both short- and long-period P waves are
shown for each station. We have grouped the data into three
basic categories based on the short-period waveforms. The pP
shoulder is most pronounced in the waveforms in the left
column and least pronounced in the right column. On the
bottom we show synthetics for *t values of 0.8, 1.1, and 1.3 s.
These *t variations can apparently explain the variation in the
observations. The pP lag time used for the calculations was
1.15 s, and the free surface reflection was assumed to be elas-
tic. The source time function was the one determined from the
near-field data. The pulse width of the first peak of the long-
period waveforms is matched by the predictions. The domin-
ant period of the signals increases.

The average short-period absolute amplitude for CANNI-
KIN is 1.8 μm for a unit gain instrument. This indicates that
the average *t in the earth is 0.92 s. However, it must be
emphasized that the scatter in the amplitude data, and hence
the uncertainty in this determination, is substantial. CANNI-
KIN was an unusually large event, so it produced a large
number of usable long-period P waves. We have measured
the amplitude of the first peaks of these signals and wherever
possible have determined the short-period to long-period am-
plitude ratio. We found that the long-period amplitudes alone
were not a strong function of *t. The short-period to long-
period amplitude ratio, however, turns out to be a stable and
sensitive indicator of *t. To match the observed ratio with our
synthetics requires that *t is 0.95 s. The observed amplitudes
and *t estimates are summarized in Table 2. Note in Table 2
the large standard deviations in the amplitude measurements.

To illustrate how the scatter in the amplitude observations
is related to variation in *t, we have plotted the short-
period/long-period ratio against short-period amplitude in
Figure 17. The dashed curve is the trajectory predicted by our
synthetic seismograms for variations in *t. The observations
follow the curve reasonably well, which supports the idea that
variations in short-period amplitude are caused by variations
in effective *t. Note, however, that to explain the range of
amplitudes, we must allow *t to range from at least 1.5 to 0.5
s. This is a larger range than is generally believed to exist in
the earth. On the other hand, it corresponds to the range of *t
values required to match the short-period waveform vari-
aditions. (This is particularly true of the MILROW and
LONGSHOT observations, as we shall show.) The correspon-
dence between the strength of the pP shoulder in Figure 16
and the amplitude can be seen in Figure 17. KIP, ATL, and
OGD exhibited a clear pP shoulder in Figure 16 and are high
in amplitude. Both behaviors are indicative of lower *t. LEM,
BHP, and PMG were from the lowest-frequency group,
grouped little or no pP effect, and plotted low on the amplitude
curve in Figure 17. These are consistent expressions of high
effective *t. Station SHK from the high-frequency group does
not follow the pattern. Nonetheless, the hypothesis that the
variations in the short-period records are primarily due to
variations in effective *t seems to explain the observations
reasonably well. It is important to note that the amplitude
pattern is not simply related to tectonic patterns. The highest-
amplitude stations, BEC and KIP, are on islands. Some of the
lowest-amplitude stations are thus on the shield or near-shield
portions of northern Europe. Thus our estimate of *t should
perhaps be referred to as an effective *t. Several different ef-
effects including scattering, defocusing by velocity gradients
near the slab, and anelastic absorption, are being modeled by
an effective Futterman [1962] attenuation operator.

CANNIKIN also produced a large pS data set. In Figure 18
we compare some of the pS waveforms to synthetics computed
from our model. We have once more divided the observations

<table>
<thead>
<tr>
<th>Event</th>
<th>Measurement</th>
<th>Average Amplitude or Amplitude Ratio</th>
<th>Standard Deviation</th>
<th>No. of Observations</th>
<th>t* s</th>
<th>t* s</th>
</tr>
</thead>
<tbody>
<tr>
<td>CANNIKIN</td>
<td>short-period P absolute amplitude</td>
<td>1.814μ</td>
<td>1.333</td>
<td>38</td>
<td>0.94</td>
<td></td>
</tr>
<tr>
<td>CANNIKIN</td>
<td>short-period/long-period P amplitude ratio</td>
<td>0.65</td>
<td>0.35</td>
<td>28</td>
<td>0.95</td>
<td></td>
</tr>
<tr>
<td>CANNIKIN</td>
<td>long-period pS amplitude (radial)</td>
<td>2.83μ</td>
<td>1.069</td>
<td>21</td>
<td>2.5</td>
<td></td>
</tr>
<tr>
<td>MILROW</td>
<td>short-period P absolute amplitude</td>
<td>0.791μ</td>
<td>0.480</td>
<td>44</td>
<td>0.85</td>
<td></td>
</tr>
<tr>
<td>MILROW</td>
<td>short-period/long-period P amplitude ratio</td>
<td>1.02</td>
<td>0.39</td>
<td>13</td>
<td>0.90</td>
<td></td>
</tr>
</tbody>
</table>
The average short-period absolute amplitude and short-period/long-period amplitude ratio for MILROW are given in Table 2. The former indicates a $t^*$ value of 0.85 s and the latter a value of 0.90 s. As noted previously, the scatter in the MILROW amplitudes tracks the scatter in the CANNIKIN amplitudes, so the MILROW observations again imply large variations in $t^*$ about the mean value.

