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DARPA-NMR-81-01

A TECHNICAL ASSESSMENT OF SEISMIC YIELD ESTIMATION

APPENDIX

PART 1 OF 2

JANUARY 1981

DEFENSE ADVANCED RESEARCH PROJECTS AGENCY DARPA

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PREFACE

The DARPA report, "A Technical Assessment of Seismic Yield Estimation," is a review of the state of current understanding of important technical issues relating to the seismic estimation of the yield of underground nuclear explosions. In preparing this review, contributions were solicited from selected government, university, and industry scientists who responded with summaries of their assessment of the state of knowledge in those areas with which they were most familiar. These contributions are collected in this Appendix.

The forty-eight separate summaries from thirty-one authors are listed by title and author in the next few pages. They are organized according to the eight subject areas of the main report. The contributions follow the listing in the indicated order with 1 to 26 in Part 1 and the remaining 22 in Part 2 of this Appendix.

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2. Robert J. Zavadil, Air Force Technical Applications Center (AFTAC)
   "Body Wave Magnitude"

3. Karl F. Veith, Teledyne Geotech, Garland, Texas
   "Body Wave Magnitude"

4. Thomas D. Eisenhauer, AFTAC
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5. Robert R. Blandford, Teledyne Geotech, Alexandria, Virginia
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6. Thomas C. Bache,* Systems, Science and Software, La Jolla, California
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8. Lane R. Johnson, University of California, Berkeley
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11. John R. Murphy, Systems, Science and Software, Reston, [illegible]nia
    "Experimental Data on Body Wave Coupling"

12. William R. Perret, Albuquerque, New Mexico
    "Evaluation of Body Wave Coupling Through Experimental Data"

13. Donald V. Helmberger, California Institute of Technology, Pasadena
    "Near Source Effects on P-Waves"

14. John G. Trulio, Applied Theory, Inc., Los Angeles, California
    "Seismic Yield Determination"

*Now at DARPA, Arlington, Virginia

15. John R. Murphy
"Body Wave Coupling Theory"

16. J. Theodore Cherry, Systems, Science and Software, La Jolla, CA
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The Influence of the Crustal Structure

17. Rhett Butler* Sierra Geophysics, Inc., Arcadia, CA
"Site Specific Propagation Effects"

18. Robert S. Hart, Sierra Geophysics, Inc., Arcadia, CA
"Factors Affecting the Estimate of m_b/Yield Bias for for Soviet Underground Nuclear Explosions

Attenuation Effects on P Waves

19. George R. Mellman and Robert S. Hart, Sierra Geophysics, Inc., Arcadia, CA
"Review of Magnitude/Yield Estimation"

20. L.J. Burdick, Lamont-Doherty Geological Observatory, Palisades, NY
"The Changing Results on Attenuation of P Waves"

21. Zoltan A. Der, Teledyne Geotech, Alexandria, VA
"Regional Attenuation Effects on P Waves"

22. Robert R. Blandford
"Regional Attenuation Effects with Special References to the Nevada Test Site"

23. Robert G. North, Lincoln Laboratory, Cambridge, MA
"Regional Attenuation Effects on P Waves"

24. Thomas H. Jordan, University of California, San Diego
"Frequency Dependence and Regional Variations of Q_{ScS}"

25. Howard C. Rodean, Lawrence Livermore Laboratory, Livermore, CA
"Regional Attenuation Effects on P Waves"

26. Charles B. Archambeau, University of Colorado, Boulder
"Regional Attenuation Effects on P Waves and Effects of Attenuation on Surface Waves"
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Surface Wave Magnitude

27. Paul W. Pomery, Rondout Associates, Inc., Stone Ridge, NY
"Definition of Surface Wave Magnitude"

* Now at CIRES, University of Colorado, Boulder, CO

28. Keith Priestly, University of California, San Diego
"Estimation of Surface Wave Magnitude"

29. Thomas D. Eisenhauer
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30. David G. Harkrider, California Institute of Technology, Pasadena
"Definition and Estimation of Surface Wave Magnitudes"

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U.S. Geological Survey

Routine Determination of Earthquake Magnitudes by the USGS National Earthquake Information Service (NEIS)

James Taggart and E. R. Engdahl

Current Methods

The technique used by the NEIS to compute body wave magnitude (m_b) has hardly changed since the early 60's (Murphy and Jordan, 1964). Short-period vertical data are reported to NEIS in terms of ground motion amplitude and period or double trace amplitude, period, and SPZ magnification at 1 second. For the latter data to be used, the frequency response of the SPZ instrument at the reporting station must be known. This latter usage is limited mainly to WWSSN stations, plus a few calibrated stations of the NEIS network telemetered to Golden. The period and amplitude data reported to NEIS are not generally monitored by us. However, we have instructed stations to report the amplitude of the largest pulse within the first five cycles of the teleseismic P or Pn. Body wave magnitudes are computed according to the formula log (A/T) + Q, defined by Gutenberg and Richter (1956), where A is the P wave amplitude in micrometers, T is the period in seconds and Q is the depth-distance factor. Body wave magnitudes are not determined for PKP arrivals, for event depths greater than 700 km, for stations having $\Delta > 109°$, for stations having $\Delta < 5°$ with an event depth greater than zero, or for stations having $\Delta < 2°$ with an event depth of zero. Magnitudes for stations whose P arrival times have residuals of greater than 10 seconds

are also not computed, primarily to remove data which may belong to another event, or which may be scaled from various crustal phases following Pn. The reported m_b is the mean of all accepted individual station values after truncation at ± 1.5 units about the mean. Individual station m_b values may also be excluded from the average on instruction by an analyst.

Long-period vertical and horizontal surface wave data are reported to NEIS in terms of ground motion amplitude and period or double trace amplitude, period, and maximum magnification. For the latter data to be used, the frequency response of the LP instruments at the reporting station must be known. Surface wave magnitudes are computed from the IASPEI (1967) *formula* $\mathrm{Log}(A/T) + 1.66\ \mathrm{Log}\ \Delta + 3.3$, where A is the maximum vertical surface wave amplitude in micrometers, T is the period in seconds, and Δ is the epicentral distance in degrees. Surface wave magnitudes are determined only for events whose focal depths are less than or equal to 50 km and for stations having $20° \leq \Delta \leq 160°$. No correction for focal depth is used in the M_s calculation. The reported M_s is the mean of all accepted individual station values computed from only vertical component data after truncation at ± 1.5 units about the mean. Individual station magnitudes computed from vectorially combined horizontal components and for reported periods $T < 18$ or $T > 22$ are also published, but not used in the average. Individual station M_s values may also be excluded from the average on instruction by an analyst.

The policy of the NEIS is to publish a PDE hypocenter and magnitudes based on limited data as soon as feasible, and later to publish monthly

listings based on much more extensive data. On January 15, 1980 the monthly listing for February 1979 was completed and the target date for catching up (allowing a 3-month lag) is August, 1980 at the current rate of two months/month.

Future Trends

In the future there may be some changes in the technique used by the NEIS to compute various magnitudes, but the observed or computed ground motion amplitude and period will continue to be reported - hence users may apply their own techniques.

We expect the NEIS to report several additional magnitudes, where applicable, in the future. Long period m_b, M_s at several frequencies, M_w, and M_m are candidates. The routine determination of most of these, as well as seismic moment, will depend upon implementation of semi-automatic processing of data from the Global Digital Seismograph Network (GDSN).

We will probably recommend that average m_b for moderate-sized, shallow earthquakes not be estimated by the NEIS from observed amplitudes at $\Delta < 20°$ unless regional attenuation functions, such as those of Evernden (1967), are available for closer distances. Even so, there is evidence that m_b attenuation in the eastern U.S. is less than that given by Evernden's EUS formula.

The USGS has plans routinely to determine focal mechanisms and phase radiation patterns semi-automatically using data from the GDSN. Application of radiation pattern corrections should reduce the scatter

of equalized amplitudes and spectral densities, except perhaps for major and great earthquakes where rupture propagation effects obscure simple patterns. The routine application of radiation pattern corrections by the NEIS presumably would initially be limited for practical reasons to earthquakes larger, say, than $M_s = 6.0$.

It would be possible to proceed another step and estimate station-path corrections for amplitude or spectral density between areally limited source regions and stations.

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Air Force Technical Applications Center (AFTAC)
Robert J. Zavadil

8 February 1980

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Attn: Dr Carl Romney

Dear Carl,

This letter is in response to your letter of 18 December requesting a statement on the two subjects discussed below. I apologize for the brevity of my comments but simply have not had the time available to do more.

I. DEFINITIONS OF BODY WAVE MAGNTIUDE:

For two decades seismologists have generally agreed on a definition of body wave magnitude:

i.e: $m_b = \log {}^{A}/_{T} + B\ (\Delta)$

Where A = Maximum amplitude within the first few cycles of the P-Wave

$B\ (\Delta)$ = A calibration function correcting for geometical spreading and attenuation with distance.

The original distance factors developed by Gutenberg[1] are still widely used. These were later modified by Vanek[2] but Vanek's are primarily used in the Soviet Bloc countries. Vieth and Clawson[3] developed a set of distance corrections in 1968 using large explosions at many different sites worldwide. This distance calibration curve of Vieth and Clawson probably represent the best universal curve available but are not generally in use today due to a reluntance to introduce a new set of corrections which result in some uncompatibility with existing data files. In addition its generally felt that when using a large number of stations the differences in the calibration curves average out. However, it would appear that the Veith-Clawson curve is significantly better and the resulting improved station magnitudes would be better suited for studies of station corrections, path effects, etc.

For measurements of magnitudes from explosions the appropriateness of dividing the amplitude by the period frequently comes into question when the effects of p^P, source spectra scaling, and period measurement problems are considered.[5] However little data has been presented which demonstrates which is the "better" approach. Another continuing problem is where to best measure the P-Wave amplitude. Most (if not all) workers agree that the measurement of the 2nd half cycle ("b" cycle) should be less contaminated than that of the maximum cycle. However, little reduced data has been presented which documents the resulting improvement.

While numerous researches have utilized spectral measurements of short-period P-Waves for a variety of studies, no concerted effort has been made which has clearly demonstrated a greater accuracy in yield estimation using spectral measurements. Part of the problem has been the lack of a sufficient amount of digital data. AFTAC is currently funding a project being done by ENSCO which is designed to attack this problem. About 100 explosions from all significant test sites have been selected and a digital data base is being developed using data from the AEDS and various high quality sources (WWSSN data are being manually digitized). The project will also include a suite of spectral measurements made on both SP and LP signals. These data will all then be available for further source estimate studies.

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3 - Veith, K. F., and G. E. Clawson, 1972, Magnitude from Short-Period P-Wave Data, Bull. Seism. Soc, Am., 6, 435-452.

4 - Basham, P. W., and R. B. Horner, 1973, Seismic Magnitudes of Underground Explosions, Bull. Seism. Soc Am., 63, 105-131.

5 - Masse', R. P., and B. G. Brooks, 1977, Measurement of Teleseismic Energy from Nuclear Explosions, AFTAC-TR- 77-17, (Classified Report)

II. ESTIMATION OF BODY WAVE MAGNITUDE:

Since the magnitudes observed from a given event are seen to be normally distributed, a simple average of the observed individual station magnitude is the accepted method of estimating event magnitude. Limiting the observations to the teleseismic range of 20^0 or $25^0$to 90^0or $100^0$is a standard procedure to avoid the effects of crust and upper mantle variations and the effects of core diffraction. In a study by Veith and Clawson[1], over 2000 stations observations from 43 explosions at 19 world wide sites in the distance range 25^0-- 90^0 showed a standard deviation of about 0.35 magnitude units about the mean. This appears consistent with personal observations of large well recorded explosions which exhibit standard deviations of a single station observation of between .30 and .35.

The selection of the observing network is very important. Besides the selection of sufficient number of well-calibrated stations tosample various distances and azimuths, care must be taken that the selected network does not truncate the sample. Truncation occurs when all stations in a selected network are not capable of recording all specific events of interest due to a lack of dynamic range. Truncation on the low side generally occurs as event falls below the detection capability of some of the network stations. Truncation on the high side occurs for large events which clip or are unreadable on some stations. Errors in event magnitude can result which approach 0.2 magnitude units due to sample truncation.[2]

Individual station corrections have been proposed from the earliest days of seismology when it was observed that certain stations always seem to have high (or low) readings. With the advent of more careful siting and calibrations, along with the recognition that many of these apparently anomalous observations were source or path related, constant station biases "grew" smaller. More recent studies have generally suggested that the constant station corrections are generally less than 0.2 mag units. However emperically derived values are strongly affected by the concentration of data from highly seismic regions and there is still uncertainty concerning the effectiveness of such corrections if applied to various aseismic regions. (I don't know if anyone has even simply applied a set of standard earthquake derived constant station corrections to a significant number of Soviet explosions sites to measure the reduction in the standard deviation). While for well recorded events the mean of any constant station corrections should be near zero and will be taken care of in the network averaging process, for events with fewer stations valid station corrections could offer a significant improvement.

Source-station corrections derived from explosions are commonly used and have been demonstrated to reduce the standard deviation to about 0.15 (from the .30 to .35) over a local area (10 - 20 Km). Earthquakes have also been used to develop source-station corrections and suggest a reduction in standard deviation to about 0.25 over a 1 - 2 region.[3] However, such regionalization using earthquakes runs the risk of resultant station corrections which reflect a particular source function rather than a station term. If so, such correction would obviously be inappropriate for use in estimating magnitudes from explosions.

1 - Veith, K. F., and G. E. Clawson, 1967, Attenuation of Short-Period P-Wave Amplitude with Distance, Geotech TR 67-58 (Classified Report)

2 - AFTAC, 1977, Surface Wave Yield Estimation and Research, TR 77-37 (Classified Report)

3 - Frye, W. H., 1970, Source Region/Station Residuals for Selected Regions of the Sino-Soviet Bloc, Geotech, TR 70-26.

I hope the above is of some value to you, and again apologize for the lack of completeness.

Sincerely,

Robert J. Zavadil

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11 January, 1980

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From: K. F. Veith

Subject: State-of-the-Art Assessment: Seismic Yield Determination

1. Definitions of Body Wave Magnitude

The early work of Gutenberg and Richter (e.g. Gutenberg and Richter, 1942, 1956a, 1956b, and Gutenberg, 1945a, 1945b, and 1945c) relates the source energy to the amplitude of a sinusoidal wave traveling through the earth by

$$E = \frac{A}{T} f(\Delta)\ g(\sin i) \qquad 1.$$

where E is the source energy,
A is the observed amplitude,
T is the observed period,
$f(\Delta)$ is the geometric spreading and attenuation function, and
$g(\sin i)$ is the energy partitioning function from acoustic boundaries.

Magnitude was defined to be a measure of the source energy according to

$$m = a \log(E) + b \qquad 2.$$

where m is the earthquake magnitude, and
a and b are proportionality constants.

This relates magnitude to amplitude by

$$m = \log(\frac{A}{T}) + Q(\Delta,h) \qquad 3.$$

where $Q(\Delta,h)$ is the adjusted proportionality factor which incorporates a specific earth model to absorb the $g(\sin i)$ term, and (a) is defined to be unity.

In normal usage, A is taken as the vertical component of the signal and Q is adjusted for the normal emergence angle (which is also a function of Δ and h).

In practice, Gutenberg developed the Q factors from theory, tied them to his shallow source observations, realized that his theory was probably only first order and made adjustments to the Q factors for observed distance and depth variations (after removal of station effects). One may seriously question the distribution and quality of the data used because they were

from earthquakes of $m_b \geq$ 7.0 which were recorded between 1900 and 1952. Therefore the earthquakes were probably complex sources and the observing stations were few and poorly distributed.

Regardless, the basic formulation (3) makes several assumptions for its validity.

1. The maximum amplitude and its corresponding period measured from the seismogram are analogous to the amplitude and period of an isolated frequency; that variations in the frequency content do not affect this measured ratio.
2. Complex sources or multipathed waveforms do not affect the ratio.
3. Regional variations in structure and attenuation either are not significant, or may be adjusted for by a constant station correction.
4. Attenuation has a constant proportionality to the distance the seismic ray travels. There is no difference in attenuation in the various regions or materials within the earth (Gutenberg, 1958,1959). This is at odds with modern theories of attenuation (Anderson and Archambeau, 1964, and Knopoff, 1964).

Questions as to the reliability of Gutenberg and Richter's Q factors were settled when Veith and Clawson (1972) developed a revised set of Q factors (P factors) from an extensive set of explosion data. While these authors recognized regional variations in attenuation factors, the only effects on the definition of surface event magnitudes is to allude to a partial cause of the station variations and to indicate that corrections may need to be distance and/or azimuthally dependant.

Addressing the assumptions leads into the various types of magnitude definitions which have been proposed. While equation (3) is applicable to many seismic phases (with appropriate changes in Q), the following discussion will concentrate on magnitude from P phases.

The maximum amplitude pulse and associated period observed in the first few cycles of a P arrival is actually the integration of data with many basic frequencies and associated amplitudes. Seismometer systems which vary in frequency response must be expected to "observe" signals of varying shapes which cannot be directly adjusted by a time domain correction for the response curve. Insofar as this effect is a characteristic of the station instrumentation and the transmission properties of the earth beneath the station, it may be removed as part of an empirical station correction factor. However, stations on or near boundaries of great contrast in tectonic regimes, or stations on island arcs may be expected to observe gross differences in frequency content of their arrivals with a corresponding need for complex station corrections to provide consistent magnitude estimates (see Byerly, Mei, and Romney, 1949).

Spectral averaged magnitudes (e.g. Chandra, 1970, and Howell et al, 1970) could indeed yield a better estimate of the energy content of the arrival, but they are subject to several problems. One concerns the window size and type which is used in obtaining the spectral estimates. Veith (1978) has shown that typical windowing functions such as cosine and Parzen windows can have significant effects upon the calculated spectra. A second is the difficulty in removing multiples from the data, particularily for the critical shallow events.

The energy from multiple arrivals is extraneous to the magnitude estimate and will tend to have a greater effect upon the spectral estimates because of the normally longer time frame utilized by such estimates.

Significant distortion of both amplitude and period is observed because of multipaths generated within the crust and upper mantle beneath many stations. Examples are the peculiar double peak observed from explosions at the Alaskan station near Burnt Mountain which is not observed at any other Alaskan station, and the great variation in signals observed across LASA (Mack, 1969). Attempts to reduce magnitude scatter by eliminating the period from equation (3) are simply assuming that the true dominant period should be constant and are acknowledging that the waveform is too complex to measure it properly. Real shifts in the observed period can occur and represent regional transmission characteristic variations. They should be treated as station effects with corresponding station corrections.

Naturally the use of P phases at regional distances requires the use of modified formulations which reflect the actual structure of the region rather than the "average" earth model of Gutenberg and Richter. Strictly speaking, the lack of agreement between the regional curves requires a discrepancy in the teleseismic curves because it reflects variations in the source region transmission and attenuation characteristics. Regional curves have been given by Evernden (1967) and Swanson (1979), among others. Regional phases have also been suggested for use at corresponding distances with varying degrees of success. Baker (1970) used Lg recorded at LRSM stations from U.S. explosions and found that the data had less scatter than Pn. Swanson (1979) found similar results for Sn, Lg, and LR in southern Africa from both earthquakes and rockbursts but equal scatter from these phases from earthquakes in South America.

2. Estimation of Body Wave Magnitude

At the present time, it would appear that regionally dependent station corrections would be the most consistent way of estimating body wave magnitudes. Without such corrections, the standard deviations of magnitude estimates from both explosions and earthquakes can be expected to be near 0.35 magnitude units (Veith and Clawson, 1972). Spectral techniques may reduce this scatter somewhat, but the results will be highly variable from station to station. Regional phases will yield standard deviations ranging from 0.25 to 0.40 units from earthquakes with the degree of scatter depending upon the complexity of the regional geology (Swanson, 1979).

A relatively extensive study of magnitude variations by Veith and Clawson (1976) yielded regional station corrections which reduced the scatter in earthquake magnitudes in ten regions of the USSR from a standard deviation of 0.35 to values between 0.25 and 0.31 magnitude units. No attempt was made however, to either reread the amplitude values or to estimate and correct for source mechanisms. It is expected that precise rules for measuring A/T and the elimination of source mechanism effects (i.e. radiation pattern corrections) could easily yield scatter reductions below 0.20 magnitude units. It is also expected that the use of digital waveform analysis may reduce some of the human analysis variations and errors which undoubtably are present in any large body of data.

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AFTAC

BODY WAVE MAGNITUDE DEFINITIONS

Thomas D. Eisenhauer

Several deficiencies in contemporary methods for determining m_b can lead to errors in yield estimates. The practice of measuring the maximum amplitude on seismograms recorded from systems peaked at high frequencies to enhance detection, is not appropriate in light of studies of the frequency dependence of displacement at the elastic-inelastic boundary (Murphy, 1977). These studies suggest that amplitudes at high frequencies should not relate to yield as consistently as amplitudes at lower frequencies. Further, recent studies by Sierra Geophysics indicate that both Q and amplification due to layering cause greater scatter in amplitudes at high frequencies. The current practice of dividing amplitude by period tends to magnify the problems caused by measurement of amplitudes at high frequencies. The measurement of amplitudes from networks of stations which have different system responses (e.g., LRSM and WWSSN) can be another source of magnitude error. My recommendation on magnitude definitions will include comments of data, measurements, distance normalization, corrections and network considerations.

Data. The data should be in digital form. As long as the data are calibrated and have adequate dynamic range, filtering can be done to reconcile differences in frequency response for various seismographic system. The stations which comprise the network should have noise levels low enough to insure that the low side of the amplitude distribution is not truncated. At least 30 stations well distributed in azimuth and teleseismic distances should be used to obtain magnitudes for explosions at each test site. Fewer stations (10 - 15), with good azimuthal and distance distribution will provide adequate data for good relative magnitudes for explosions within a test site.

Measurements. Both time and frequency domain measurements should be evaluated to settle on the best type of measurement. I recommend that the instrument response be removed for frequencies greater than about 0.5 Hz and that the time series amplitude be measured within the first 3/4 cycle (b amplitude). Measurements from data with higher cutoff frequencies will be necessary when S/N is a problem. Measurements from time series with flat response eliminates the need for a precise period measurement to correct for system response. Measurement of the b amplitude reduces the potential for interference due to P^P or spall. Spectral amplitudes should be averaged over a frequency band from about .5 to 1.2 Hz.

Again, a spectral amplitude at higher frequencies will be necessary when S/N is a problem. The spectra should be computed with the shortest time window (gate) necessary to resolve the amplitudes at frequencies of interest. It should be noted that P^P and spall could strongly influence the spectral amplitudes.

Distance Normalization. Part of the scatter in magnitudes is due to errors in normalization for distance. If new measurements are made, new distance normalization values may have to be developed. As a minimum the amplitudes for an explosion at each test site should be normalized to one magnitude and evaluated as a function of distance. As a starting point I recommend that the P factors of Veith and Clawson, 1972) be used to compute the initial magnitudes. The velocity structure of the earth and the average Q for body waves can be used to compute the shape of the distance normalization curves as a check on the observations.

Network Magnitudes: Some sort of averaging of individual station magnitudes is necessary to obtain a single estimate of magnitude. Problems are encountered when it is necessary to use different networks in order to have sufficient observations for the various test sites. Differences in attenuation and layering beneath the stations could cause some difference in magnitudes with the same yield, coupling and regional attenuation characteristics. This problem can be alleviated by evaluating a matrix composed of explosions at different test sites and magnitudes at all available teleseismic stations. The matrix can be solved to obtain magnitude corrections for each station and magnitudes for each event. The event magnitudes would, on average, be equivalent to the magnitude obtained if each event were observed at all stations. Similar procedures are used to develop magnitude corrections for multiple events at a single test site. The matrix method for combining data from different networks would compensate for anomalous attenuation due to variations in Q.

Corrections: Magnitude corrections have been proposed for the source region, station and P^P (Marshall et al, 1977). These corrections were developed for amplitudes measured from analog records. Measurement of amplitude at lower frequencies and matrix methods for computing magnitudes from different networks may eliminate the need for source and station corrections. With digital data the effects of P^P should be taken into account when the measurements are made. No matter how carefully the measurements are made and the distance normalization is handled there will still be scatter in the magnitudes due to propagation effects which cannot be accounted for. Consequently the practice of developing magnitude corrections for explosions within a test site will still have merit. If it is necessary to compute magnitudes from amplitudes measured at high frequencies it may be desirable to correct the magnitude to the equivalent magnitude at low frequencies. Quantitative estimates of the corrections will have to come from future studies.

Estimation of Body Wave Magnitude
Robert Blandford
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I assume that the body wave magnitude estimation procedures of interest are those of the AEDS network. I have no first-hand knowledge of these procedures, however I have heard that use is made of a large number of WWSSN stations to determine an average magnitude for a shot at two different sites and that then station corrections for each site are defined so that each individual AEDS station yields this magnitude.

Let me first comment on how the calibrating WWSSN network should be distributed. In Figures 1 and 2 from von Seggern (1977) may be seen the same 23 WWSSN stations arrayed around the events Boxcar and Milrow. To the extent possible the stations are in the teleseismic range and evenly distributed in azimuth. Even in that study there is, no doubt, too great a concentration of stations in the "northeast" quadrant. Concentrations e. g. in the United States and Europe must be avoided. An even distribution is crucial because Chang and von Seggern (1977) have shown at LASA that there are substantial azimuthally varying magnitude biases which, however, average out azimuthally to only a few hundreths of a magnitude unit. (Chang and von Seggern trace these magnitude variations to focussing and defocussing in the mantle; so there is no guarantee that the biases will average to zero, in fact the contrary should there be a suitable "lens" structure in the mantle). Substantial variations in the azimuthal terms between LASA subarrays in the study by Chang and von Seggern also show that calibrations may need to be recalculated for test sites only 50 km apart.

As much as possible the WWSSN network should be in common between the two test sites to avoid effects of crustal amplification due to differing surface crustal velocities. For LRSM stations Der et. al. (1977) have shown that large magnitude effects may be traced to these differences. If the stations cannot be in common, then only hard rock sites should be exchanged for each other and if this is not possible, then crustal effect corrections must be applied.

When the problems of equal azimuthal distribution, station commonality and crustal corrections have been correctly handled for the calibrating network, then we must return to questions of possible bias for subsequent events at the same test site, even when station corrections are applied. Ringdal's maximum likelihood estimation procedure, see Ringdal (1976), von Seggern and Rivers (1978), Ringdal (1978) must be applied for the cases of clipped signals and signals below the noise. If there is variation in the signals then it is always possible that the signal was not detected at a station because of a "blind spot" to that station from a new location within the test site; and Ringdal's procedure gives the proper approach for avoiding bias in this situation.

There are other considerations which might be followed in an ideal system which might be better discussed under the Topic "Definitions of Body Wave Magnitude". I feel that most of these considerations are less important than those discussed above, and so I will only mention them.

Ideally a single cycle would be measured on a common instrument and not be corrected for or divided by period. This would avoid the error, on the order of 10% introduced by period measurement.

Calibration events should be close in yield to those events of interest in order that system and earth response variation with period does not bias the results.

The amplitude and delay of pP should be estimated by the method of maximum likelihood (Shumway and Blandford, 1977) and the magnitude appropriately adjusted. Due to the general prevalence of small reflection coefficients this should be an effect on the order of less than 0.1 m_b.

In discussing the expected variances of well-recorded events . . . it is difficult for me to discuss what might be obtained with present procedures within which might be imbedded biases of various sorts; so, assuming that the procedures outlined above are followed, let me outline a sequence of scenarios.

1. Explosions of equal yield are detonated at equal depth in the same medium over a region 50 km in diameter similar to the region in Montana where LASA was located. Then the calibration procedure suggested above is applied.

Then the study of Chang and von Seggern (1977) applies, see Figure 3. The standard deviation at an individual station using reciprocity would be about 0.4 and the standard deviation of the mean would decline by $\sqrt{N}$ where N is the number of measuring stations. This would be the relative precision of estimated yields within that test site.

2. Two identical test sites of the above type which are, however at substantially different azimuths from most of the detecting stations, e. g. Semipalatinsk and a region around RKON.

For this case we need the result for a separation in Figure 3 not of 60 kilometers but of, say, 6000 kilometers. One guess for such a number would be to assume that the fluctuations in teleseismic magnitude about the mean are truly random. Then if the standard deviation is the usual 0.35 m_b the corresponding number would be $\sigma \sqrt{2} \times 0.35 = 0.5$ m_b. This is in agreement with calculations the same as those required for Figure 3 but applied to von Seggerns' (1977) m_b data for Boxcar and Milrow. These yield $\sigma = 0.56$; more data of this sort are needed. Then we have the result that the relative yield would be determined as $0.56/\sqrt{N}$, only slightly worse than for tests 50 km apart. Note that N cannot, probably, be usefully increased above 20 before we begin to oversample the "pattern" emerging from the source and obtain correlated magnitudes.

3. Finally, if there are questions of shot medium, effects of pP or effects of absorption in the upper mantle; then these effects must be estimated and allowed for.

Brief Comments on Unassigned Topics

Robert Blandford
Teledyne Geotech

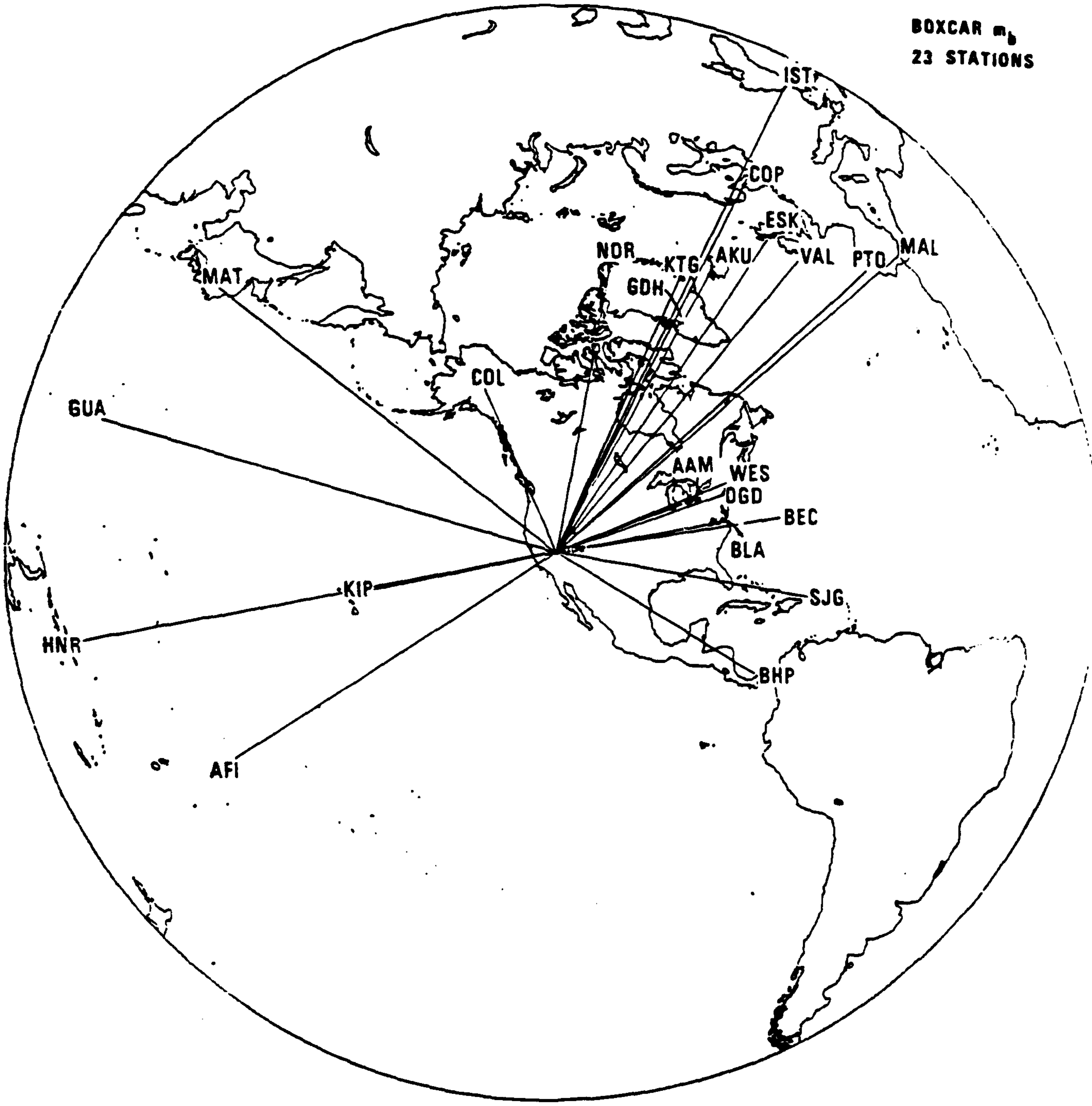
Definitions of Body Wave Magnitude

von Seggern (1977) has defined a spectral magnitude as the integral over the first 6 seconds of the instrument corrected displacement spectrum in the frequency band 0.8-1.4 Hz. The technique was compared to conventional calculations for 100 events at ANMO and good agreement was found. This work was done in order to develop a suitable m_b measurement for automatic processing with the NEP system.

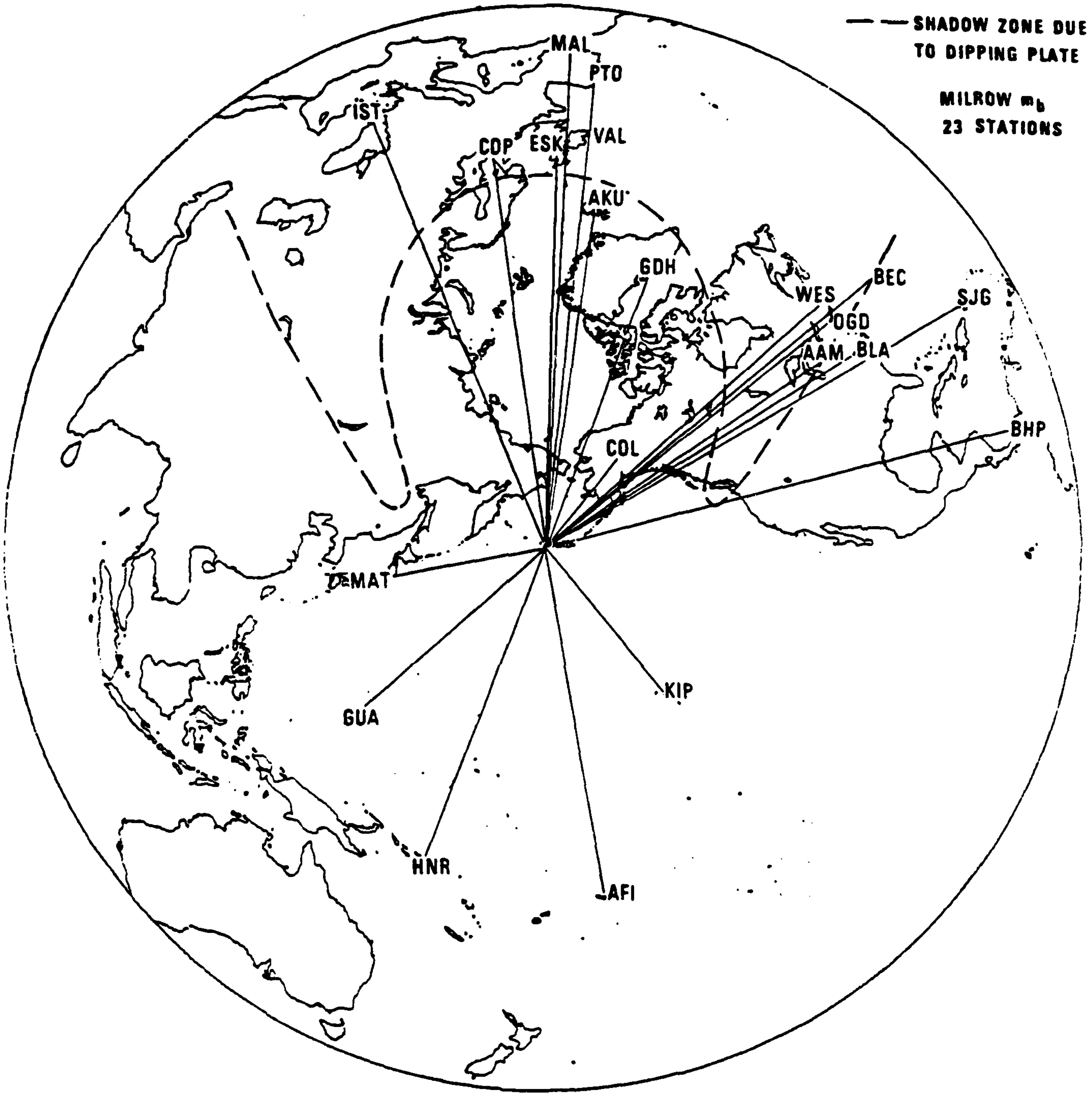
In a number of studies carried out at the SDAC the a, b, or c phase corrected or not for period were measured. In the narrow range of magnitude interest, 5.0-6.0 no significant difference has ever been noted so far as magnitude determination was concerned, although effects of pP, etc., can be detected.

von Seggern, D. W., 1977, Methods of automating routine analysis tasks in preparing a global seismic bulletin, TR-77-13, Teleldyne Geotech, Alexandria, Virginia 22314.

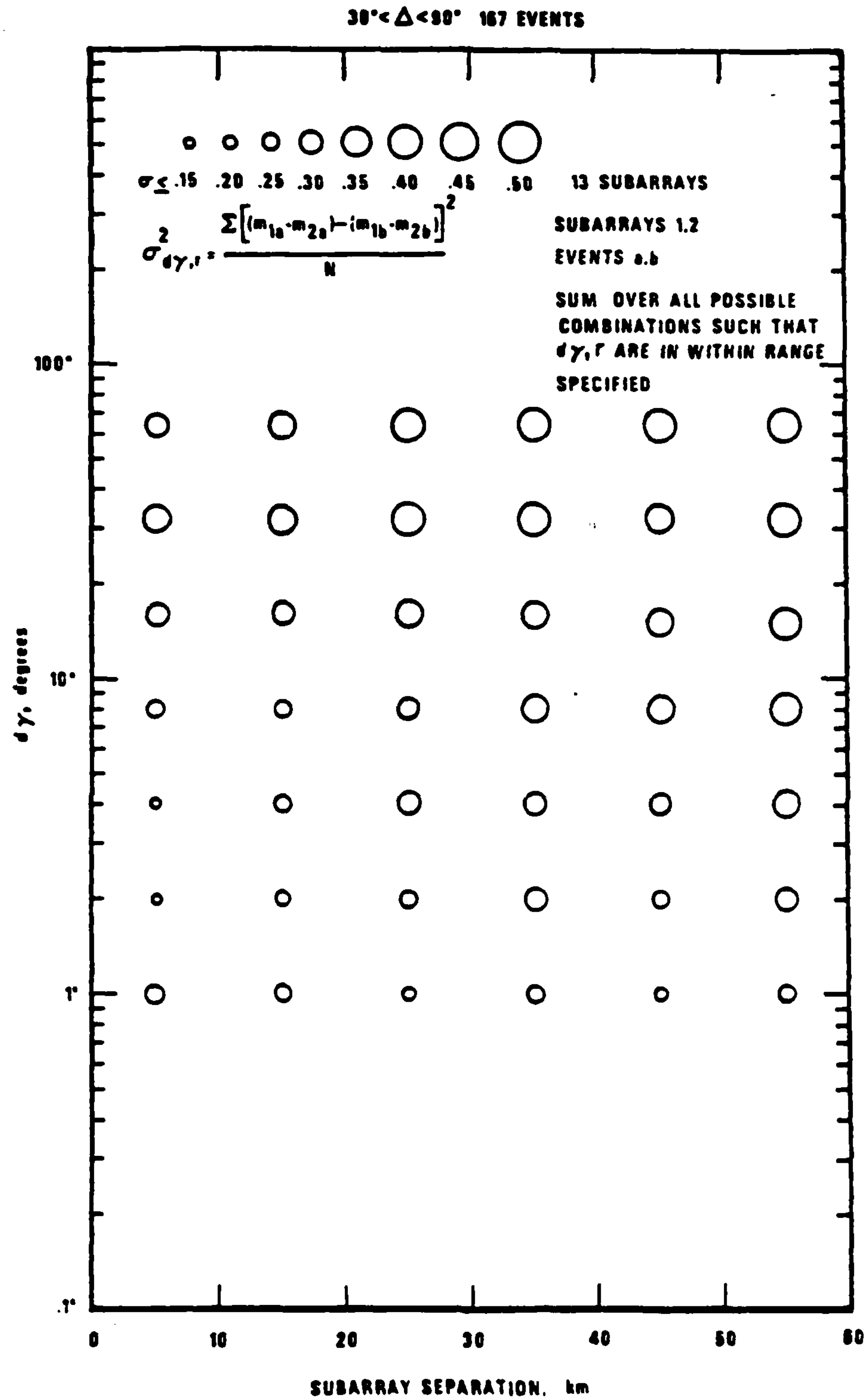
The only way I can see to reduce the statistical fluctuation is to obtain fundamental causal knowledge of the focussing and defocussing beneath source and receiver.



Locations of WWSSN stations used to estimate BOXCAR m_b (90° equidistant azimuthal projection from the Nevada Test Site).



Locations of WWSSN stations used to estimate MILROW m_b (90° equidistant azimuthal projection from Amchitka Island).



Standard deviation of the imprecision of calibration, $\sigma(\Delta m_b)$, expressed in terms of subarray separation and the central angle between two rays impinging at LASA.

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1. DEFINITIONS OF BODY WAVE MAGNITUDE

by

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6

Introduction

Body wave magnitude, m_b, is an important single parameter used to describe an event recorded at many stations. Conventional m_b is based on direct measurement made by an analyst on analog playouts of the data and includes a correction for the response of the seismometer at the apparent period of the phase measured. When digital data are available, as is increasingly the case, this procedure is unnecessarily cumbersome and prone to error.

We have done some work at Systems, Science and Software (S^3) to define better methods for determining m_b from individual station recordings. These "station" m_b must then be combined in some statistical way to determine an "event" m_b, but we have not been especially concerned with that.

First, we will describe a semi-automated procedure that preserves conventional ways to measure m_b, but essentially eliminates measurement errors and systematic errors due to mixing recordings from different seismometers.

We have also been developing and testing a fully automated method for determining a spectral magnitude we call $\hat{m}_b$. This represents a more radical departure from current practice, but the results certainly indicate that it should be seriously considered.

A Semi-Automated Time Domain m_b

If digital data are available, we suggest the following procedure for determining a standard time domain m_b. First, a standard seismometer response is selected. All seismograms are filtered to appear as if recorded by this seismometer. We will show examples that indicate systematic errors of as much as 0.2 m_b units can result from mixing seismometer responses. The time and amplitude of peaks within a selected time window are then determined automatically by a parabolic fit to a moving three-point window. The peaks to be used for m_b are then selected by an analyst and the automatically determined amplitude and period is used to compute m_b.

Bache, Day and Savino (1979) and Bache (1979) give some interesting examples of the application of this algorithm to recordings of eleven Pahute Mesa explosions at several teleseismic stations. The HNME results are illustrative. Five of these events were recorded by the 18300 seismometer while the others were recorded by the KS36000. The response curves are plotted in Figure 1. The data are shown in Figure 2 as they were recorded. The seismograms are replotted in Figure 3 after filtering the 18300 recorded events to appear as if recorded by the KS36000. The effect is to remove some of the high frequency details.

In Table 1 we list two estimates for m_b for the KS36000 recorded events and three estimates for the others. The SDAC values are taken directly from SDAC event reports.

The S^3 m_b measurements were made using the semi-automated procedure. For the five events with m_b from both instruments, the differences are striking. The T from the 18300 recordings are 0.14 seconds shorter, on the average, than the T from the KS36000 records. As a result, the m_b are an average of 0.14 units smaller. The differences are greatest for STILTON and

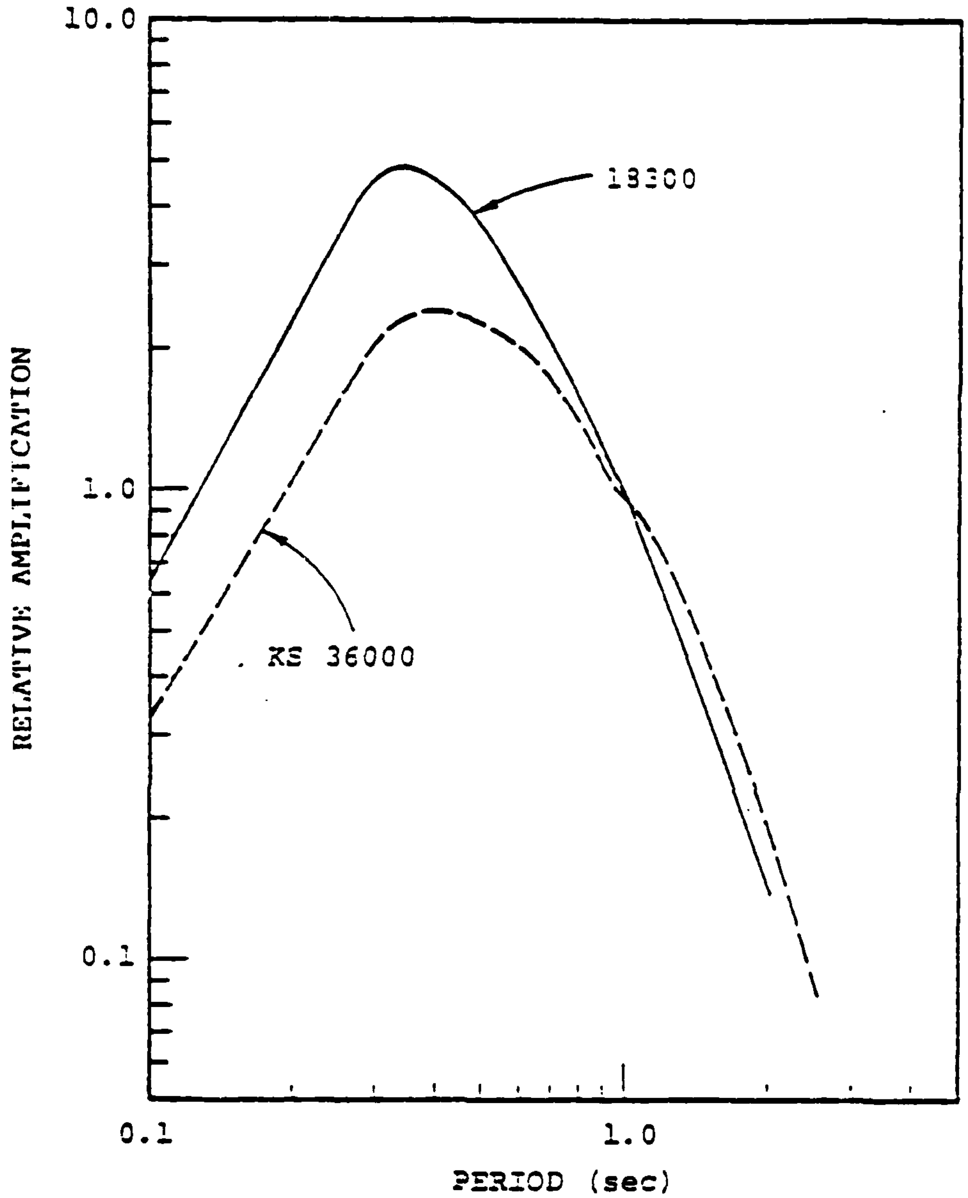


Figure 1. Relative amplitude response for two instruments used at the HNME site.