The LONGSHOT source function was not measured from near-field records as those of the larger events were. The source parameters were simply estimated from the yield scaling laws. We thus can only use LONGSHOT as a consistency check on the scaling laws. LONGSHOT was shallow enough so that we could not see a $pP$ shoulder in the teleseismic data. We therefore cannot independently measure the $pP$ arrival time. To compute synthetics for comparison with the observations, we adopt the source time function predicted by the scaling laws, adopt the $pP$ lag time estimated by spectral techniques of 0.55 s by Cohen [1970] and King et al. [1972] and assume an elastic free surface reflection. (This $pP$ lag time is again larger than the elastic predictions of 0.4 s.) The resulting synthetics are compared to some of the observed short-period waveforms in Figure 20. We have again divided the observations into three categories. The $t^*$ range necessary to explain the variations, 0.3, 0.9, and 1.3 s, is the same range as was required for MILROW. We attempted to test the sensitivity of our LONGSHOT synthetics to the various parameters we needed to adopt. We found that the synthetic waveforms and amplitudes were relatively insensitive to variations in $K$. This is because for large $K$ values like 16.7 s$^{-1}$ the RDP appears as a step function in the short-period pass band. The rise time, which is the feature controlled by $K$, is essentially instantaneous. The $pP$-$P$ lag time is not constrained by the waveforms to be later than elastic as we have assumed, but since both MILROW and CANNIKIN had delayed $pP$ times,

$$t^*$$ necessary to map out the waveform variations in Figure 19 are 0.3, 0.5, 0.9, and 1.3 s. This is very close to the variation indicated in Figure 17. The MILROW long-period $P$ wave data set was much smaller than for CANNIKIN. Four representative waveforms are shown on the right of Figure 19. They are matched reasonably well by a synthetic with a $t^*$ of 0.9 s.

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is contained within our estimate of the effective attenuation operator a component of energy loss due to scattering. Here ray path have a substantial effect on the wave field. Thus there characterize the very complex process of attenuation by a how that field attenuates and evolves as it propagates to tele-

It is well known that lateral variations in velocity all along the single number, the $t^*$ parameter in the Futterman $Q$ operator. 

A very valid criticism of the study is that we are attempting to apply our scaling laws over more lateral variations in effective $t^*$ would have no basis for constraining a more complex model. If the results of this study are ever compared to those of other works, it would be best to consider our estimate of $t^*$ as an average value for the frequency band of the WWSSN short-period instrument, which is around 1 Hz.

**Conclusions**

In the course of this investigation we have developed several methods and achieved several results that could be of some importance. We established the accuracy and utility of the rays plus modes approach for modeling near-field seismograms. We introduced and tested scaling laws for the Helmberger and Hadley [1981; Frazier and Filson, 1972]. The second reason is that even though it is generally accepted that attenuation is frequency dependent there is no consensus on what frequency dependent model is appropriate [Burdick, 1982]. Choice of any given frequency dependent attenuation operator would be as arbitrary as choice of the frequency independent one. The third reason is by far the most compelling one. The variations in the teleseismic observations can be explained well in terms of variations in the simple Futterman operator. This refers both to variations in the waveforms (Figures 16, 18, 19, and 20) and variations in the simple Futterman operator. This refers both to variations in the waveforms (Figures 16, 18, 19, and 20).

**DISCUSSION**

Viewed in its entirety, this experiment can be seen as an attempt to characterize the strength and frequency content of a seismic wave field from near-field data and then to measure how that field attenuates and evolves as it propagates to teleseismic distances. A very valid criticism of the study is that we characterize the very complex process of attenuation by a single number, the $t^*$ parameter in the Futterman $Q$ operator. It is well known that lateral variations in velocity all along the ray path have a substantial effect on the wave field. Thus there is contained within our estimate of the effective attenuation operator a component of energy loss due to scattering. Here we use the term scattering in a generalized sense to mean any effects due to departure of the earth from radial symmetry. Included within this would be effects due to the fact that WWSSN station site characteristics vary. The remainder of our effective attenuation operator is then due to true anelasticity. However, it is generally accepted that such anelasticity is frequency dependent. The Futterman operator is frequency independent in the seismic band. It would seem that a much more complex model for the attenuation process is warranted.

We have elected to use the simple Futterman [1962] operator for three reasons. The first is that it allows us to compare our results with previous work [Helmberger and Hadley, 1981; Frazier and Filson, 1972]. The second reason is that even though it is generally accepted that attenuation is frequency dependent there is no consensus on what frequency dependent model is appropriate [Burdick, 1982]. Choice of any given frequency dependent attenuation operator would be as arbitrary as choice of the frequency independent one. The third reason is by far the most compelling one. The variations in the teleseismic observations can be explained well in terms of variations in the simple Futterman operator. This refers both to variations in the waveforms (Figures 16, 18, 19, and 20) and variations in the simple Futterman operator. This refers both to variations in the waveforms (Figures 16, 18, 19, and 20).

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