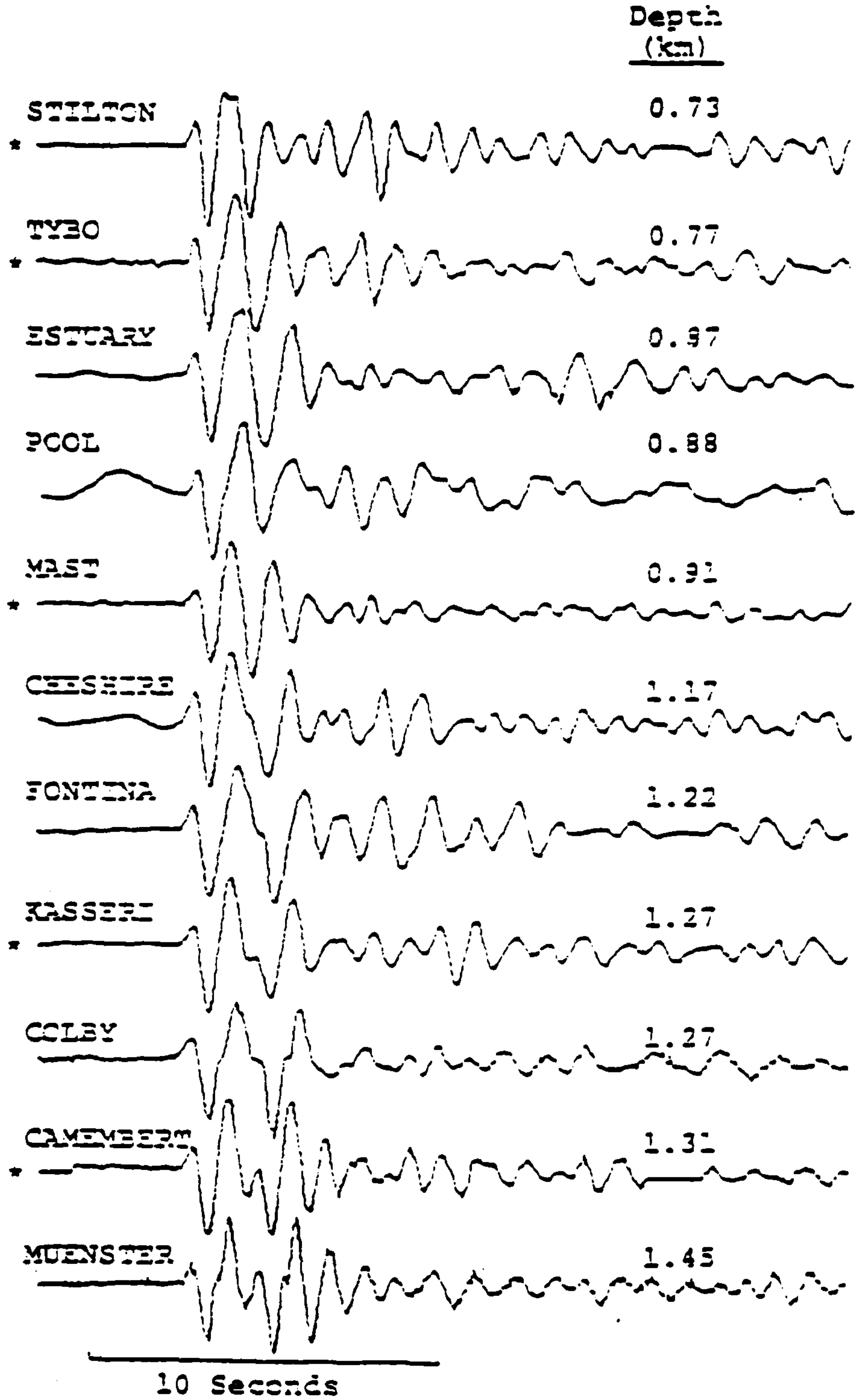


Figure 2. The ENME recordings of eleven Pahute Mesa events are arranged according to increasing depth. The asterisk denotes events recorded by the 18300 seismometer.

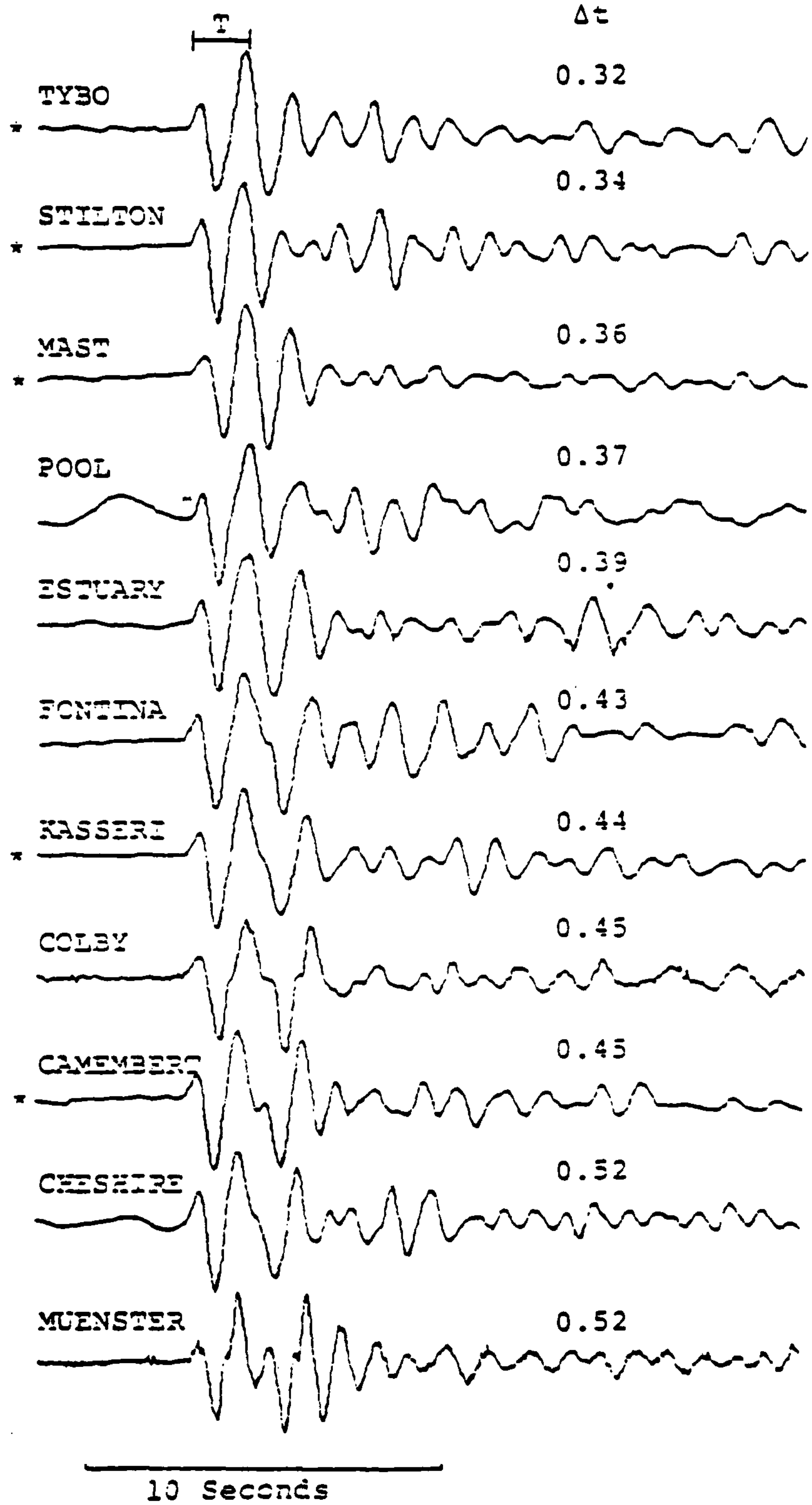


Figure 3. The HNME recordings are arranged according to increasing source-surface travel time. All 18300 recorded events (marked with an asterisk) have been filtered to appear as if recorded by the KS36000 instrument.

TABLE 1

CONVENTIONAL m_b for HNME RECORDINGS*

	SDAC** Data		S^3 Measurements 18300		KS36000	
Event	m_b	Period	m_b	Period	m_b	Period
STILTON	5.55	0.7	5.60	0.8	5.86	1.1
POOL	6.37	1.3			6.29	1.3
ESTUARY	6.25	1.5			6.07	1.3
TYBO	5.37	1.4	6.20	1.2	6.26	1.2
MAST	6.21	1.1	6.10	1.0	6.25	1.1
CHESHIRE	6.03	1.0			6.02	1.1
CAMEMBERT	6.25	1.0	6.24	1.0	6.37	1.1
MUENSTER	6.39	0.8			6.49	0.9
COLBY	6.38	0.9			6.50	1.3
KASSERI	6.46	1.0	6.50	1.1	6.59	1.2
FONTINA	6.48	1.3			6.43	1.3

*All period measurements were made from the first peak to the second peak as shown above the TYBO record in Figure 3. The amplitudes were measured from first trough to first peak.

**From SDAC event reports -- authors: J. R. Woolson, K. J. Hill, D. D. Solari, M. S. Dawkins, M. D. Gillespie, R. R. Baumstark, R. J. Markle, D. J. Reinbold.

the reason can easily be seen by comparing the two waveforms (Figures 2 and 3). However, the average differences for the other four events are 0.10 seconds and 0.11 m_b units, still quite large.

The differences between our (mixed instrument) m_b and those given by the SDAC reports can mostly be explained by differences in the period. Our period measurements are very accurate since they were done automatically. A major difference occurs for TYBO where the amplitude in the SDAC report must be in error. Using copies of the station logs and digital playouts of the calibration steps, we recalibrated all the data. Thus, the gain we used is probably not identical to that used by SDAC. Ignoring TYBO, the differences between the SDAC m_b and ours obtained from recalibrated data with our procedure are between -0.18 to 0.12 m_b units.

An Automated Magnitude Measure, $\hat{m}_b$

A major product of the S[3] research program is the MARS signal analysis program. This program is based on the application of a series of Gaussian narrow-band filters to the data. Applications include the following:

1. Determination of phase and group velocity dispersion of surface waves. This capability was used in the work described under Topic 13.

2. Detection. MARS was implemented as a P wave detector during the VSC conducted discrimination experiment.

3. Discrimination. The MARS program computes high and low frequency spectral estimates called $\bar{m}_b(f)$. The discriminant used by S[3]

in the discrimination experiment is based on comparison of these $\bar{m}_b(f)$ values with earthquakes and explosions falling in different portions of the plane.

A natural extension of this work is to use MARS to automatically provide the magnitude needed for yield determination. Ultimately, the program could automatically detect, discriminate and estimate yield.

Bache (1979) and Bache, Day and Savino (1979) proposed a particular algorithm for determining a MARS based magnitude called $\hat{m}_b$. This algorithm was tested by processing recordings of eleven Pahute Mesa explosions from six teleseismic stations. The $\hat{m}_b$ values are compared to the m_b determined by the semi-automated procedure described in previous paragraphs. An important aspect of the comparison is via a linear regression on log yield. We conclude that the $\hat{m}_b$ is at least as good a magnitude measure as the most carefully determined time domain m_b for the high signal/noise data processed.

Calculation of $\hat{m}_b$, Some Illustrative Examples

The $\hat{m}_b$ algorithm used in the reports by Bache (1979) and Bache, *et al*. (1979) worked very well, but was primitive because it failed to account for the presence of seismic noise. The discrimination experiment work led to the development of more sophisticated algorithms for using MARS output to determine spectral amplitude (Masso, *et al*., 1979; Savino, *et al*., 1979). These algorithms incorporate corrections for the presence of interferring phases and for seismic noise.

We now propose a slightly altered algorithm for computing $\hat{m}_b$. The properties of this new $\hat{m}_b$ should be nearly the same as the properties of the $\hat{m}_b$ used in the studies by

Bache (1979) and Bache, *et al*. (1979). The computation is done as follows:

1. The seismogram is processed by MARS to determine the peaks of the narrow band filter envelope functions.

2. The peaks are processed by the MARS detection algorithm to identify one or more undispersed P wave arrivals. These two steps are precisely those used by Savino, *et al*. (1979) in the discrimination experiment.

3. A particular frequency is selected for determining $\overline{m}_b(f)$. For discrimination $\overline{m}_b(f)$ is computed at a high (e.g., 2 Hz) and a low (e.g., 0.5 Hz) frequency. We will compute $\overline{m}_b(f)$ in exactly the same way at f = 1 Hz and call this value $\hat{m}_b$.

We demonstrate the $\hat{m}_b$ algorithm by applying it to a synthetic seismogram with superimposed seismic noise. The particular noise sample used is from the AI data set for RKON. Three seismograms were constructed based on the ratio of the largest peak on the synthetic to the largest peak on the noise sample. This ratio was 100, 3 and 1, respectively.

In Figure 4 we show the results of processing the peak synthetic/peak noise = 100 case. The seismogram is shown at the top. The asterisk indicates the $\overline{\overline{t}}_g$ defined by Savino, *et al*. (1979). It represents the noise weighted mean arrival time of the detected signal.

The basic information used to compute $\hat{m}_b$ is shown at the bottom. The heaviest line is a plot of log $(A \cdot f_c)$ versus the filter center frequency. The A is the filtered signal

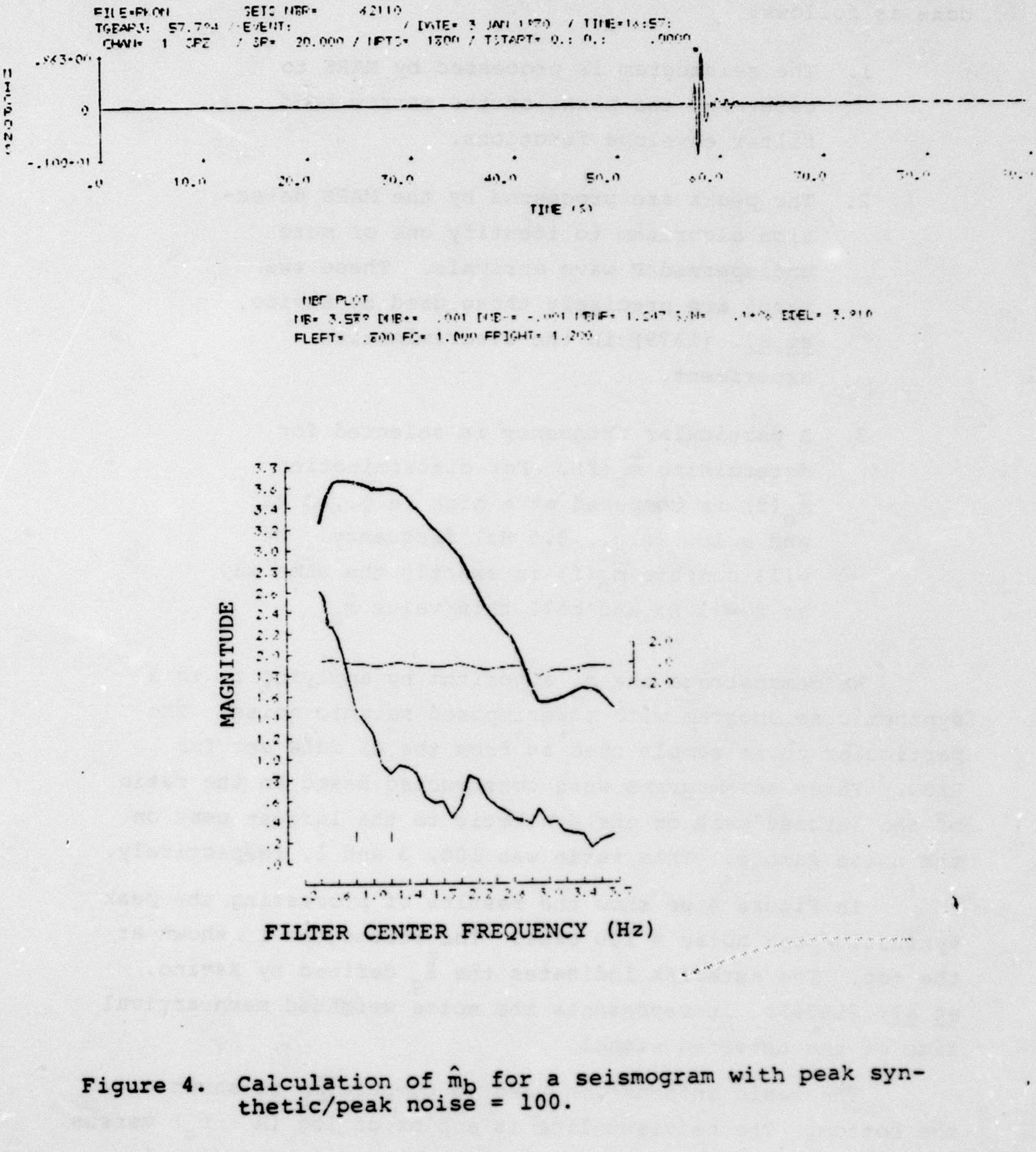


Figure 4. Calculation of $\hat{m}_b$ for a seismogram with peak synthetic/peak noise = 100.

amplitude computed by MARS. Each amplitude has an associated arrival time, t_g, which is plotted as the lightest line. The t_g scale is at the right with zero being the time of the asterisk. The amplitude scale at left is in magnitude units, since the standard Gutenberg-Richter distance correction has been added to log $(A \cdot f_c)$. The lower line is the log $(A \cdot f_c)$ for the noise sample in m_b units. The two vertical lines near the bottom mark FLEFT and FRIGHT, the limits of the band used to compute $\hat{m}_b$ at FC = 1 Hz from the best fitting (least squares) parabola to the five values in this band.

The main information about $\hat{m}_b$ appears with the graph. The MB is $\hat{m}_b$. The square of the signal/noise at 1 Hz is denoted by S/N and BDEL is the distance correction for MB. The noise introduces some statistical uncertainties into the $\hat{m}_b$ and these are given as DMB+ and DMB-.

The $\hat{m}_b$ for this seismogram is 3.589. This may be compared to the time domain m_b of 3.776, which is computed from m_b = log (A/T) + 3.61 with T = 0.72 seconds and A being the peak-to-peak amplitude. Systematic differences between the time domain m_b and the spectral measure $\hat{m}_b$ are expected. But the $\hat{m}_b$ is believed to be a more convenient and consistent measure of the spectral energy in the P wave.

In Figure 5 we show the Fourier spectrum for the synthetic seismogram (without any noise) plotted with the MARS determined spectrum from Figure 4. The MARS spectrum is simply a smoothed version of the Fourier spectrum over most of the frequency band, which demonstrates the accuracy of the MARS processing.

In Figure 6 we compute $\hat{m}_b$ for a seismogram with peak synthetic/peak noise = 3. The $\hat{m}_b$ is 3.583, only 0.006 different from the $\hat{m}_b$ computed with much less noise. The uncertainty limits are appropriately larger.

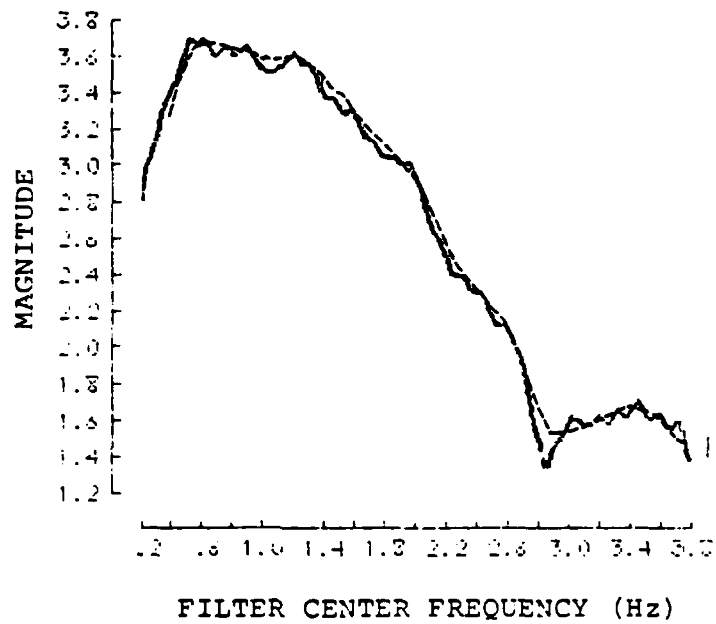


Figure 5. Comparison of the Fourier velocity (amplitude times frequency) spectrum to the MARS log (A · f_c) spectrum (dashed line). The amplitude is in m_b units.

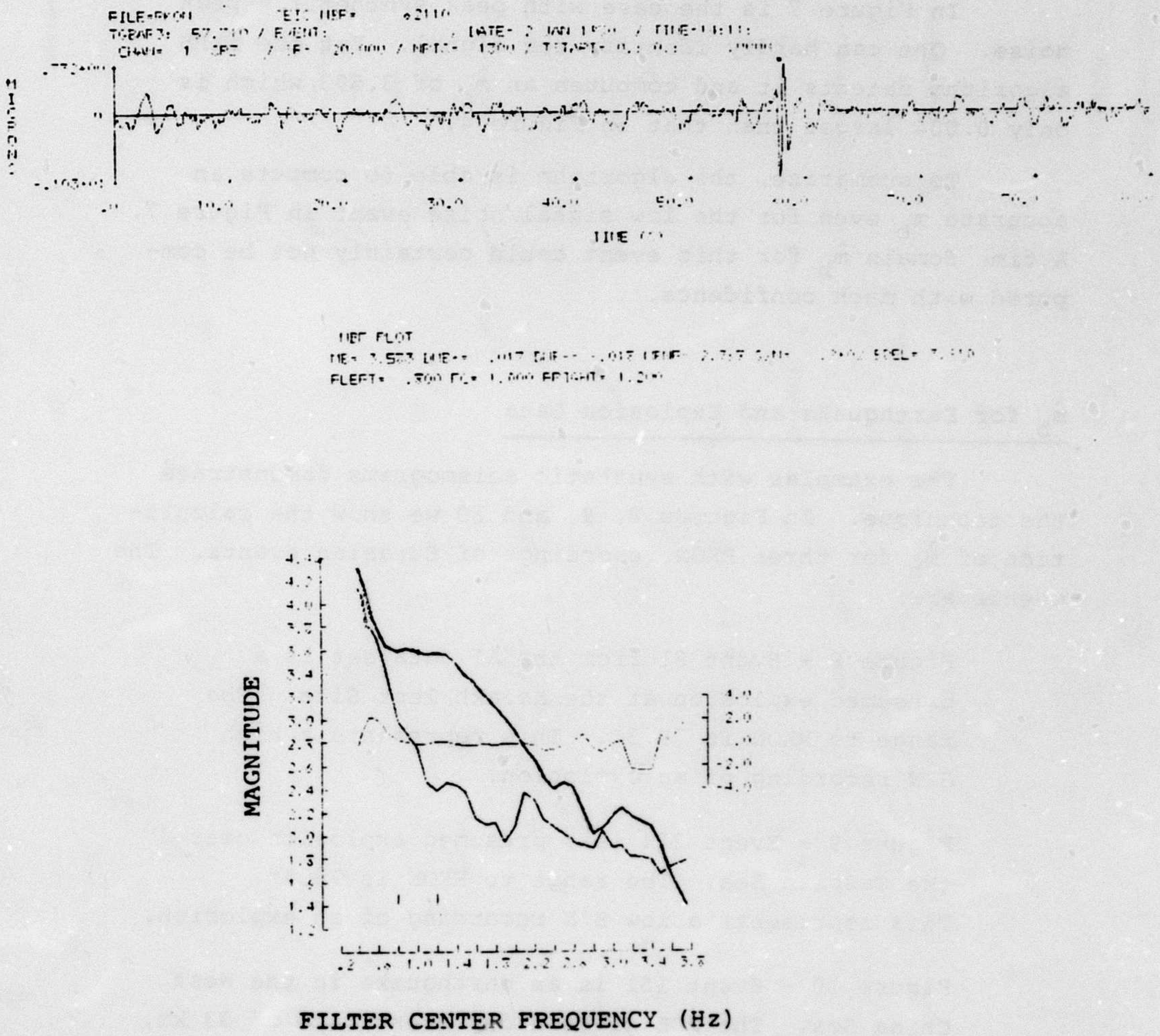


Figure 6. Calculation of $\hat{m}_b$ for a seismogram with peak synthetic/peak noise = 3.

In Figure 7 is the case with peak synthetic = peak noise. One can hardly identify the signal. Yet the MARS algorithm detects it and computes an $\hat{m}_b$ of 3.593, which is only 0.004 larger than that in Figure 4.

To summarize, the algorithm is able to compute an accurate $\hat{m}_b$, even for the low signal/noise event in Figure 7. A time domain $\hat{m}_b$ for this event could certainly not be computed with much confidence.

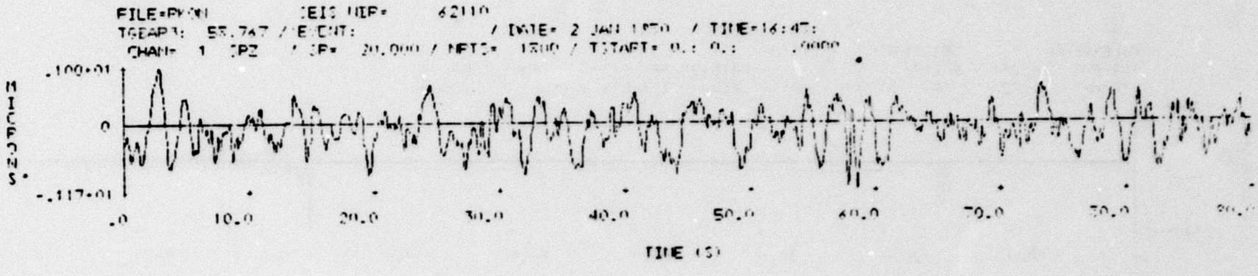
$\hat{m}_b$ for Earthquake and Explosion Data

The examples with synthetic seismograms demonstrate the technique. In Figures 8, 9, and 10 we show the calculation of $\hat{m}_b$ for three RKON recordings of Eurasian events. The events are:

Figure 8 - Event 81 from the AI data set is a presumed explosion at the Kazakh Test Site. The range to RKON is 79.3°. This represents a high S/N recording of an explosion.

Figure 9 - Event 274 is a presumed explosion near the Caspian Sea. The range to RKON is 76.4°. This represents a low S/N recording of an explosion.

Figure 10 - Event 151 is an earthquake in the West China Sea. The PDE gives a depth estimate of 33 km, indicating that it was shallow. The range to RKON is 93.5°. The MARS detector identifies a P wave arrival at a time of 799 seconds from the PDE origin time. This is marked with an asterisk. The expected arrival time for P from the Herrin tables is 791 to 797 seconds for depths between 40 and 0 kilometers. Therefore, it seems that the



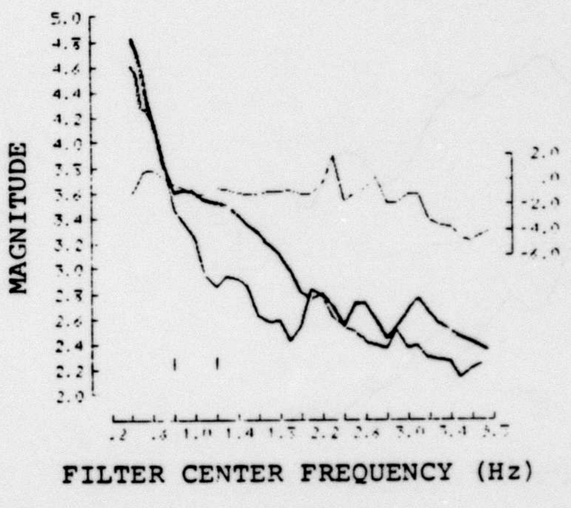


Figure 7. Calculation of $\hat{m}_b$ for a seismogram with peak synthetic/peak noise = 1.

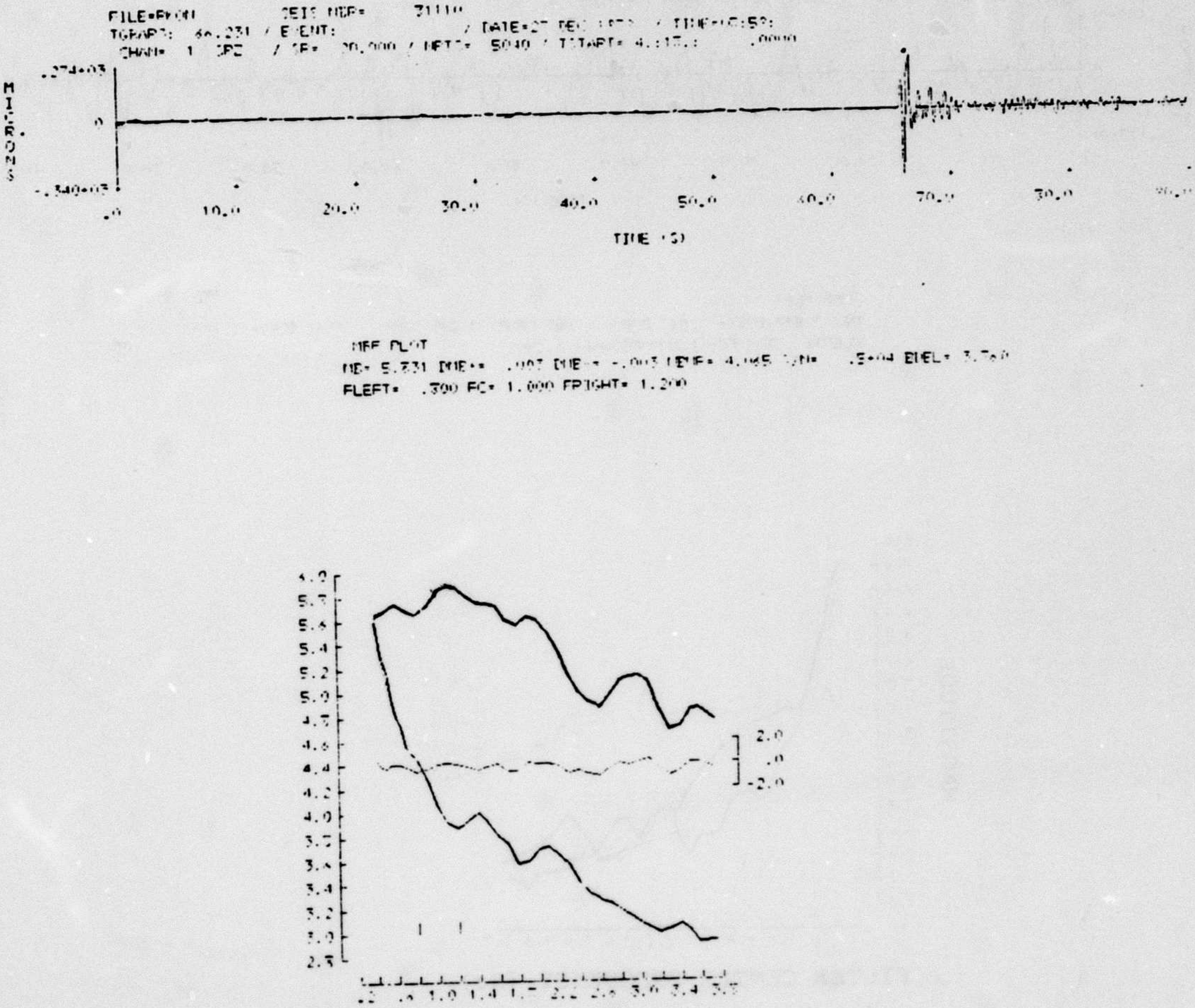


Figure 8. Calculation of $\hat{m}_b$ for the RKON recording of AI Event No. 81.

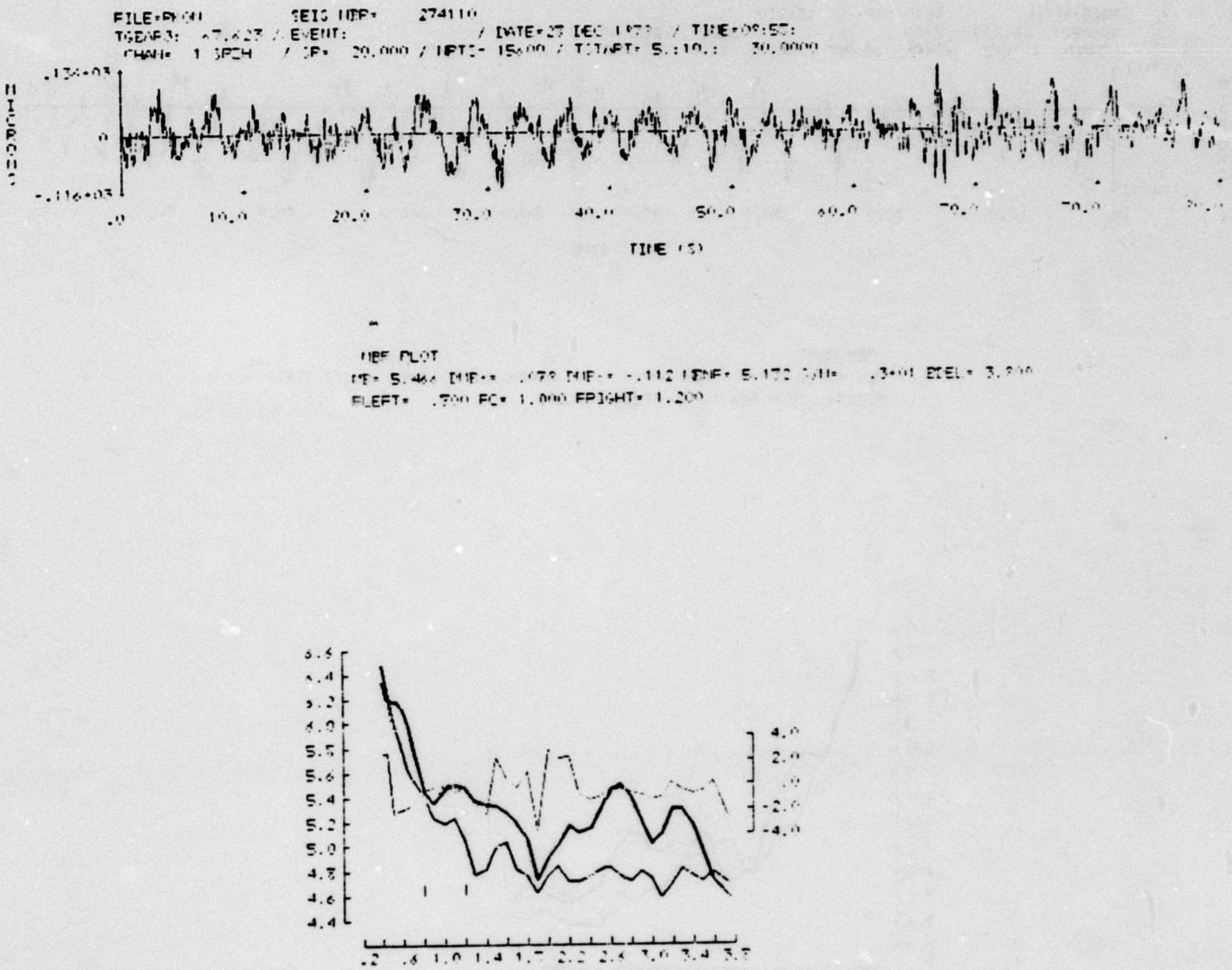
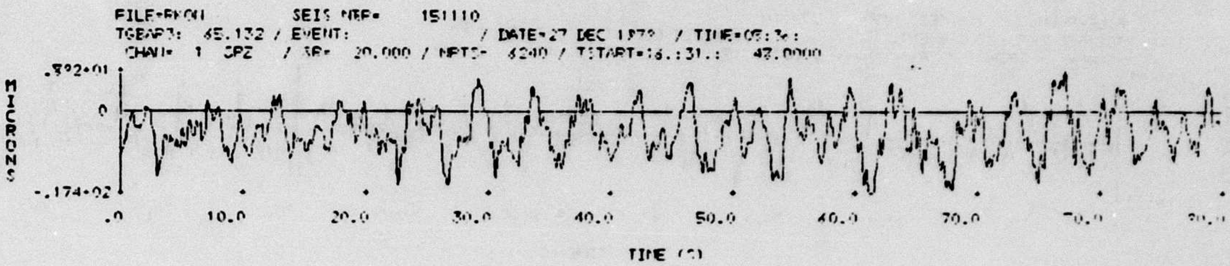


Figure 9. Calculation of $\hat{m}_b$ for the RKON recording of AI Event No. 274.



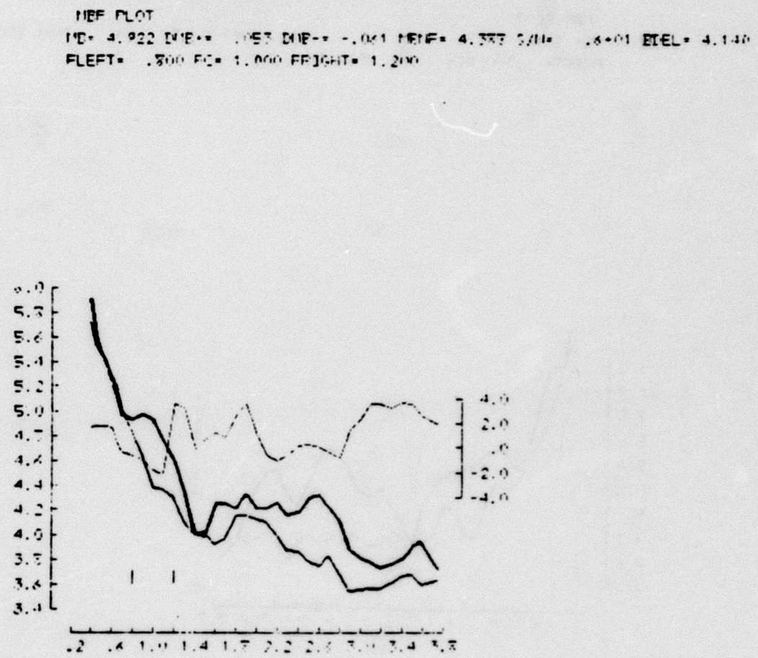


Figure 10. Calculation of $\hat{m}_b$ for the RKON recording of AI Event No. 151.

algorithm correctly detects the P phase and gives an $\hat{m}_b$. We might also mention that the MARS discriminant identified the event as an earthquake.

Conclusions

We began this summary with a rather pedestrian point. If we insist on using time domain m_b for digital recordings, we should correct all seismograms to a common instrument response and let the computer determine the amplitude and period.

Much more important is our demonstration that a capability now exists to automatically compute a spectral m_b. Our demonstration of this may be summarized as follows:

1. In this summary we have shown that the $\hat{m}_b$ is an excellent estimate for the P wave spectral amplitude at 1 Hz. This may be the best indicator for seismic yield we can have.

2. Bache (1979) computed $\hat{m}_b$ for many events and stations. This was done with an earlier version of the algorithm that included no noise corrections. The $\hat{m}_b$ was compared to conventional m_b measured with our semi-automated technique and it was shown that:

   - The station-to-station scatter of $\hat{m}_b$ was no greater than that of m_b.
   - The scatter in $\hat{m}_b$ versus log W is no greater than that for m_b.

We would suppose that the results would be better if noise corrections were included. Finally, we emphasize that the $\hat{m}_b$ is computed by the same MARS process used for detection and discrimination.

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ESTIMATION OF BODY-WAVE MAGNITUDE

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Otto W. Nuttli

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Conventionally body-wave magnitude, m_b, is determined from the amplitude of 1-Hz, vertical-component P waves recorded at teleseismic distances. If the threshold of detection of P-wave ground motion is assumed to be 10 millimicrons, an event with m_b of less than 4.6 cannot be detected at a distance of 30°. If the threshold is 3 millimicrons, the corresponding m_b value is 4.0. Only seismographs with 1-Hz magnification of 100,000 and greater and with noise levels of less than 1 mm of trace amplitude can attain detection levels of 10 millimicrons and less.

To determine m_b values of earthquakes or explosions of $m_b \lesssim 4$ to 4.5, it is necessary to use amplitudes of waves recorded at regional distances. Unless the scaling of seismic spectra is taken into account, the waves used to estimate m_b should be of 1-Hz frequency, the same as of P waves at teleseismic distances. The development of the necessary equations to obtain m_b from the amplitude of phases other than teleseismic P is empirical. That is, one selects earthquakes large enough to be recorded teleseismically but not too large so that amplitudes of regional phases can also be obtained from seismograms. Then the data usually are fitted to an equation of the form

$$m_b = B + C \log \Delta + \log A. \qquad (1)$$

B depends on the excitation of the particular phase and the epicentral distance level, and C depends on the type of phase, its attenuation and the epicentral distance interval. The attenuation in turn is a function of the frequency of the wave.

Evernden (1967) was one of the first to use this approach to derive m_b formulas for regional P phases in the United States. He obtained empirical formulas relating the amplitudes of Pn and Pg to m_b in the western United States. He also obtained a relation of Pn amplitude to m_b in the eastern United States. He noted that the attenuation of Pn in the eastern United States is significantly less than in the West.

Nuttli (1973) related the amplitude of 1-Hz Lg waves to m_b for eastern North America. The Lg phase is as much as an order of magnitude larger than Pn in this region, so that the use of Lg enables m_b to be determined for events as small as m_b = 2 to 2.5. Nuttli (1973) showed that the observed Lg amplitudes satisfied a theoretical curve for the attenuation of dispersed surface waves. The plot of the curve on log-log paper is not linear, but over a limited range of distances it can be approximated by a straight line. Thus, instead of a single formula, such as equation (1), one will have a set of formulas with different coefficients B and C for different distance ranges.

Nuttli (1979b) carried out observational studies on the excitation and attenuation of short-period crustal phases in Iran. He found the most prominent phases to be first P (Pn or mantle refraction), Pg, Sn and Lg. He gave formulas, similar to equation (1), for determining m_b from the amplitudes of Pg and Lg. He also gave a calibration function, $\beta(\Delta)$, to be used to calculate m_b from the amplitude of the first P arrival at distances of 200 to 1300 km, according to the formula

$$m_b = \log A + \beta(\Delta). \qquad (2)$$

The P-wave calibration function for Iran is similar to that found by Nuttli (1979a) for southern Asia, and by Nuttli (1972) for nuclear explosion data. It also is similar to that of Veith and Clawson (1972), but differs significantly from that of Gutenberg and Richter (1956).

Adams (1977) has compiled a list of formulas used for determining magnitudes of near earthquakes in Europe, Asia, Africa, Australasia and the Pacific. From this compilation it is obvious that many seismologists do not attempt to differentiate between m_b, M_S and M_L, and apparently operate under the mistaken assumption that there is only one magnitude value for an earthquake.

When regional formulas are used for m_b determination, the standard deviation for an individual event is usually 0.2 to 0.3 units, if a sufficient number of stations are available. This is similar to the standard deviation of m_b obtained from teleseismic P-wave amplitudes. However, in exceptional cases the amplitude of a crustal phase can vary from the average value by as much as 1.0 m_b unit. Thus, m_b determination based upon the amplitude of a single phase at a single station can be in error by as much as 1.0 units. This can be reduced by using several phases recorded at the single station, which will tend to minimize the effects of focal mechanism variation.

From a limited amount of data (eastern North America, southern Asia and Iran), the theoretical extrapolation of 1-Hz Lg amplitude data back to 10 km epicentral distance for an m_b = 5.0 event gives essentially the same amplitude (within $\pm$0.1 m_b unit). This suggests that the excitation of Lg

is independent of source region. If this proves to be true for the entire world, then all that is necessary to obtain m_b-Lg formulas for a given region is to determine the value of the coefficient of anelastic attenuation (or absorption) for that region. At present we are attempting to do this for the WWSSN stations of Asia.

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NEAR-SOURCE EFFECTS ON P WAVES

8

The matter of the generation of elastic waves by an explosion buried in the earth would appear to be a relatively simple problem. Thus the exercise of characterizing an explosion through the analysis of the radiated elastic waves should be quite tractable and yield definitive results. In practice, this has not proved to be a simple process, and there remain several unanswered questions concerning the relationship between an explosive source and the waves that are emitted from the source region.

Part of the difficulty undoubtedly relates to the fact that observed seismic waves contain the combined effects of the source and the propagation between source and receiver, and it is often difficult to separate these two effects. However, with current knowledge about earth structure and the improving capability for calculating the propagation effects, only a diminishing amount of the difficulty can be attributed to this cause. In what follows, therefore, it will be assumed that propagation effects outside the immediate source region can be calculated and are not a major obstacle in the interpretation. However, it is important to keep in mind this basic fact that the effects of source and propagation are often indistinguishable, and unless one is known the other can not be uniquely determined.

Setting aside the effects of propagation outside the source region, we must consider the possibilities that the waves generated by an explosion are not as simple as we might expect or that the waves are modified by effects very near the source. In most cases it is not very

meaningful to try to separate these two possibilities, and so it is useful to combine them and consider the waves that propagate outward from a general source region, which is taken to be a region surrounding the explosion and including part of the crust in the immediate vicinity.

The subject of this summary will be our current understanding of these waves that propagate outward from the source region of a buried explosion. The discussion will be mainly restricted to the P waves, and thus only body waves will be considered.

Theoretical Considerations. The problem which has the most physical similarity to that of a buried explosion and also has an analytic closed-form solution is that of a pressure pulse applied to the interior of a spherical cavity in a homogeneous elastic medium. Although many differences remain between this idealized mathematical problem and the actual situation of an explosion in the earth, it is reasonable to expect that its solution might provide a first approximation to the more complicated problem.

The solution to the problem of a pressure pulse in a spherical cavity is well-known, and numerous treatements can be found in the literature (for example: Jeffreys, 1931; Sharpe, 1942; Blake, 1952; Favreau, 1969). Because of the spherical symmetry of the problem, the solution can be expressed in terms of a scalar function known as the reduced displacement potential, which is only a function of the reduced travel time. The usual procedure is to specify the pressure history within the cavity, solve for the reduced displacement potential, and then obtain the displacement at any point by taking the gradient of

the reduced displacement potential divided by the distance from the source.

Because of its simple form, the reduced displacement potential has become a popular device for characterizing an explosive source. Since it is independent of distance, it can be determined at any convenient distance from the source. This is helpful in dealing with the complications of an inelatic zone which surrounds large explosions. The initial cavity of a contained explosion will in general be surrounded by successive zones of vaporization, melting, cracking, and inelastic stresses before reaching a zone where the assumptions of linear elasticity are appropriate. In place of the radius of the initial cavity, it is customary to use the inner boundary of the region of elastic behavior as the effective source radius. This is often called the elastic radius. With this modification of the problem the pressure history within the cavity gets replaced by the stress wave which arrives at the elastic radius.

It should be noted in passing that the problem where the source consists of shear stresses applied to the interior of a spherical cavity also has known solutions (Jeffreys, 1931; Honda, 1960). However, so far these results seem to have found little application in the problem of a buried explosion.

A spherically symmetric source, such as a pressure pulse in a spherical cavity, generates only P waves. However, if such a source is placed in a medium with a preexisting shear stress, then S waves will also be generated (Archambeau, 1972). Closely related to this mechanism

is that of triggering a tectonic earthquake on a nearby fault (Andrews, 1973). In either case the secondary source related to the shear stress has the form of a *shear dislocation*, and so its radiation pattern is a quadrapole and it is roughly an order of magnitude more efficient in generating S waves than P waves. Thus, observational studies of this effect have been based primarily on S waves and surface waves (Press and Archambeau, 1962; Aki et al., 1969; Archambeau and Sammis, 1970; Toksoz et al., 1971; Lambert et al., 1972; Aki and Tsai, 1972; Toksoz and Kehrer, 1972). It appears from the results of these studies that, while reasonable levels of prestress can be important in the generation of S waves and surface waves, the direct P waves generated by this secondary source will be small compared to those generated by the explosive source. However, the mechanism whereby S waves from the secondary source are converted to P waves at nearby boundaries could be an important source of P waves.

The next step in constructing a more realistic model for a buried explosion in the earth is to consider the effects of material inhomogeneity in the vicinity of the explosion. Such inhomogeneity causes the reflection and refraction of primary waves from the source, the conversion of wave type from P to S and vice versa, and the generation of interface waves. For shallow explosions, a very important inhomogeneity is the free surface of the earth. Geologic layering, the water table, and fault surfaces are other types of inhomogeneities which can also have significant effects.

Attempts to obtain exact solutions to the problem of an explosion in the vicinity of an inhomogeneity have not been very successful.

The problem of a finite spherical source embedded in a homogeneous elastic halfspace has been considered by Ben-Menahem and Cisternas (1963) and by Thiruvenkatachar and Viswanathan (1965, 1967). This is a problem with mixed boundaries and its solution is very difficult, the answer usually being expressed as an infinite series. Because of the rather untractable form of the results, the analytical treatment of this problem has not yet contributed any practical results to the problem of a buried explosion.

Provided one is willing to approximate an explosion with a point source, the effect of vertical inhomogeneity in the vicinity of the source can be handled in a satisfactory manner. Two approaches are commmonly used, the method of generalized rays (Helmberger and Harkrider, 1972) or the method using propagator matrices (Fuchs, 1966; Hasegawa, 1971).

The treatment of lateral inhomogeneity near an explosive source is a more difficult problem. This would include such features as dipping layers, faults, and surface topography. In general, numerical methods are required to calculate the effects of this type of inhomogeneity, but usually the detailed knowledge about the geometry of such features is not sufficient to justify an elaborate computational treatment.

The presence of inhomogeneity near the explosive source can also lead to additional inelastic effects. Spallation is an important effect of this type. When the P wave from an explosion is reflected at the free surface with a change in sign, the associated stress can cause failure in tension of the near surface material. When failure occurs,

part of the near-surface material may separate and move ballistically upward, eventually falling back and impacting the earth at some later time. This closure of the spall is sometimes referred to as slapdown. This process of spallation has been fully described and documented in the literature (Eisler and Chilton, 1964; Eisler et al., 1966; Chilton et al., 1966; Perret, 1972; Viecelli, 1973; Springer, 1974). The energy in the P wave which leaves the source in an upward direction is converted into a reflected pP wave, a reflected sP wave, a surface wave, inelastic effects, and the waves generated by the slapdown. Viewed from a distance, the primary effects of spallation upon the P-wave coda is a diminished amplitude of the pP wave and an additional phase at the time of slapdown.

On the basis of the preceding discussion it is possible to construct a general model of a buried explosion. The model is primarily an elastic model so it begins with a stress pulse applied at the elastic radius. This generates an outward propagating P wave and, if the source region is prestressed, also an S wave. These outward propagating waves interact with inhomogeneities in material properties near the source to produce reflected and interface waves. Waves reflected from above the source may also cause spallation and the associated slapdown. The outward propagating waves may also trigger secondary shear dislocations at near-by stress concentrations. All of these effects combine and interact to produce the waves that propagate out from the general source region.

Experimental Results. Consider now the question of what can be learned about a buried explosion from an analysis of experimental data. This is a basic inverse problem. The effects, primarily the waveforms of elastic waves, are to be used to estimate properties of the cause, the explosive source. The usual approach is to construct a general model of the source containing a number of undetermined parameters. The observational data are then analyzed to deterimine whether the model is capable of explaining the data and, if so, what values should be given to the parameters.

Because of the advantages already mentioned, it has become common to characterize an explosive source by its reduced displacement potential, and so the experimental determination of the reduced displacement potential has received considerable attention. The generalized form of the reduced displacement potential as a function of time is an abrupt start, a smooth rise to a maximum over a finite time, and then a decrease to a static level. Three numbers - the rise time, the static value, and the ratio of maximum to static values - describe the major features of such a function. In the frequency domain, the first time derivative of the reduced displacement potential has a correspondingly simple form. Going from high to low frequencies, the spectrum rises at some slope, reaches a maximum near what is called the corner frequency, and then decreases to a constant value at low frequencies. In terms of the time domain parameters, the spectral amplitude scales with the static value, the frequency scales with the rise time, the ratio of the spectral maximum to the low frequency level depends upon the ratio of maximum to static values of the reduced displacement

potential, and the high frequency slope depends upon the abruptness of the beginning of the reduced displacement potential.

The simple model of a pressure pulse applied to the interior of a spherical cavity can be used to relate the parameters of the reduced displacement potential to the physical properties of the explosive source. The static value depends upon the static pressure, the cavity volume, and the material properties. The rise time depends upon the cavity radius and the material properties. The ratio of maximum to static values, sometimes called the overshoot, depends upon the time history of the pressure pulse and the material properties. Considering the pressure pulse to consist of two main parts, an impulse and a step, the overshoot will increase as the ratio of impulse to step increases.

For a more complete parameterization of the reduced displacement potential, it can be approximated with an analytic function. Haskell (1967) argued that displacement, velocity, and acceleration should all be continuous at the elastic radius and used a fourth order polynomial in time. Von Seggern and Blandford (1972) required that only the displacement be continuous and used a second order polynomial. Mueller and Murphy (1971) used the theoretical solution for a pressure pulse within a cavity and a semi-empirical expression for the shape of the pressure pulse to arrive at an analytic expression for the reduced displacement potential. The basic difference in these three models is primarily at the high frequencies, where the Haskell model falls off with a -4 slope on a log-log scale, while both the von Seggern-Blandford and Mueller-Murphy models fall off with a -2 slope. The experimental data (von Seggern and Blandford,

1972; Murphy, 1977; Burdick and Helmberger, 1979) at teleseismic, regional, and near distances mostly favor the -2 slope of the von Seggern-Blandford and Mueller-Murphy models. From an analysis of very-near data, Peppin (1976) found evidence for a slope of at least -3 at the high frequencies.

The problem of estimating the reduced displacement potential for an explosion can be approached from at least three directions. The dynamic equations for the explosion and the surrounding inelastic region can be solved numerically and the calculations carried out to the elastic radius (Holzer, 1966; Rodean, 1971). Another approach is to measure the ground motion as near the source as possible while still in the elastic region, and then calculate the reduced displacement potential on the basis of these measurements (Werth et al., 1962; Werth and Herbst, 1963). In most cases the reduced displacement potential can only be determined for times less than about 0.5 sec because the effect of the free surface and other departures from spherical symmetry begin to affect the results at later times. A third approach is to record elastic waves at near to teleseismic distances and then attempt to infer the reduced displacement potential which best explains the observations. What emerges from this approach is an apparent reduced displacement potential, because, as already discussed, the waves that emerge from the general source region can consist of considerably more than the direct P wave from the explosion. Such additional effects must be taken into account in the interpretation of the data.

Because of the abundance of easily accessible data, numerous studies have used waveform data recorded at near to teleseismic distances to

investigate the details of explosive sources. Various methods of interpretation have been employed. A few of the representative studies will be summarized below.

Molnar (1971), Kulhanek (1971), and Wyss et al. (1971) all studied the spectra of teleseismic P waves and found that the spectra were modulated. This can be explained, at least partly, by the interference of the P and pP waves. Filson and Frasier (1972) and King et al. (1972) fit theoretical models to the spectra of teleseismic P waves to estimate parameters of the reduced displacement potential.

Aki et al. (1974) studied both near and teleseismic data and found evidence for a large overshoot ratio in the reduced displacement potential. Peppin (1976) analyzed near and regional data and did not find evidence for overshoot. Murphy (1977) studied a variety of data from near to teleseismic distances and concluded that the data were consistent with the source model of Mueller and Murphy (1971).

Frasier (1972) deconvolved teleseismic P waves and found evidence of a pP phase plus a later phase, possibly due to slapdown. Bakun and Johnson (1973) applied homomorphic deconvolution to teleseismic P waves and found indications of the pP reflection and also a later phase which was consistent with an interpretation in terms of slapdown. Burdick and Helmberger (1979) used synthetic seismograms to model teleseismic P waves and found that the data could be explained by substantial overshoot in the reduced displacement potential and reflected crustal phases, but did not require a slapdown phase.

Stump and Johnson (1977) and Stump (1979) have developed a general inverse method for estimating the second-rank seismic moment tensor. The trace of this tensor is equivalent to the reduced displacement potential and the deviatoric components provide a means of expressing other effects, such as tectonic stress release.

There is obviously not total agreement among the observational studies concerning the source properties of explosions. This is partly due to the different methods of interpretation which have been employed. It is also due to a certain degree of nonuniqueness that exists in the basic problem. This is compounded by the fact that propagation effects must usually be taken into account in the interpretation of the data, and any uncertainty in earth structure can get translated into nonuniqueness in the source properties. For instance, a peaked spectrum of the P wave coda can be produced in at least three ways: overshoot of the reduced displacement potential, interference caused by reflected waves, or over-correction for the effects of attenuation in the interpretation process.

<u>Conclusions.</u> The available theoretical models of an explosive source embedded in a realistic crustal structure appear to be sufficiently general to explain the major features of the observational data. It is clear that considerably more than the direct P wave emerges from the general source region. Reflections from inhomogeneities near the explosion, particularly the free surface, are definitely important. The importance of spallation and slapdown upon the generation of P waves is still rather uncertain. Tectonic strain release may be important in some instances, and, when it is significant, it is probably not the direct P waves from the secondary sources but rather the S to P converted phases which contribute most to the P wave coda. The entire matter of how S waves are generated in the vicinity of an explosion is still not completely understood, and some new process, such as the acoustic fluidization suggested by Melosh (1979), may eventually provide the answer.

The methods of interpreting the observational data to infer source properties are steadily improving, and the increased use of synthetic seismograms and more complete inversion schemes should be very helpful. This progress in interpretation will require a more accurate knowledge of earth sturcture, because it is doubtful that the accuracy of inferred source properties can ever be greater than that of the earth model used in the interpretation. The anelastic properties of the earth present a current problem in this respect, because the correction for attenuation which must usually be applied in the interpretation process is still rather uncertain and this can have a major effect at the high frequencies.

Hopefully, continued improvements in the methods of interpretation and knowledge of earth structure will reduce the inherent nonuniqueness in this inverse problem.

The reduced displacement potential has become a popular means of summarizing the properties of an explosive source and has been quite useful. But is should be emphasized that in most cases this is only an apparent reduced displacement potential. The requirement for spherical symmetry holds, if at all, only for a few tenths of a second, and asymmetry in the source region is an important factor in most P wave codas. The importance of this point is the realization that the apparent reduced displacement potential for a given event may be different as viewed from different distances, different azimuths, or different types of data.

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LAWRENCE LIVERMORE LABORATORY

Donald Springer

February 11, 1980

Col. George Bulin, USAF
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Dear George:

This is my tardy response to the request for a "state-of-the-art" assessment as outlined in Dr. Carl Romney's letter dated December 18, 1979.

I am tardy because of my workload here at the Laboratory (and then the "earthquakes"), but also because I have taken extra time to search through some of the older literature.

<u>Experimental Data on Body Wave Coupling</u>

It has been shown many times that the efficiency with which underground nuclear explosions generate and radiate seismic body waves is related to several factors: among them being explosion medium, yield, and depth of burial (references 1-10). Most of these studies were hampered by the difficulty in separating these source factors from other near-source factors and propagation factors. Thus, correlations of seismic amplitudes with coupling factors are not very precise-ranging under controlled conditions (similar medium, etc.) from 15-20% to 50-100% for amplitude vs yield variations have generally been explained as being caused by unknown source factors as well as near-source factors (other than coupling factors). I will discuss some of these near-source factors later, but I wish to dwell on source factors first.

Given all other factors near equal, seismic body-wave amplitude (that is, $\log_{10}A$) vs yield relationships have been shown to have slopes of about 0.9 (references 11 and 12). Values for the slope can vary due to whether $\log_{10}A$ or $\log_{10}A/T$, where T is dominant period, is used. Lower slopes are obtained when amplitudes of head waves (rather than body waves) are used (references 12 and 13). The reasons for this are not well understood, but probably involve the explosion source function having some characteristics of a peaked spectrum (a pulse in the time domain) and the attenuation characteristics of the propagation path. I'd like to point out that the scaling relations mentioned above generally only apply to waves of about 1 second period. If the explosion source function is assumed to have a peaked spectrum, then the slopes of amplitude vs yield relations will be frequency dependent. Some studies have suggested such frequency dependence (references 14-25), but no really comprehensive empirical study has been reported.

Many other theoretical studies (references 26-37) are available on the subject of amplitude-yield scaling. I believe the medium dependence of the shape of the source function is not well understood, although perhaps well enough for first-order predictions of teleseismic body waves at about 1 second period. The high-frequency (3 Hz) and low frequency (0.5 Hz) characteristics of the source deserve further study. It should be clear from explosion-cavity studies (references 38-41), that the longer-period amplitudes in the

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P-wave source depend on source medium in a different way than the shorter-period amplitudes in the P-wave source (cavity radii have a rather weak medium dependence).

Studies of close-in data (references 42-51) have shed some light on medium dependence (and other factors), but a more comprehensive study of all near-field data should be considered for the future.

The effect of depth-of-burial on generated seismic body waves has been treated by some of the above studies and other (references 52-55). I believe the effect on short-period P waves is not great (although I wouldn't argue that such an effect does not exist); however, the effect on long-period P waves (and perhaps Rayleigh waves) could be more significant.

Related to depth of burial is one of the near-source factors affecting body wave generation; namely, the free surface. A number of studies (references 56-60) suggest that both the surface reflected P wave (pP) and the spall-closure wave (P_s) will influence the nature and amplitude of teleseismic P. These factors deserve more study in the future, because the dependence on depth, medium, and geologic structure is not understood. The dependence of body-wave amplitudes on underlying structures has been noted in reference 11, but this too needs further study. The comprehensive study of near-field data I suggested earlier could help in these areas also. Surface spall has been noted before by other workers (references 61 and 62) as well as those mentioned earlier.

Next, the geometry of the explosion could be a significant factor (references 63-66) although their appears to be little effect in some cases (reference 67).

The way in which seismic amplitudes from multiple explosions superpose (and scale) have been reported (references 68-70) and is worth noting here. In addition, the effects of linear arrays of charges have been studied by others (reference 71), although some investigations concentrate on the effect on surface-wave generation.

Tectonic release caused by an underground explosion may also influence body-wave and surface-wave seismic radiation (references 72-76), but this phenomenon is not well understood and deserves further study.

Differences that may exist in seismic radiation from chemical explosions compared to nuclear explosions should be of some current interest. Thus, some studies of chemical explosions deserve mentioning (references 77-80).

Many other workers have reported on various aspects and factors mentioned above. A comprehensive list of references may be impossible to put together but I have listed most of those I know of. While I don't necessarily have the same points of view as reported by some of these workers, I believe their work should be mentioned for a balanced evaluation. It should be noted that most of the Soviet articles concern effects of chemical explosions.

Regional Attenution Effects on P Waves

It is universally accepted that attenuation effects on P waves varies from one region of the world to another (references 1-10). Unfortunately, in most cases isolating attenuation effects from scattering effects is a near-impossible task; thus, attenuation determinations (of say, quality factor, Q) typically include the effects of scattering, dispersion, and frictional losses (absorption) together. These individual effects have been studied in the laboratory (references 11-15), but much remains to be understood. I believe the effects on P wave magnitudes could amount to as much as 0.4-0.5 units (relative to the most efficient transmission), although generally less (0.1-0.2 units). It will be a long and difficult task to regionalize (in a very precise way) the earth as far as attenuation characteristics go, but many are working in this arena (references 16-50). I include many of the surface-wave studies because those results are relevant to P waves also. Other reference to laboratory and seismological studies are included for consideration (references 51-66). An overall reference of value is a compilation of Soviet works (reference 67).

Site Specific Propagation Effects

As mentioned in the first section, near-source effects may be significant (references 1-2) causing focusing or defocusing of seismic waves. This can be at least 0.3 magnitude units according to reference 2, although these and other results should be assessed carefully. It is not known to what extent other factors, such as variation in coupling medium, may also be helping to give the variations that are attributed solely to the underlying, near-source geology. I should point out that some of the other studies mentioned in the first section could very well be discussed under the heading of this section.

Reference 3 discusses the insignificant effect that the shape of the Tatum Salt Dome had on the radiated P-waves for Salmon. This was a theoretical study, however, that experimental data has never confirmed (nor refuted).

Reference 4 discusses some local effects and may be pertinent to this topic.

I hope the discussions above will be of some value to your considerations. They certainly are not comprehensive.

Sincerely,

Donald Springer

DS:dt

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Teledyne Geotech
Robert R. Blandford
Experimental Data on Body Wave Coupling

Blandford (1976) found for the event series Buteo, Rex, Scotch, and Benham that regional amplitude and spectra fit the predictions of cube-root-scaling for the Werth and Herst tuff potential; and teleseismically fit the predictions of cube-root-scaling for the granite potential. This implies that the "reduced displacement potential" is a function of take-off angle. There is no evidence in the data to require a dependence of reduced displacement potential on depth, although there is no doubt in my mind that such dependence must exist for great enough ranges of depth; in particular for very shallow depths.

Blandford (1978) found that there was substantial evidence for variation of reduced displacement potential with depth for explosions in salt; thus data from Salmon, Gnome, and several Soviet explosions in salt could not be matched with cube-root-scaling of the Salmon reduced displacement potential. The theory of Mueller and Murphy did a better job of matching the amplitude at 1 Hz, however it also failed to match the spectral trends at high frequencies where the observations showed too much high frequency for either theory. The discrepancy between Salmon and Gnome may be explained by the fact that Gnome is in layered salt and has a high dirt content; thus it's reduced displacement potential may be different due to a difference in medium as compared to the Salmon dome salt.

Phase Pg and Lg do not show as great a variation with medium as do the phases Pn and teleseismic P, suggesting that the shot point medium influences the propagation of Pn and P in such a way as to enhance the variation in coupling. This is as would be expected due to ray curvature resulting in a small portion of the focal sphere going to teleseismic distances for slower shot media. Blandford and Klouda (1980)

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Site Specific Propagation Effects

A full discussion of the analysis by Chang and von Seggern (1977) showing variations of amplitudes at LASA due to mantle structure is given under the topic "Estimation of Body Wave Magnitude". There we see that for different azimuths of approach the relative magnitudes at stations only 50 km apart can vary by 0.4 m_b units in standard deviation. This is due to structure in the upper mantle, and is difficult to predict.

Der et. al. (1979) have shown the large average effects which can result from differences in velocity of the surface crustal layers. For stations not on hard rock these corrections should be made, and the art of so doing is well founded.

We all know from a casual reading of the literature that large variations in teleseismic absorption can occur at such locations as Yellowstone and the Geysers.

Chang, A., and D. W. von Seggern, (1977). A study of amplitude anomaly and m_b bias at LASA subarrays, TR-77-11, Teledyne Geotech, Alexandria, Virginia.

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Near Source Effects on P Waves

A full discussion of analyses by Chang and von Seggern (1977) showing variations of amplitudes at LASA due to mantle structure is given under the topic Estimation of Body Wave Magnitude. By reciprocity these results imply similar effects for outgoing waves.

Chang, A., and D. W. von Seggern, (1977). A study of amplitude anomaly and m_b bias at LASA subarrays, TR-77-11, Teledyne Geotech, Alexandria, Virginia, 22314.

SYSTEMS, SCIENCE AND SOFTWARE

11

EXPERIMENTAL DATA ON BODY WAVE COUPLING

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Body wave data are currently available from U.S. underground explosions in the following media: alluvium (wet and dry), tuff/rhyolite (wet and dry), granite, salt, shale and limestone/dolomite. These data fall into two subsets; body wave magnitude data and free-field measurements of the seismic source function. In this review, I will summarize what I think is known from these two data sources and assess to what extent they present a coherent picture of body wave coupling.

With regard to the body wave magnitude data, the sample compiled by Alewine and Young at VSC (Personal Communication, 1977) has been selected for analysis because these magnitudes were carefully determined in a consistent fashion for all the events in the sample. These data seem to support the following conclusions:

(1) The subset of the sample consisting of explosions in NTS wet tuff/rhyolite (T/R) emplacement media is the only one which is complete enough to permit the definition of a

statistically significant m_b/yield curve over the yield range of potential interest (i.e. from about 1 to 1000 kt).

(2) A linear regression of m_b on log W (i.e. assuming all the error in m_b) for the NTS wet T/R subset of the sample gives:

$$m_b = 3.92 + 0.81 \log W \qquad (1)$$

The standard error of estimate associated with this fit is 0.12 magnitude units (i.e. approximately 68% of the data lies in the region $m_b \pm 0.12$ with m_b given by equation (1)).

(3) The m_b data for U.S. explosions in granite (Hard Hat, Shoal, Pile Driver) cannot be distinguished from the predictions of equation (1) in any statistically meaningful sense.

(4) The m_b data for NTS explosions in dry, unconsolidated material (alluvium, tuff) fall low of equation (1) by about 0.5 ± 0.25 magnitude units.

(5) The m_b data for NTS explosions in dolomite and limestone are widely scattered but generally fall low with respect to equation (1).

(6) The only available m_b value for salt (i.e. Salmon) agrees almost exactly with the value predicted for that yield by equation (1).

(7) The m_b value for Gasbuggy, which was detonated in shale, agrees very well with equation (1). However, the m_b values for the Rulison and Rio Blanco events, which were also detonated in shale, fall low of equation (1) by about 0.3 magnitude units.

Now, to some extent, the above observations are relevant to the assessment of the relative body wave coupling in the various source media. However, as is indicated by the

Gasbuggy/Rulison comparison, other factors can also affect the observed teleseismic m_b value. Another example of this is provided by the Faultless explosion which was detonated in wet tuff only a short distance north of NTS. Given the medium and the proximity to NTS it might be expected that the Faultless m_b value agreed with the prediction of equation (1). In fact, the observed m_b value was higher than that predicted by equation (1) by a statistically significant amount. Thus, regional differences in propagation paths (and possibly other factors) can significantly affect m_b and, since the m_b data used to infer the above-listed conclusions represents a wide variety of geographic locations and geologic environments, it is not clear to what extent they correlate with differences in body wave coupling at the source. In order to address this question, I have examined the available free-field data from explosions in the various media using the recent compilations of Murphy (1978) and Murphy and Bennett (1979). In the following discussion the seismic coupling as a function of source media will be assessed by comparing the observed reduced displacement potentials (RDP's). As a basis of comparison, I have selected the RDP that would be predicted for the given yield and depth of burial using the Mueller/Murphy scaling relations (Mueller and Murphy, 1971). Figure 1 shows that this prediction provides a good fit to the observed RDP data from the Discus Thrower and Rainier events in tuff and thus should provide a reasonable basis for comparison.

Figure 2 shows a comparison of the observed RDP's from four events in dry alluvium with the corresponding predicted RDP's for events of the same yields and depths of burial in wet T/R emplacement media. It can be seen that these data clearly indicate that dry alluvium is a weak coupling medium with respect to wet T/R, in agreement with the differences noted in the teleseismic m_b data. Figure 3 shows a similar

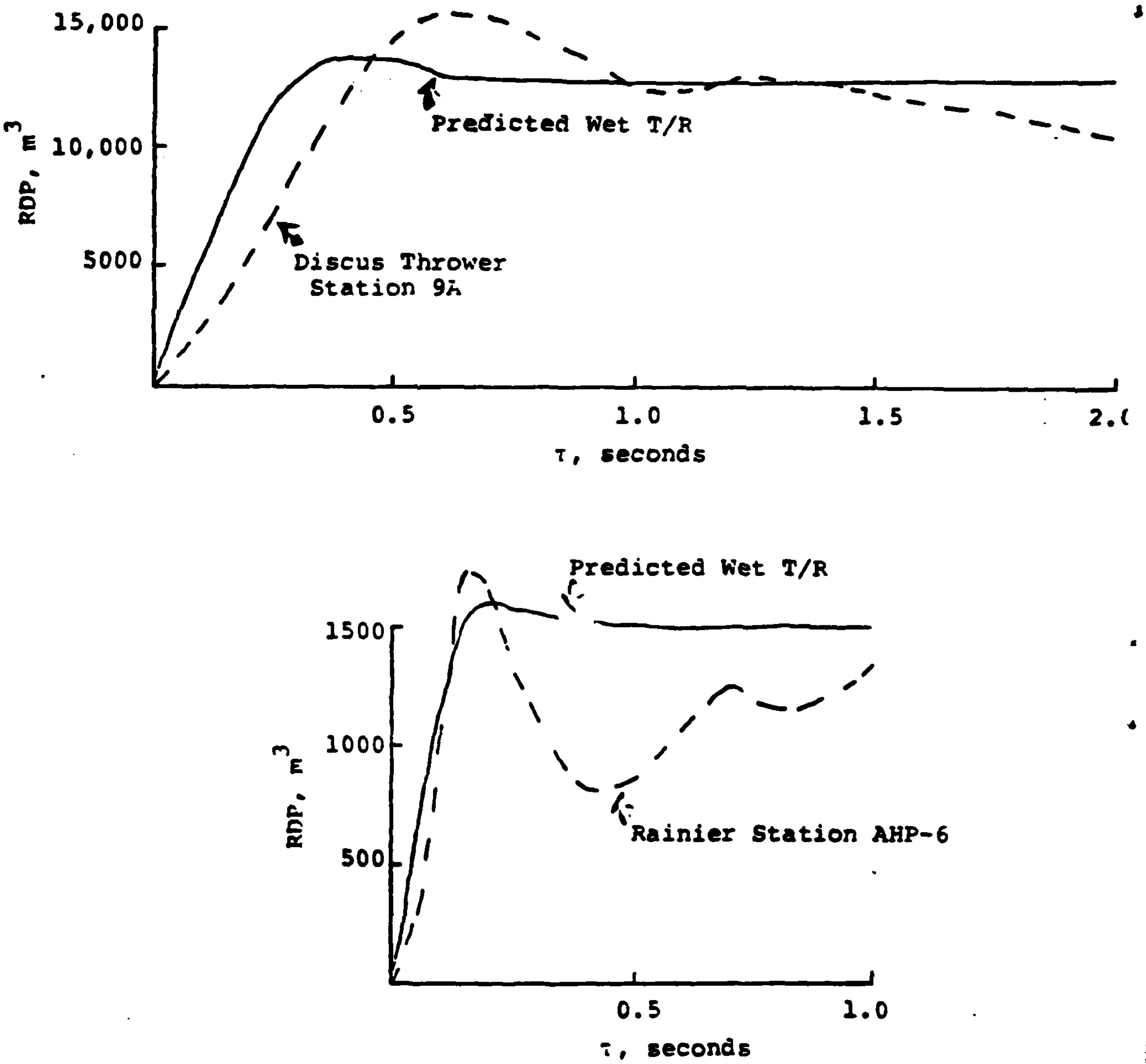


Figure 1. Comparison of Observed Tuff RDP's With RDP's Predicted For the Same Yield and Depth of Burial in a Wet Tuff/Rhyolite Emplacement Medium.

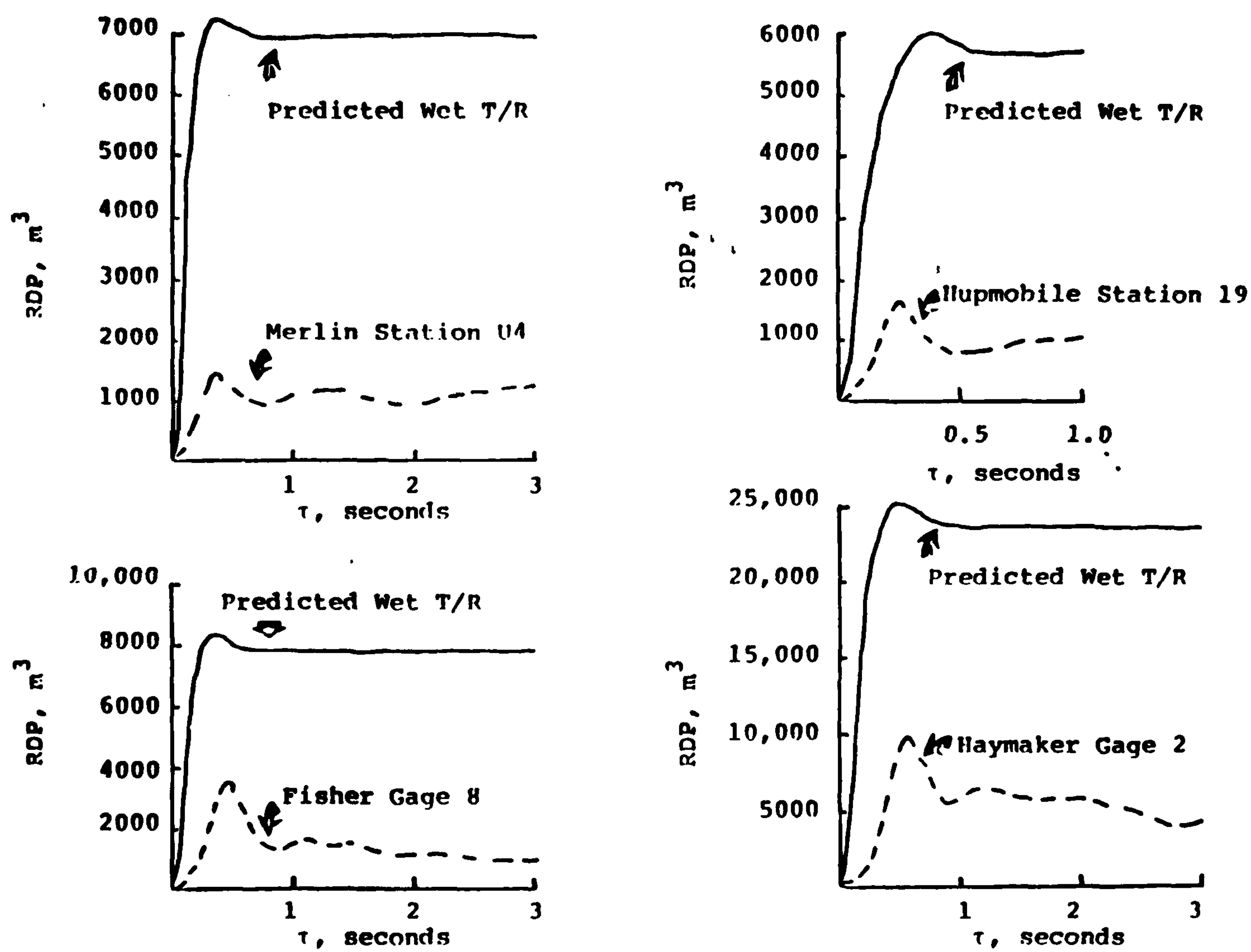


Figure 2. Comparison of Observed Alluvium RDP's With RDP's Predicted For the Same Yield and Depth of Burial in a Wet Tuff/Rhyolite Emplacement Medium.

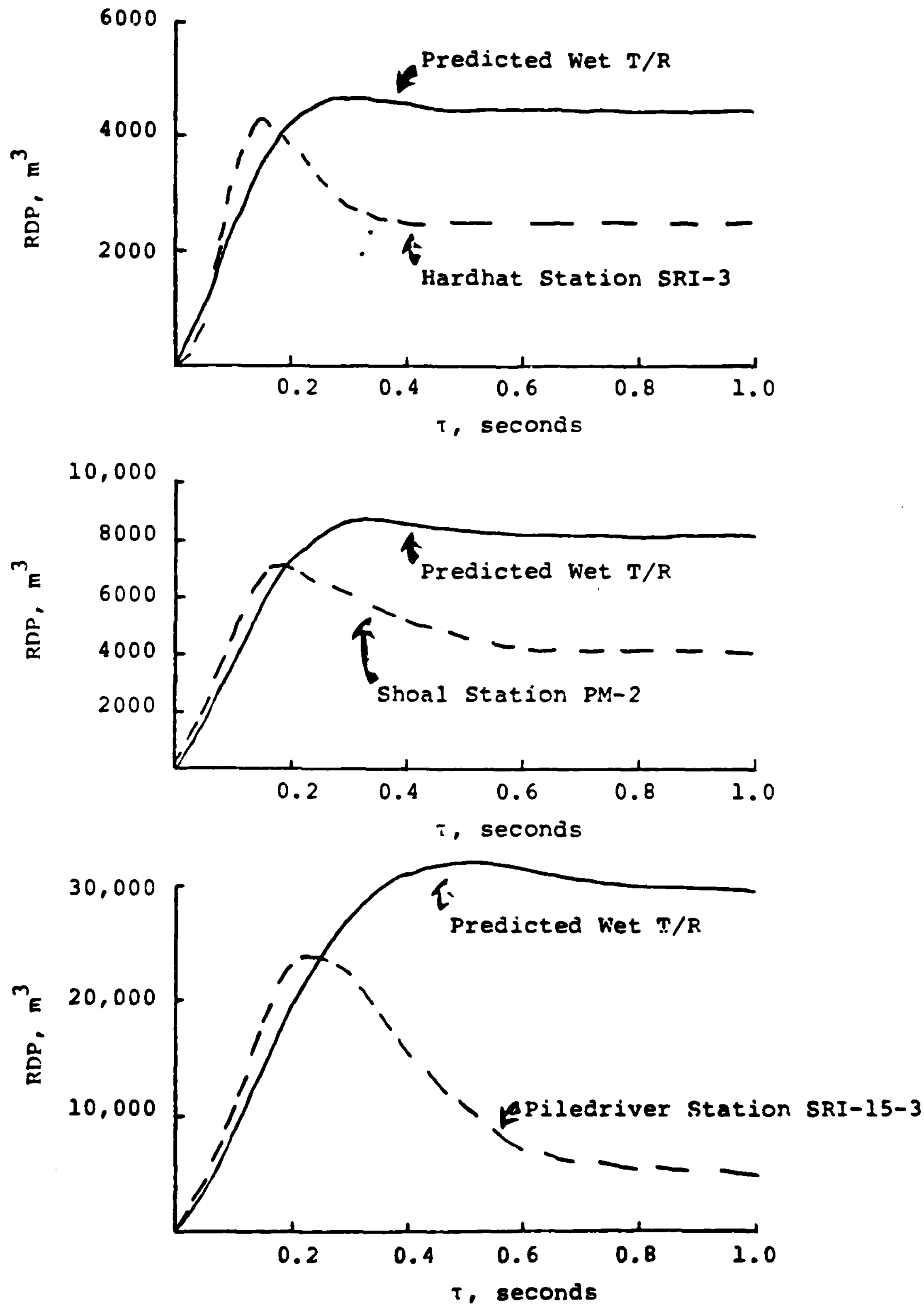
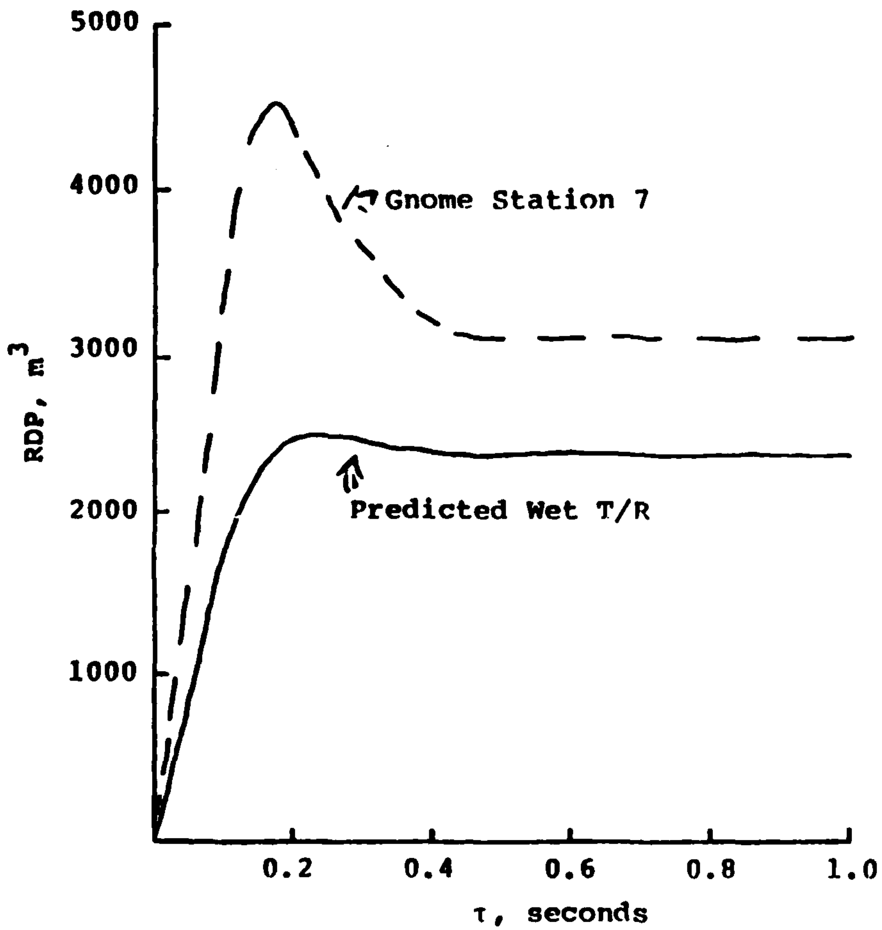


Figure 3. Comparison of Observed Granite RDP's With RDP's Predicted For the Same Yield and Depth of Burial in a Wet Tuff/Rhyolite Emplacement Medium.

comparison for events in granite. It is well known that it is difficult to specify "the" observed RDP for events in granite because the observed data from any one event show wide scatter, presumably due to the effects of block motion etc. In any case, the observed RDP data presented here do suggest that granite is a somewhat less efficient coupling medium than wet T/R. The teleseismic m_b data, on the other hand, indicate essentially identical coupling for these two media.

Figure 4 shows the comparison for explosions in salt. It can be seen that these RDP data indicate that salt couples better than wet T/R. On the other hand, the only available teleseismic m_b data point for salt (i.e. Salmon) suggests that the seismic coupling in the two media is about the same. Finally, Figure 5 shows the RDP comparisons for shale (Gasbuggy) and dolomite (Handcar). It can be seen that these data suggest that the seismic coupling in both media is low with respect to wet T/R. This agrees with the observed teleseismic m_b values for Handcar and the Rulison and Rio Blanco events in shale. However, it is not consistent with the observed Gasbuggy m_b value which was found to be in good agreement with that predicted by equation (1).

Thus, the observed m_b values and RDP data are in qualitative agreement in most cases, although there are some notable discrepancies. One source of these discrepancies is related to the fact that the coupling into the teleseismic transmission path can be expected to vary with the source medium. That is, if different source media are taken to overlay the same upper crustal structure, the transmission of energy out of the source layer will depend on the impedance mismatch at the layer boundary and thus on the physical properties of the source layer. Bache et al. (1975) have shown that this effect can be approximately accounted for by multiplying the source region RDP by



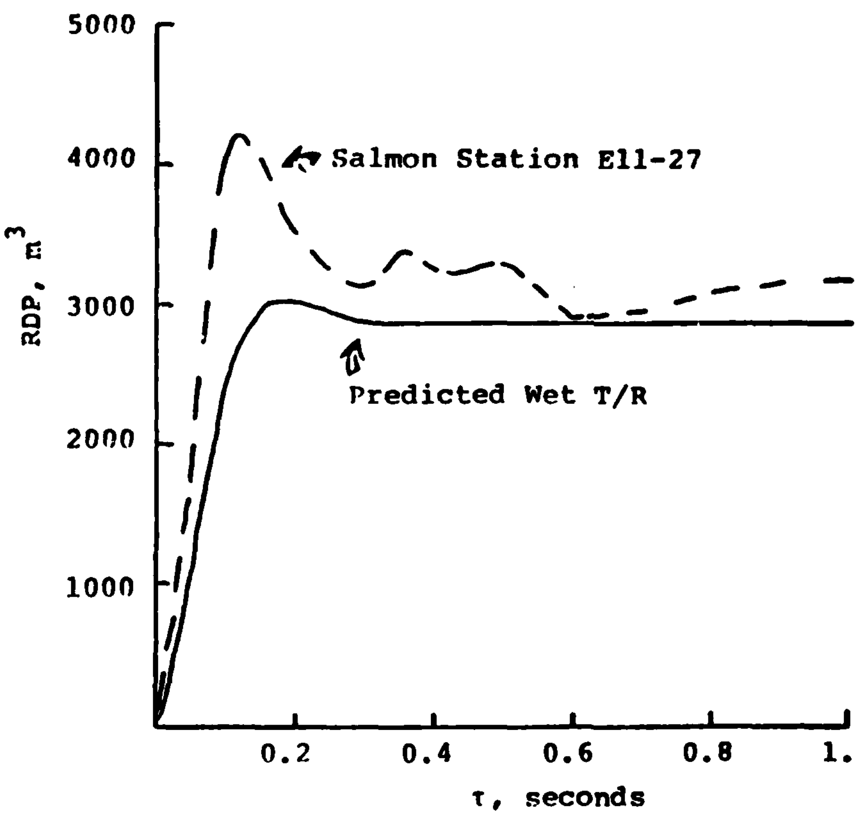


Figure 4. Comparison of Observed Salt RDP's With RDP's Predicted For the Same Yield and Depth of Burial in a Wet Tuff/Rhyolite Emplacement Medium.

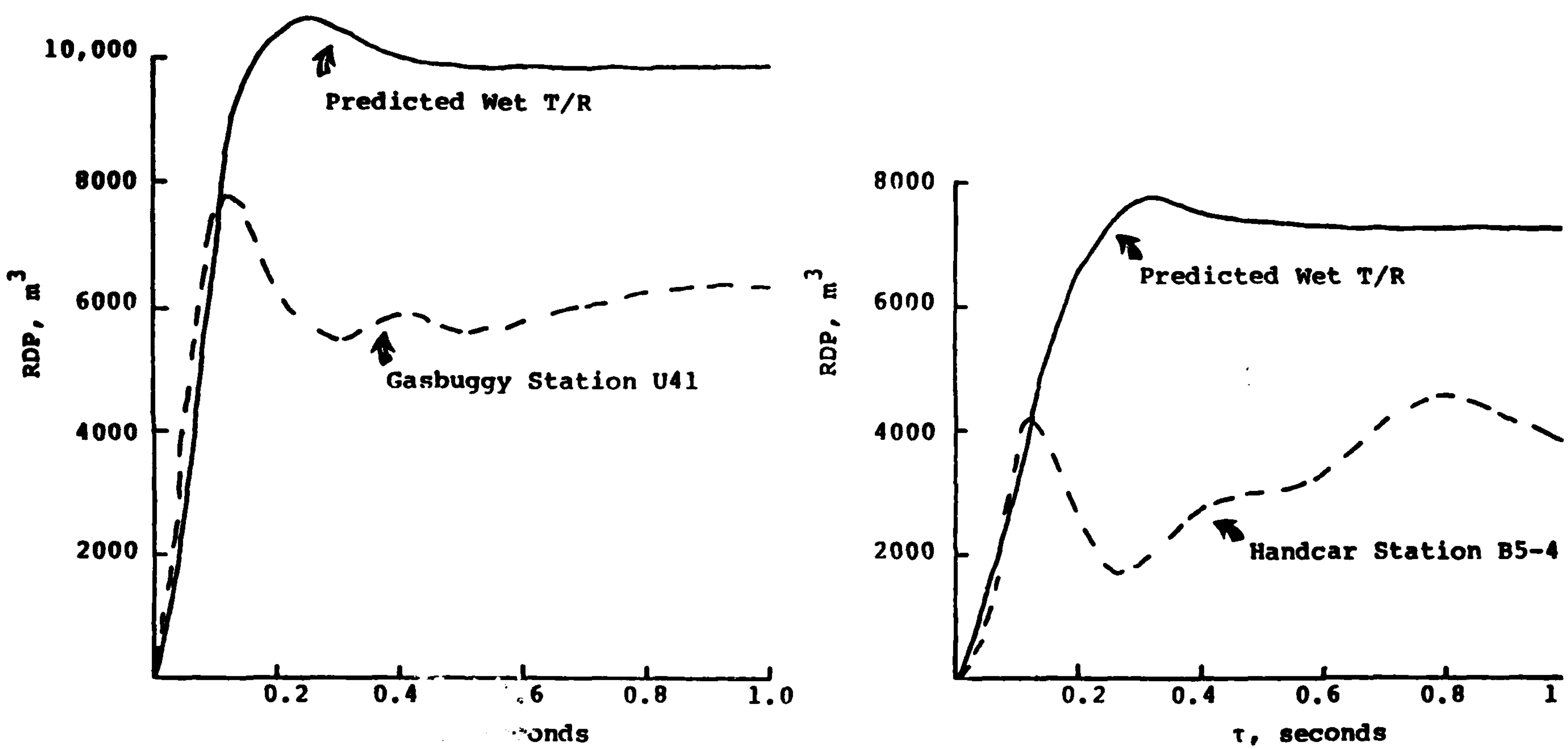


Figure 5. Comparison of Observed Shale (Gasbuggy) and Dolomite (Handcar) RDP's With RDP's Predicted For the Same Yield and Depth of Burial in a Wet Tuff/Rhyolite Emplacement Medium.

the compression wave velocity in the source medium.

When the RDP's of Figures 1-5 are corrected in this manner for the effects of coupling into the teleseismic transmission path, it is found that the free-field data suggest: (i) that there will not be much difference between the m_b/yield curves for events in wet T/R, granite, shale and dolomite emplacement media, (ii) that the m_b/yield curve for explosions in dry alluvium is expected to lie well below that for wet T/R and (iii) that the m_b/yield curve for explosions in salt is expected to lie significantly above that for wet T/R.

Thus, both the m_b and free-field data indicate that the body wave coupling for explosions in wet tuff/rhyolite, granite, shale and dolomite is about the same and that the body wave coupling for events in dry alluvium is significantly lower. However, although the salt RDP's clearly suggest higher coupling, the only available salt m_b value (Salmon) is very comparable to that expected for that yield in wet T/R. However, Salmon was deeply overburied and I feel that consequently its m_b value is lower that it would have been if the explosion had been detonated at the normal containment depth of $122W^{1/3}$ m typical of the wet T/R sample used to derive equation (1). Consequently, I conclude that the seismic coupling in salt is more efficient than for any of the other media considered here, in agreement with the observed free-field data.

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16 January 1980

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12

Dear Carl,

Herewith is my response to your request of 18 December 1979 for assessment of the status of siesmic yield determination. It is in two parts, of which the first is concerned with aspects of close-in free field measurements relevant to seismic source evaluation and the second considers the significance of close-in surface motion data to generation of surface waves.

This response took longer than I had anticipated, in part, because of a new approach I have applied to estimating the energy developed in spall closure impact. The results for the Milrow and Boxcar events imply that the impact energy is probably negligible compared to the P-wave source energy; of the order of 1/6 to 1/10 of the seismic source energy derived from free field data. This approach might be profitably employed in analysis of surface data from other events.

The reference list attached may seem egotistical but my excuse is that each of these reports is pertinent to the issue and involves data with which I am intimately acquainted.

I am particularly interested in the derivation of source energies and reduced displacement potentials from free field data and of surface wave initiation and spall impact energies from surface motion data and think that I may be able to contribute effectively in these areas.

Sincerely,

Bill

William R. Perret

EVALUATION OF BODY WAVE COUPLING THROUGH EXPERIMENTAL DATA.

Free field data from the zones of nonlinear and linear response surrounding a contained underground nuclear explosion are significant to seismic source estimates: (a) for evaluation of attenuation patterns, (b) as models for interim stages of seismic source computer codes, (c) for definition of magnitude and form of seismic source functions through energy and reduced displacement potential calculations and (d) for estimating elastic radii.

Seismic source parameters are assumed to be defined at the transition from nonlinear to linear response of the earth. This transition has generally been identified with (a) a decrease in the rate of peak radial stress, i.e., particle velocity, attenuation from the inverse second power to the inverse first power of radial distance, (b) the onset of constant calculated energy with increasing distance and (c) corresponding constancy in derived reduced displacement potential with increased distance. Application of these criteria to real data involves approximations, particularly since the transition is not a discontinuous one.

Energy estimates are derived by computing energy flux from radial particle velocity records and summing this flux over a spherical surface of radius equal to the radial range at which the velocity was recorded.

Reduced displacement potential data, derived from integrals of radial displacement records are as reliable as the records in which peak values are generally good, but residuals are often uncertain.

Free field ground motion has been recorded for at least 60 contained nuclear explosions from the Rainier event in 1957 through 1970[1]. Some of these projects involved only vertical arrays of gage stations above the shot point; others included horizontal arrays at, above and below shot level. In general data from vertical arrays above shots are not suitable

for seismic source evaluation because signals reflected from the surface distort records at depths within the region of non-linear response. Some horizontal arrays did not extend into the region of linear response and data from them are of doubtful value to seismic source studies.

Of the free field data known to me[1], those from the following events include station ranges adequate to seismic source evaluation: Merlin[2] (alluvium); Rainier[3] (tuff); Hard Hat[4], Shoal[5] and Pile Driver[6,7] (granite); Handcar[8], Gasbuggy[9] and Discus Thrower[10] (layered sediments); Salmon[11] and Sterling[12] (dome salt); Boxcar[1] (volcanics). Borderline data sets which might be useful are those from: Fisher and Haymaker[13], Mud Pack[14] and Events C, D and N of Reference 1.

Seismic source energies have been calculated for 24 events[15,16] and have been used to estimate coupling efficiencies of four rock types. Reduced displacement potentials have beenderived for numerous events; some included in the referenced reports and others reported in geophysical journals. Not all potentials considered definitive of seismic source functions in the latter reports were derived from gages within the linear response region.

Those free field data which were recorded by Sandia Laboratories exist in Sandia archives as the original analog FM tapes and in many cases as adjusted digital final data tapes also. I have no information concerning the status or availability of free field data obtained by other organizations.

William R. Perret

EVALUATION OF NEAR-SOURCE SURFACE DATA.

Through 1972, surface motion data at surface-zero and at various distances out to 22.5km have been recorded for more than 57 contained nuclear events and then for many more of which I have no specific knowledge. Most of the former data sets include records from horizontal ranges of the order of 2 to 3 times shot depth, but a few extend well beyond that limit.

Displacement hodographs of surface motion in the vertical-radial plane have been derived from surface records for numerous events[11,17,18]. Onset of retrograde motion in these hodographs has been interpreted as the start of Raleigh waves, especiallywhen that phase increases in duration and amplitude with distance. This phase becomes evident in many hodographs at horizontal ranges equivalent to half to one times shot depth.

Spallation of the earth above an explosion is identifiable in vertical surface motiion records by distinct signatures. These data have been used in a few instances to define very roughly the lateral extnet of spall.

Spall-closure impact is considered a possible source of seismic surface waves, but several factors tend to limit its significance. These factors include: multiplicity of spalls, lateral extent of spall, thickness of spall gaps and spalled layers, rock type, and sequence of spall openings and closures. New calculations based on surface motion records from the Milrow[17] and Boxcar[18] events indicate that assumed simultaneous impact over the entire spalled area produced energy equivalent to approximately 4 to 5kt for Milrow and 4 to 7kt for Boxcar. These calculations disregard the sequential nature of spall impact in both time and space and assume (1) a single spall opening equal to the displacement represented by the negative phase of the surface vertical particle velocity N-wave, and (2) a spalled mass thickness derived from surface-zero data or assumed from other infromation and considered to taper to 1/4 the surface-zero thickness at the most

remote surface station which showed positive evidence of spall. All of these assumptions yield conservative, i. e., higher, energy estimates suggesting that the results quoted may be excessive by a factor of two. Thus as a source of a body wave or surface wave phase, the spall impact is probably real but minor.

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NEAR-SOURCE EFFECTS ON P-WAVES

by Don Helmberger

Although we have been monitoring explosions in the near-field
for many years it still is not clear that we know much about the
effective source description. In fact, recent numerical results indicate
that most near-in data is probably within the non-linear zone making
the conventional RDP's questionable and secondly, the latest evidence 13
for regional changes in frequency dependent Q's makes it difficult to
estimate RDP's from teleseismic data sets. Thus, the various studies
relating very-near-in data (distances less than a source depth) of the
type summarized by Murphy (1978) to teleseismic data becomes even more
difficult to access. It appears that a better appreciation of local
observations at distances 2 to 20 source depths can help clarify the
situation that is, at these ranges the non-linear effects should be less
significant and a more reliable RDP obtained. Unfortunately, separating
the propagational distortions and secondary disturbances from the
effective RDP at these ranges is no easy task.

Although there have been numerous U. S. Coast and Geodetic Survey's Special Projects data collections at local ranges by King (1969) and his associates, the three-component measurements reported on by McEvilly and his students seem to be the most accessible, see Peppin (1977). The best results reported on to date were obtained for three shots fired on Pahute Mesa, normaly Jorum, Hadley and Pipkin. The first two of these were recorded at a constant range of 8km but at several azimuths and the latter event at several azimuths ranging from 2 to 12km. The observations from the first two megaton shots show considerable azimuthal differences on the horizontal components which can be interpreted

as tectonic release or in terms of moment tensor components, see Stump and Johnson (1977) and Stump (1979). These results are produced by a powerful inversion technique that determines the time functions and moment strengths necessary to fit the observed waveforms with assumptions about the local crustal structure. A halfspace model was used in these preliminary attempts so that complications produced by structure are forced into source excitation.

A somewhat less ambitious analysis of these observations is given by Hadley (1979). In this study, it was noted that the vertical components were, in general, much stronger than on the horizontals at the record onset which was interpreted as caused by the direct P-ray diving into the faster substrata and reaching the surface at a much steeper angle than predicted by halfspace models which suggest nearly complete radical motion. Attempts at modeling the Pahute Mesa structure by a flat layered stack are presented and synthetic seismograms compared with the observations over the first few seconds of motion. An effective RDP(t) for Jorum was produced following a trial-and-error procedure where a trade-off between overshoot in the RDP(t) against the strength of pP(t) occurs making a unique answer difficult. It would appear that perhaps some new definitions would be useful, with a RDP(t) appropriate for direct P and RDP(t) appropriate for pP since we suspect that pP is deleted in high frequency and generally delayed. Both of these features are produced by numerical experiments conducted with linear wave propagational models, Scott and Helmberger (1980). If one supposes that the surface directly above the source is allowed to be less than perfectly reflecting (energy lost to spall) one obtains the above effects.

We normally assume that RDP(t) is independent of azimuth with some justification, at least McEvilly data is reasonably consistent with this assumption after some geophysical interpretation. Namely, it is observed that direct P changes its strength with azimuth but the ratio of vertical-to-radial motion changes accordingly which apparently is caused by changes in geology, see Hadley (1979). Thus, it would appear that much more data and analysis must be conducted on this type of observation before definitive conclusions can be obtained.

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APPLIED THEORY, INC.

STATE OF THE ART ASSESSMENT: SEISMIC YIELD DETERMINATION

1. Coupling and Source Theory

Computing power is cheap enough to permit accurate integration of the equations of continuum motion for a wide class of interesting sources. But — there's much less to that fact than meets the eye. True, some basic features of material- 14
velocity waveforms are almost medium-independent; to that extent, little penalty is paid for ignorance of the mechanical properties of soils and rocks. Otherwise, however, that ignorance is so deep that conclusions reached in source calculations are to be trusted only if they don't depend much on medium-models — a severe limitation on the usefulness of large-scale computation, whether for body- or surface-wave prediction. Even so, the following properties of deep, nearly-spherical shots (>150 m/kt$^{1/3}$)[1] were discovered mainly by calculating spherical fields of explosively-driven motion:

a) Owing to a decrease in the fraction of the yield lost to the medium as heat, m_b increases at first as R_c is increased from tamped-charge values (the L^3 explanation is different, but the result is similar). Calculated increases in m_b run from ~0.1 to 0.2, with the maximum m_b at R_c-values of 2 to 4 m/kt$^{1/3}$.[2]

b) At the smallest R_c that meets the practical criterion for full decoupling ($P_o \approx P_c$), air has a near-maximum heat capacity.[3] Hence, further increases in R_c cause m_b to increase; R_c has an optimum value, R^*, in the sense that m_b has a minimum at R^*. R^* is probably close to the practical value

of R_c for full decoupling, but is uncertain for real media due to their inelastic behavior at low stress levels (<1 bar).

According to calculations, values of m_b decrease most rapidly in salt at values of R_c from 8 to 15½ m/kt$^{1/3}$, and from 5 to 13 m/kt$^{1/3}$ in granite; m_b drops by ~1.3 on those R_c-ranges.[2]

c) For tamped bursts (and in lesser degree for overdriven cavities), final cavities are held open mainly by locked-in compressive hoop stresses that develop as material fails at its ultimate limit (von Mises limit) of shear strength, early in the shot (<0.03 sec/kt$^{1/3}$). Those stresses extend outward to several cavity radii; elastic compression associated with them accounts for 30% to 50% of the final cavity volume. That's one reason why, at ranges where material behavior is linear, displacements are not simply related to final cavity volumes even for rather simple materials.[2]

d) No better data on ground motion exist than for the Salmon and Cowboy events. Still, the data leave permanent displacements much in doubt. Permanent displacements from gauge records for tamped Cowboy shots — though subject to wide scatter and possible systematic error — show a trend toward values of r^2D_∞ that decrease with r out to the greatest instrumented ranges in the Winnfield dome (~3200 m/kt$^{1/3}$), becoming negative well before that range is reached. A similar result is suggested by D_∞-values for the Cowboy cavity shots, but with still greater scatter. What might cause D_∞ to be <0 at many times the final value of R_c is clear physically; it was seen and explained as a bona fide feature of some fields computed for deep tamped bursts, years before any evidence appeared that D_∞ could actually be <0.[4]

So far, we find that a change from $D_\infty>0$ to $D_\infty<0$ affects RVP-spectral-amplitudes only slightly unless $D_\infty \approx 0$. The main effect is felt in the phase of the RVP-spectrum, which aproaches a 180° reversal as $f \to 0$. For $D_\infty/D_{max} = -\frac{1}{2}$ (the most negative value now within reason) and Salmon yield (5.3 kt), the phase change relative to the case $D_\infty/D_{max} = \frac{1}{2}$ is >45° for f < 2 hz, and >135° for f < 0.3 hz. For $D_\infty/D_{max} \approx 0$, the amplitude decrease exceeds a factor of 2 for f < 0.4 hz (of course, if $D_\infty=0$, the factor grows without limit as $f \to 0$). A further effect is to put in doubt the validity of analyses of coupling in which elastic behavior of the medium is assumed, even for fully decoupled shots; indeed, it seems clear that deformation must be not just inelastic, but nonlinear, to produce the result $D_\infty<0$ — at least in homogeneous, isotropic media.

e) Solid-earth sites show a systematic increase in strength and stiffness with depth, or overburden. The increase in strength matters most, but all depth effects combined don't cause m_b to fall by as much as 0.1 in granite and salt media, as depth goes from 100 to 300 $m/kt^{\frac{1}{3}}$. Depth changes in a single medium have much less effect than differences between media (at the same depth, m_b varied by 0.6 among tamped shots in salt, wet sandstone and three granites).[2]

2. Experimental Data and the Basic New Insights They Give

Study of gauge records from field tests has yielded far-reaching results. The most consistent, credible ground-motion data, as well as the most complete, were obtained from the Salmon and Cowboy events.

Firstly, near-elastic deformation is simply not observed in the field (and hardly ever in laboratory shots). Secondly, bursts in salt domes are spherically symmetric and reproducible, well within any practical monitoring requirement (and most requirements for ground-motion research). Within measurement accuracy — the highest afforded so far by field tests — the rules of simple scaling hold for those shots for yields from 10^{-4} to 5.3 kt, at least.

The persistence of strong inelastic effects to the farthest ranges of ground-motion measurement was not predicted by the computational models used for full-blown nonlinear source calculations; the models conflict with observed nearly-spherical motion in other basic ways as well. More importantly, a linear source of far-field motion has yet to be defined by experiment (while elastic sources may not exist). As a result, sourcemen are left without any strong constraint on the linear near-fields they compute, and propagationmen have no strong constraint on the sources they assume. Thus, given the unknowns, not-well-knowns, and complexity, of both source and propagation models, the ability to produce synthetic seismograms that look realistic has no clear meaning. It also becomes difficult to place meaningful error-bars around the decoupling factors deduced from ground-motion measurements, and to say how consistent

they are with the factors obtained by surface-seismic measurement. Fortunately, the same ground-motion data show that a linear source can almost certainly be defined by conducting CE shots in salt domes, and they provide a firm basis for small-scale simulation of NE shots. Details follow.

a) Scrutiny of Salmon and Cowboy fields for departures from isotropy and homogeneity showed that neither c_P nor U_{max} varied systematically with direction. Maximum deviations from the mean wavespeed with direction lie within measurement error for Salmon (<10%) while for the Cowboy events the maximum deviations (~10%) appear significant. Directional variations in U_{max} at fixed r are not significant for Salmon (≤5%), but they may be for Cowboy (≤30%).[5]

b) Regression fits to Salmon peak velocities and displacements yield power-law exponents of 1.89 and 1.60 respectively, with variances of .05 and .04; variances from the least-squares power-law fits amount to 9% and 7%. For tamped Cowboy shots, simply-scaled to a common yield, the exponents 1.53 and 1.50 (for peak particle velocity and displacement) have variances of .07 and .045; variances from the fits amount to 35% and 19%.[6]

c) From the first[7], values of U_{max} and D_{max} from tamped Cowboy shots, simply-scaled to a common yield, were plotted on log-log paper vs. slant range. So treated, the data don't scatter much about the least squares straight line through them [2.b)], and show almost no systematic dependence on yield. Treating Cowboy cavity shots with a common R_c-value in that way, five more least-squares lines of U_{max} vs. r, and five of D_{max} vs. r, are obtained.[8] If arranged in the order of decreasing values of U_{max} on the r-interval covered by the gauges for each line,

the six slopes show a definite trend toward smaller (less negative) values. The trend is slow and gentle, and the limit actually approached may not be that of elasticity; even for U_{max}-values of ~1 cm/sec (peak radial stresses ~1 bar), the slopes appear at least as negative as -1.2 and -1.25 for D_{max} and U_{max}, respectively.[9]

d) Data from the Cowboy 10 shot, and some theoretical work, led to a value of 2.6 kj/gm for the Cowboy CE (a pelletized form of TNT; mean density 1 gm/cc).[10] The nominal value was 4.2 kj/gm. Shortly thereafter, it was found by direct calorimetric measurement at L^3 that powdered TNT of density 1 gm/cc gave an energy of 3.6 kj/gm when heavily clad, and 2.45 kj/gm when allowed to expand freely into a large cavity; at normal density (1.63 gm/cc), heavily clad, the available energy of TNT is 4.6 kj/gm.[11]. It appears that the energy released on expansion of oxygen-deficient CE's will generally be path-dependent, in which case yields can only really be pinned down by following actual shot-paths. Thus, the energy actually released in tamped Cowboy shots may still be uncertain by ±10%.[12]

e) The compression of Cowboy data under simple scaling rules suggests in a small way that those rules might apply to dome-salt; tamped Cowboy charges weighed from 20 to 1003 lbs, and the largest cavity charge weighed 1902 lbs. When scaled to Salmon yield (assuming that 1000 English tons of Cowboy CE yields 1 kt of energy), the Cowboy gauges cover the range-interval from 280 to 5540 m; Salmon gauges run from 166 m to 744 m. The straight lines that fit closely the Salmon and tamped Cowboy data for log D_{max} vs. log r, are nearly parallel [see 2.b)], with the Salmon line above that for Cowboy. Thus, elastic and other processes that determine the rates of decay

of the similarly-shaped Salmon and Cowboy displacement pulses, are sensibly scale-independent. Indeed, the two lines virtually became one when the yield of Cowboy CE was corrected [see 2.d)]. Near-coincidence of those lines is surprising; HE and CE sources surely differ in the motions they cause at very small ranges (where ground motion can't be measured by gauges of Salmon and Cowboy type). How and why they merge at ranges less than 10 final cavity radii, should be an absorbing tale.

Salmon and tamped-Cowboy velocity pulses are also generally similar in shape, but have some persistent differences that show up in log-log plots of U_{max} vs. r. Again, the Salmon line lies above the tamped-Cowboy line, but its slope is significantly more negative [2.b)]. However, at least half the difference in slope vanishes if the Salmon velocity pulses, simply-scaled to Cowboy yields, are passed through the Cowboy gauges and scaled back to Salmon yield.[12] The remaining slope-difference probably has statistical meaning; it may be a sign that the higher-frequency components reflected in U_{max} (as opposed to D_{max}) don't simply-scale, but it's just as well explained by the slow turning of U_{max}-vs.-r curves toward elastic (1/r) decay [2.c)].

f) If salt deforms elastically for all $r > r_o$, then motion at r_o determines, without approximation, how a spherical explosively-driven field develops beyond r_o. Thus, for each of several measured velocity pulses, we computed the values that U_{max} and D_{max} would have had as functions of r, if material had behaved elastically beyond the range of any given gauge ("elastic extrapolation").[13] With negligible error,

all the resulting curves of U_{max} vs. r, whether extrapolated from close-in or far-out gauges, proved to be inverse-range curves — a result in conflict with measurement [see 2.b)]]

Obtained by elastic extrapolation of the closest-in pulses measured, D_{max} falls more rapidly than 1/r at first, but soon becomes sensibly proportional to 1/r. From gauges at larger ranges, elastically-extrapolated pulses gave D_{max} as virtually proportional to 1/r. Those results also conflict with measurement; in reality, D_{max} unmistakeably falls more rapidly with range than that.

g) The importance of inelastic decay of D_{max} for body-wave monitoring was established by breaking measured pulses into decade-wide harmonic bands, and transforming each band back to the time domain.[13] It was then evident that D_{max} is determined predominantly (at Salmon yield) by .5-5 hz components of the outgoing wavetrain — the core of the body-wave-detection band. Further, between .75 and 4 km, motion decays in amplitude for that band at an average rate equivalent, in the patois of seismology, to Q=3. Predictably, U_{max} is dominated by components of higher frequency, but still serves to signal inelastic deformation.

Squeezing the data harder leads to two major conclusions that must remain tentative until more complete and consistent data are available than the Cowboy events gave: (i) Well beyond the range of Salmon measurement, pulse-decay in salt is nonlinear, and not just inelastic. Specifically, for the 1-10 hz band, rates of D_{max}-decay found by elastic extrapolation tend to be greater near their ranges of origin than those found for that band directly from the measured pulses themselves. It thus appears that energy is fed from other bands, presumably of higher frequency, to that one.

(ii) At the farthest ranges from the over-decoupled Cowboy 10 shot, decay of the .3-3 hz band may be effectively elastic.

h) When simple scaling rules hold, NE-CE equivalence factors become complex functions of scaled range and frequency.[14] At ranges (if any) where the medium also behaves elastically, equivalence factors become complex functions of scaled frequency alone. For practical purposes, the functions in question may assume simple forms, but not as simple as the single-number description generally used.

3. Source Theory; Surface Waves

In calculating surface wave sources, symmetry is limited to rotation about a vertical axis through the shot point. As symmetry is reduced, the class of material deformations encountered widens. Hence, the demands made on material models in computing surface wave sources are greater than for spherical motion - and the caveats of Section 1 apply here a fortiori.

Demands on the numerical art are also relatively heavy, since motion must virtually have ceased at the final time of calculation. Well before that, live stresses are much smaller than overburden stresses, almost everywhere. A method was found to assure that computed fields would approach numerical conditions of equilibrium in the late-time limit, with gravity correctly accounted for, and even if stress-strain relations are always nonlinear [2,c),f)]. Computer cost and storage limitations proved a more severe problem - a problem not foreseen, and closely tied to the finding that fracture of near-surface material is a prime feature of near-fields from shallow-buried ($<150\mathrm{m/kt}^{\frac{1}{3}}$) 150-kt bursts.[15] Cracking, and especially its nonlinear effects, gave rise to some of our main results. A slowdown in near-field evolution is one such result — but it's also a source of practical computing trouble. Specifically, conjecture that notable motion would last only for ~1 sec after burst, gave way to actual computation of 2-3 sec of motion. Even that may not be long enough, and calculation can't be carried past ~3 sec without substantial computer code revision.

Given tighter numerical and material properties limitations, more care than usual must be taken in drawing conclusions from calculations. Nevertheless, new conclusions of note have emerged, as follows:

a) For yields in the 150-kt range, and burial depths to at least 150m/kt$^{\frac{1}{3}}$ (which covers common U.S. and Soviet test practice), inelastic processes triggered by reflection dominate the reflected wave. In particular, the waveform is altered because (i) cracking limits reflected tensile stresses to about overburden-levels, and (ii) reassembly of cracked material under gravity and weakened continuum stresses, takes much longer than does radiation of energy out of an elastic near-field. Further, c_p is cut sharply because the progress of signals from the ground surface is repeatedly interrupted by cracks; as a result, elastic treatment of reflected waves (pP-reflection in particular) leads to overestimates of burial depth. For actual burial depths of 250 to 700 m in sedimentary rocks, the depths estimated elastically would be too large by a factor that decreases from 2.5 to 1.4.[2] [Yields would also be overestimated for bursts in media that behave like the "wet sandstone" of the calculations. On a plot of m_b vs. depth, typical U.S. tests fall near a minimum when cracking effects are taken into account, while m_b increases monotonically with depth (to 150m/kt$^{\frac{1}{3}}$) if reflection is elastic.]

It appears that omission of cracking causes a major drop in M_s, a result that will be tested as more experience is gained with M_s-computation.

b) The amplitude of the wave striking the ground surface from below depends strongly on nonlinear properties of the near-field medium. Hence, while incident waveforms don't vary much, and the tensile strengths of almost all geologic solids appear negligible, nonlinear reflected-wave effects are medium-dependent. For a rock like granite, the deduction of burial depth from seismograms, and depth corrections to m_b, are more likely to succeed

than for a sedimentary rock; such key properties as air-filled porosity and shear strength vary more from site to site in the latter. Also, if scaled burial depth be fixed, overburden suppresses cracking more and more as Y increases. That effect is probably outweighed in real media (at least for yields in the low-kt range) by more-nearly-elastic behavior at greater depths, which causes a stronger pulse to reach the surface.

c) For many cratering bursts, it looks possible to deduce yield from crater dimensions to within a factor of 1.5-2. Specifically: (i) Crater radii vary slowly as shot-depth runs over most of the interval from zero to containment (55-75 $m/kt^{1/3.4}$, depending on medium); across more than half that interval, radii change by only 25%.[16] (ii) Except for near-surface and nearly-contained shots, crater radii appear reproducible to within 20% for a given depth and yield. (iii) Using the known shape of the Sedan crater, but not its known dimensions, its radius was "predicted" with an error <20%.[17]

A key step in our Sedan calculation was to adjust the properties of the medium to match the ground-surface jumpoff velocity, as measured (to ?%) above the shot. Jumpoff velocity (not likely to be told us for others' shots) is predictable if we know how U_{max} decays in the shock driven to the ground surface — but for more media than "dry hard rock", "wet rock", etc. Luckily, since U_{max} falls in simple power-law fashion [2.b)], even lightly instrumented CE tests (lab-scale included) can establish its decay experimentally. That's the only way; while they might have some diagnostic value, calculations of U_{max} can't be trusted (Section 2). Calculations are needed, though, to extrapolate from the few media in which crater di-

mensions are known as functions of yield and shot depth, to other media. Success in that task rests on the weakness of soils and rocks in tension, and knowledge of jumpoff velocity.

As near-surface and nearly-contained limits are approached, yield estimates will become less accurate. Shots near those extremes of burial depth might be recognized by (i) the relatively broad, flat craters they produce, and (ii) observations of other quantities, such as residual radioactivity.

The insensitivity of crater radius to burial depth, a boon to yield estimation, makes it hard to assess burial depth. Nor will crater depth turn the trick; it's much less reproducible than radius. Thus, the prospects seem dim for deducing burial depth from crater dimensions, to within ± 20 m/kt$^{1/3.4}$; that's not much better than ± 35 m/kt$^{1/3.4}$ which follows from the mere fact that a crater formed.

4. Next?

Definition of an actual linear source is the single most critical need in NMR; along with that goes the search for a regime of tangibly-elastic behavior. Both tasks are chiefly experimental; moreover, the field data now at hand make it plain that they should be carried out first for dome-salt. To the extent that simple scaling holds [2.e)], CE shots alone will suffice for tamped bursts; Salmon has provided the NE data with which to define equivalence factors.

Only with an experimentally-certified linear source can monitoring research be split cleanly into source and propagation parcels (see start of Section 2. above). The same shots can also show us what significance to attach to elastic potentials, analyses and models. Further, surface-seismic data can be obtained, and with it new insight into the transformation of pulses from source to seismometer (insight that could influence Salmon-Sterling conclusions). A key task in the events should be the measurement of D_∞ at several ranges, partly to see whether it stays positive, but mainly to correct objectively for the drift that characterizes records from ground-motion gauges. In addition, the energy released by several CE's, including Pelletol, should be determined from in situ shots by making passive measurements of cavity volume — shots that will also show, for the yields covered, how closely the resulting cavities follow simple scaling rules. Careful post-shot surveys of the medium around the shots would yield data to compare to that found for salt near the Salmon cavity; in that way the stage would be set for deciding whether differences between mined and explosively-formed cavities can be determined by simulation.

Implementation of the tasks noted is either under way or planned. Beyond them, four paths appear most probable:

(i) CE simulation of decoupled NE bursts;

(ii) repetition of the main salt shots, but in granite stock;

(iii) performing a final CE shot in dome-salt at much-increased yield, as a direct check on the accuracy of simple scaling rules;

(iv) more tamped salt-dome-shots in the 200-1000 lb range, either to fill gaps in data from the initial shots, or to pursue unexpected results.

Computers would play a major role in designing tests of type (i); NE-driven motion of fluids (and air in particular) has been calculable since the '50's. Motion from cavities overdriven by NE could also be simulated, provided either that cavity-wall motion is negligible, or that observed and calculated wall-motions differ little. Data from all shots should be used to evaluate the results of laboratory-scale simulation, which would continue apace; in fact, lab "prediction" experiments would go a long way toward establishing what can and can't be learned from tests at that scale.

Path (ii) should be followed at NTS, since direct comparison of NE and CE data might then be possible; NE shots in granite, in particular, would serve several purposes besides linear-source-definition. Bursts in cavities, both mined and explosively-formed, are among the events that could be conducted for both NE and CE at NTS; that can't be done in dome-salt. In any case, NE events should take place only after CE tests that fix the minimum dimensions of a linear source — information vital to rational NE-test design. Requirements for ground-

motion instrumentation would also be set thereby. Of the drawbacks to such an NTS program, we must cite these, at least: Adherence to the rules of simple scaling is a material-specific matter, and seems less likely for a cracked and jointed medium like NTS granite than for dome-salt. Cracks and joints also make the medium inhomogeneous over distances of ~1 m; many more ground-motion measurements will be needed than in salt to assure that a good picture of the field is obtained despite local aberrations. Further, there is a more-than-slim chance of observing motion that can't reasonably be viewed as spherically symmetric; that contingency must be covered, which again multiplies gauge-related costs.

Path (iii) needs to be taken to put all simulation work on a rigorous footing, and to provide NE-CE equivalence factors free of any substantial guesswork. Until that's done, a fundamental gap will remain in the arguments by which both linear sources and equivalence factors are defined. No amount of computing or theorizing will tell us whether scale effects on cavity growth are big or small when yields are varied from 10^{-3} to 5.3 kt (even though, beyond reasonable doubt, motion simply-scales in dome-salt at ranges greater than ~10 final cavity radii). At 2 cavity radii, for example, the question is open: Salt may be much stronger at the higher strain-rates of tamped Cowboy shots than for Salmon. If so, then simple-scaling-factors will have to be adjusted upward in order to reach correct conclusions about NE bursts from motion measured in lower-yield CE shots. Contrariwise, it might turn out that even the results of lab tests can be used without such adjustment.

Tests in NTS granite would serve directly to (i) expand the base of surface- and body-wave data for granite media, and

(ii) tell us what the effects of burial depth on those waves really are [3a),b)]. At CE-test scale, the impact of gravity on ground motion is negligible; overburden is then absent, which enhances cracking — but since scaled distances between cracks increase as yield is lowered, tensile strength tends to increase as well. Still, major features of inelastic reflection should be present in 10-ton bursts, namly, quenching of the pP-wave and late arrival of the weak reflection that replaces it. At that scale, however, a sizable risk is also run of learning nothing: Gauge records could be rendered incoherent by local field variations (witness Mine Shaft). To see how data-quality depends on yield, and what the chances are of observing coherent near-fields from nuclear bursts, preliminary CE shots at 1 and 100 tons would be well-advised (they would have other uses too; above).

NOTES AND REFERENCES

1. Abbreviations: m = meters; sec = second; gm = gram; cc = cubic centimeter; kt = kilotons of energy (1kt = 4.184 x 10^{19} erg); P_c = long-term (~¼ sec/$kt^{1/3}$) pressure of cavity gases in a contained shot; P_o = overburden stress; R_c = scaled cavity radius (m/$kt^{1/3}$); m_b = body-wave magnitude; D_∞ = permanent displacement (radial or vector); ω = angular frequency (radians/sec); f = cyclic frequency in hertz (hz); CE = chemical explosive; NE = nuclear explosive; r = slant range; U_{max} = maximum radial particle velocity; D_{max} = maximum radial particle displacement; c_p = longitudinal wavespeed; kj = kilojoules (1 kj = 10^{10} erg); L^3 = Lawrence Livermore Laboratory; NMR=nuclear monitoring research; NTS=Nevada Test Site.

2. Being model-dependent, the quantitative data quoted are subject to change. The qualitative result does not appear model-dependent.

3. L. Doan and C. Nickel, "A Subroutine for the Equation of State of Air", RTD (WLR) TM-63-2, Air Force Weapons Laboratory, May 1963. This statement is firm because, (i) unlike geologic solids, air has well known mechanical properties, and (ii) very little explosive energy is lost to the surrounding medium in a fully decoupled shot.

4. Progress letter to Maj. T. Stong of DNA from Applied Theory, Inc. for the period 14 February 1974 to 15 July 1974 under contract DNA001-74-C-0238.

5. J. Trulio, S. Bless and G. Voland, "Isotropy of Natural Salt: Implications of Underground Explosive Testing", ATR-74-41-1, Vol. II (Final Report; Contract DNA001-74-C-0238), Applied Theory, Inc., September 1975. The Cowboy maxima quoted are probably overestimates; frequency response limitations and systematic differences between the two main types of gauge deployed have only recently been quantified.

6. W. Perret et al, "Free-Field Particle Motion from a Nuclear Explosion in Salt, Part I, Project Dribble, Salmon Event", VUF-3012, Sandia Laboratory, November 1967. RMS deviations from the regression fits were computed by N. Perl at ATI.

7. B. Murphey, "Particle Motions near Explosions in Halite", SC-4440(RR), Sandia Corporation, June 1960.

8. J. Trulio and N. Perl, "Simple Scaling and Nuclear Monitoring", ATR-77-45-2 (Final Report; Phase IV: Contract No. DNA001-75-C-0304), Applied Theory, Inc., April 1978.

9. Except for one case in which the displacement data are questionable, departures from a smooth decrease in decay-rate, toward elastic behavior, lie within measurement error (Ref. 8, p. 23).

10. Ref. 8, Section 2.8.1.

11. Private communications from the chemistry staff at L^3.

12. Talk given by J. Trulio at meeting of 7-8 February 1979 at Darpa Hqtrs. There is now little doubt that frequency-response limitations of the Cowboy gauges appreciably reduced the U_{max}-values from tamped Cowboy shots - as Perret hinted they might (Ref. 6, p.113). The uncertainty that remains stems from incomplete documentation of the frequency-response characteristics of those gauges. Salmon gauge-response was satisfactory; Cowboy pulses hardly changed on scaling them to Salmon yield, passing them (scaled) through Salmon gauges, and scaling them back to Cowboy yield.

13. Letter of April 3, 1979 to C. Romney from J.Trulio, and attachments.

14. Ref. 8, Section 2.7.

15. N. Perl, F. Thomas, J. Trulio and L. Woodie, "Effect of Burial Depth on Seismic Signals", Volume 1, PSR Report 815 (Final Report for Contract DNA001-76-C-0078; May 1979).

16. H. Cooper, "Estimates of Crater Dimensions for Near-Surface Explosions of Nuclear and High-Explosive Sources", RDA-TR-2604-001 (September 1976).

17. N. Perl and J. Trulio, "Effect of Burial Depth on Seismic Signals", Volume II, PSR Report 815 (Final Report for Contract DNA001-76-C-0078; May 1979).

SYSTEMS, SCIENCE AND SOFTWARE

15

BODY WAVE COUPLING THEORY

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The approach taken in deriving the Mueller/Murphy source model (Mueller and Murphy, 1971) was to ignore the details of the energy propagation in the nonlinear regime and use near-regional and free-field empirical seismic data to infer an analytic approximation to the nuclear seismic source function as well as general scaling laws which can be used to describe the variation of the source function with yield and depth of burial in a given source medium. One of the limitations of this approach is that calibration data from an explosion in a particular medium are required to form a base from which extrapolations can be made to other explosions in that same medium. At the present time, the model has been calibrated for explosions in salt, granite, wet tuff/rhyolite and shale emplacement media.* These models have been extensively tested against the available free-field, regional and teleseismic data measured from explosions in the various media and it has been demonstrated

* It appears that adequate calibration data are now available for alluvium and dolomite source media, but the data have not yet been analyzed to define source models for these media.

that they are reasonably consistent with the available observational constraints (Murphy, 1977; Murphy, 1978).

The theoretical, far-field P wave displacement spectra predicted at a fixed yield (100 kt) and depth of burial ($h = 122W^{1/3}$ m) for these four media are compared in Figure 1 (Murphy, 1978). Now, these spectra have been computed by assuming that the source medium is infinite in extent. Bache et al. (1975) have proposed an approximate correction for the effects of local crustal structure which is implemented by multiplying the far-field displacement spectra by the square of the compressional wave velocity in the source medium. The results of applying this correction to the spectra of Figure 1 are shown in Figure 2. Now, assuming that the correct relative coupling factors lie somewhere between those of Figures 1 and 2, the following conclusions can be drawn. First, in agreement with the early findings of Werth and Herbst (1963), salt media such as those represented by the Salmon and Gnome events are predicted to couple the best of the four media studied here. Second, in the frequency band around 1.0 Hz, which defines the relative m_b value, the spectral amplitude values for wet tuff/rhyolite, granite and shale differ by less than a factor of 1.5. Thus, the m_b/yield curves for these three media are not predicted to be very different, in agreement with the observed trends.

The predicted yield and depth dependence of the seismic source functions for the four media discussed above are essentially identical. They result from introducing a depth dependence into the familiar cube-root scaling laws to take into account a variety of observations which suggest that the elastic transition pressure is proportional to the overburden pressure (Mueller and Murphy, 1971; Murphy, 1977). Note that for explosions at a fixed depth, the scaling model reduces to simple

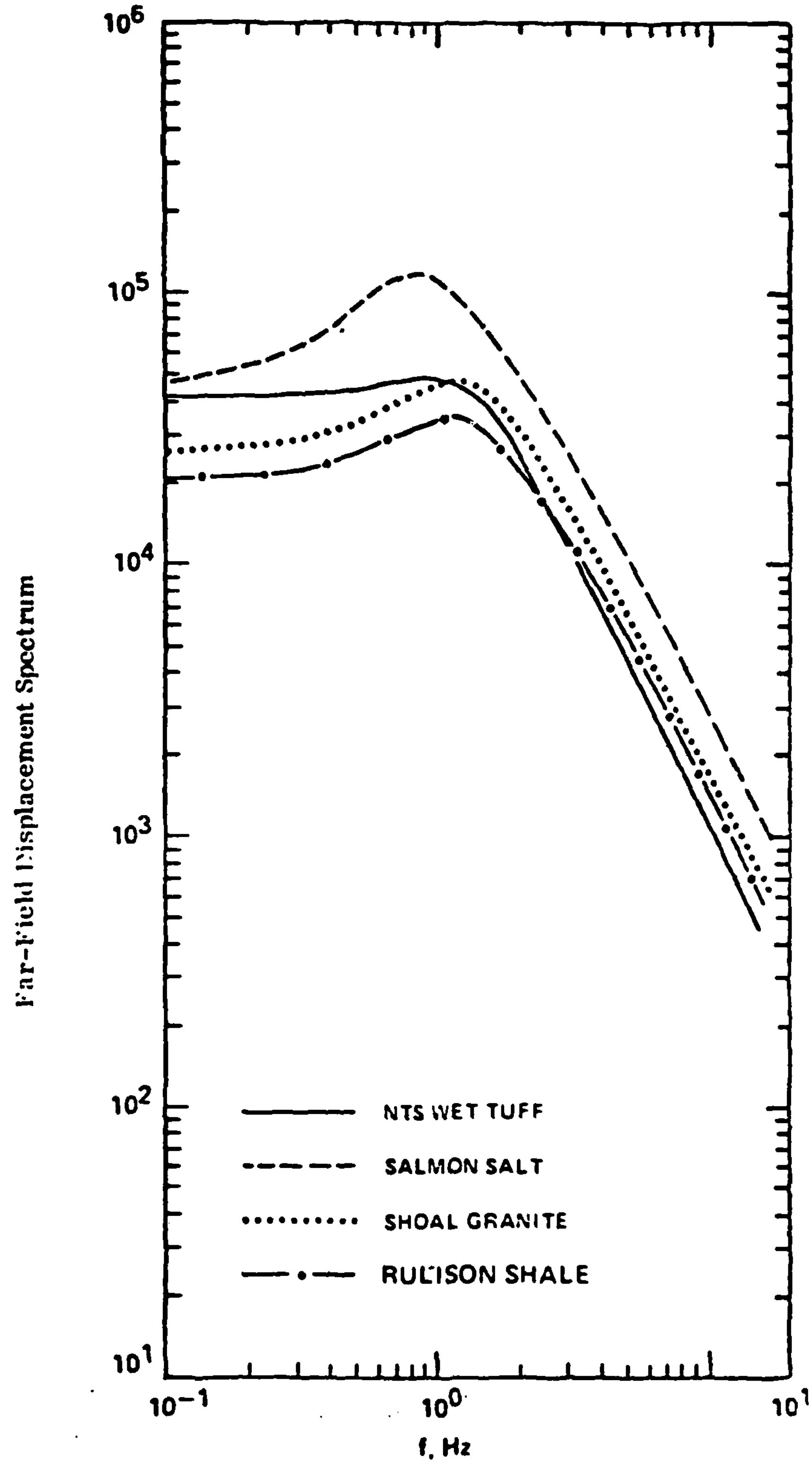

Figure 1. Far-Field Displacement Spectra as a Function of Source Medium, W = 100 kt, h = $122W^{\frac{1}{4}}$ m.

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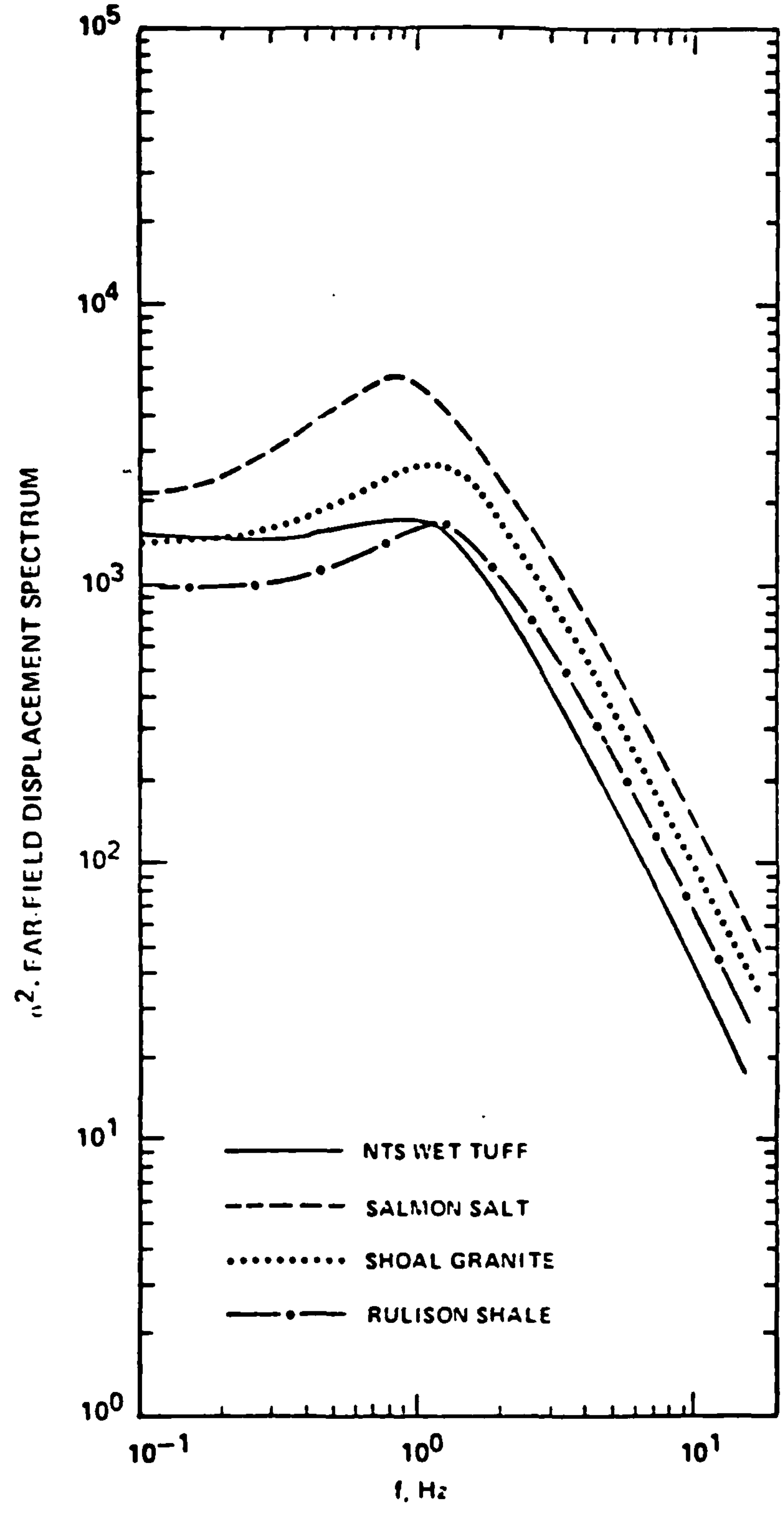

Figure 2. Far-Field Displacement Spectra as a Function of Source Medium, Corrected for Near-Source Crustal Structure, W = 100 kt, h = $122W^{1/3}$ m.

cube-root scaling. In this case, considering the far-field displacement spectrum, the low frequency spectral amplitude level is directly proportional to yield (W), the high frequency spectral amplitude level is proportional to $W^{1/3}$ and the corner frequency is proportional to $W^{-1/3}$. On the other hand, Figure 3 shows the predicted yield dependence of the far-field displacement spectrum at a fixed scaled depth ($h = 122W^{1/3}$ m) for explosions in wet tuff/rhyolite emplacement media. Here, the low frequency spectral amplitude level is proportional to $W^{0.76}$, the high frequency spectral amplitude level is proportional to $W^{0.5}$ and the corner frequency is proportional to $W^{-0.20}$. It has been shown (Murphy, 1977) that these modified scaling laws have significant implications with respect to the short period magnitude/yield relationship and that the resulting modifications with respect to a cube-root-scaling-based magnitude/yield relation are in good agreement with the observed data.

The predicted dependence of the far-field displacement spectrum on source depth at a fixed yield (10 kt) is illustrated in Figure 4 for explosions in salt. It can be seen that at a fixed yield the corner frequency predicted by the model increases with depth approximately as $h^{0.5}$ while the low frequency amplitude level is predicted to decrease with increasing depth approximately as $h^{1/3}$ and the high frequency amplitude level is predicted to increase with increasing depth approximately as $h^{0.5}$.

The source spectra discussed above have been combined with the yield and depth scaling laws to define theoretical m_b/yield curves for the various media (Murphy, 1978). The theoretical Western U.S. m_b/yield curves computed for explosions in salt, granite, wet tuff/rhyolite and shale are shown in Figure 5 assuming $h = 122W^{1/3}$ m, $t^* = 1.0$. For purposes of the comparison with the observed data, the absolute level of these

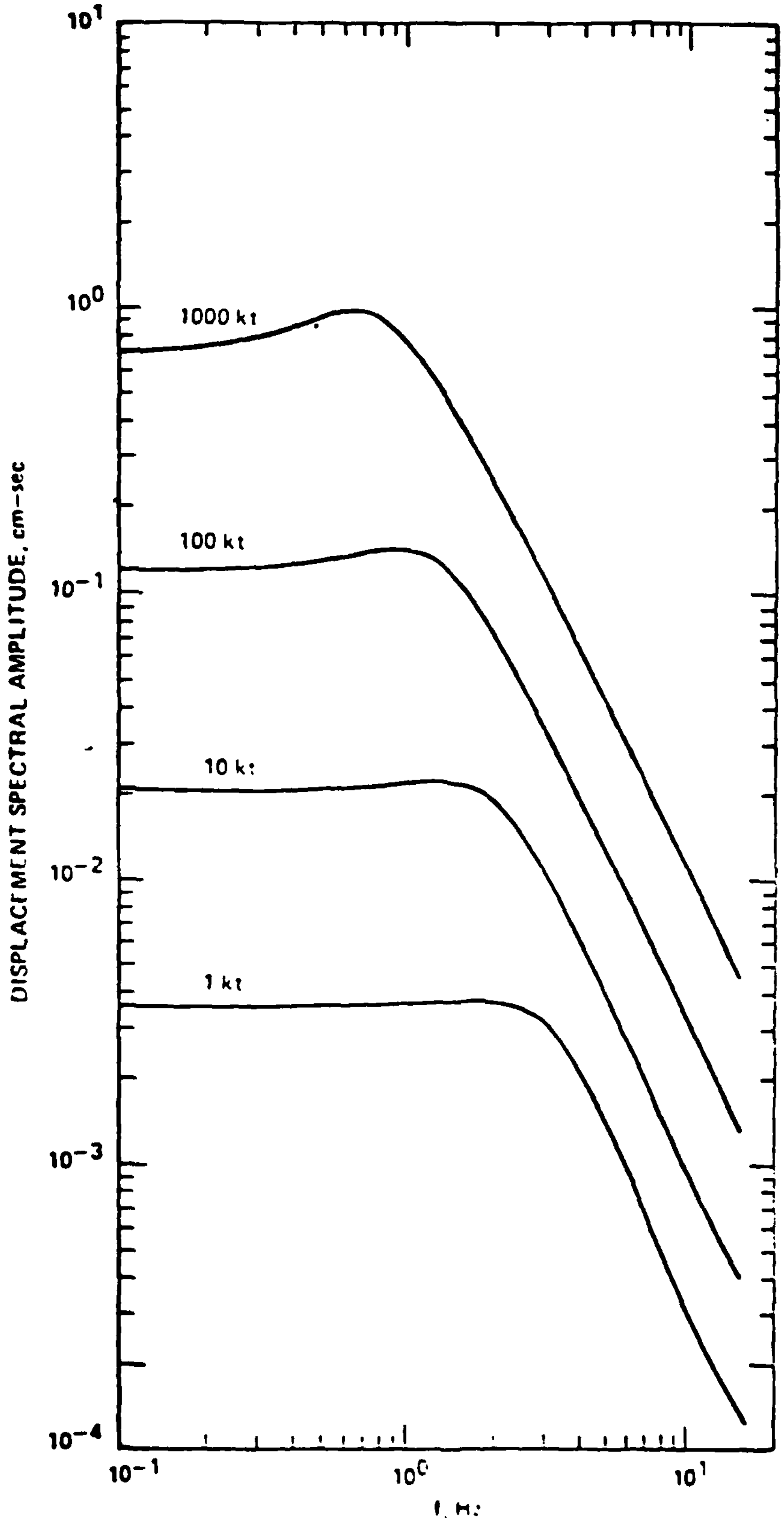

Figure 3. Far-Field Displacement Spectra as a Function of Yield, NTS Tuff/Rhyolite, $h = 122W^{1/3}$ m.

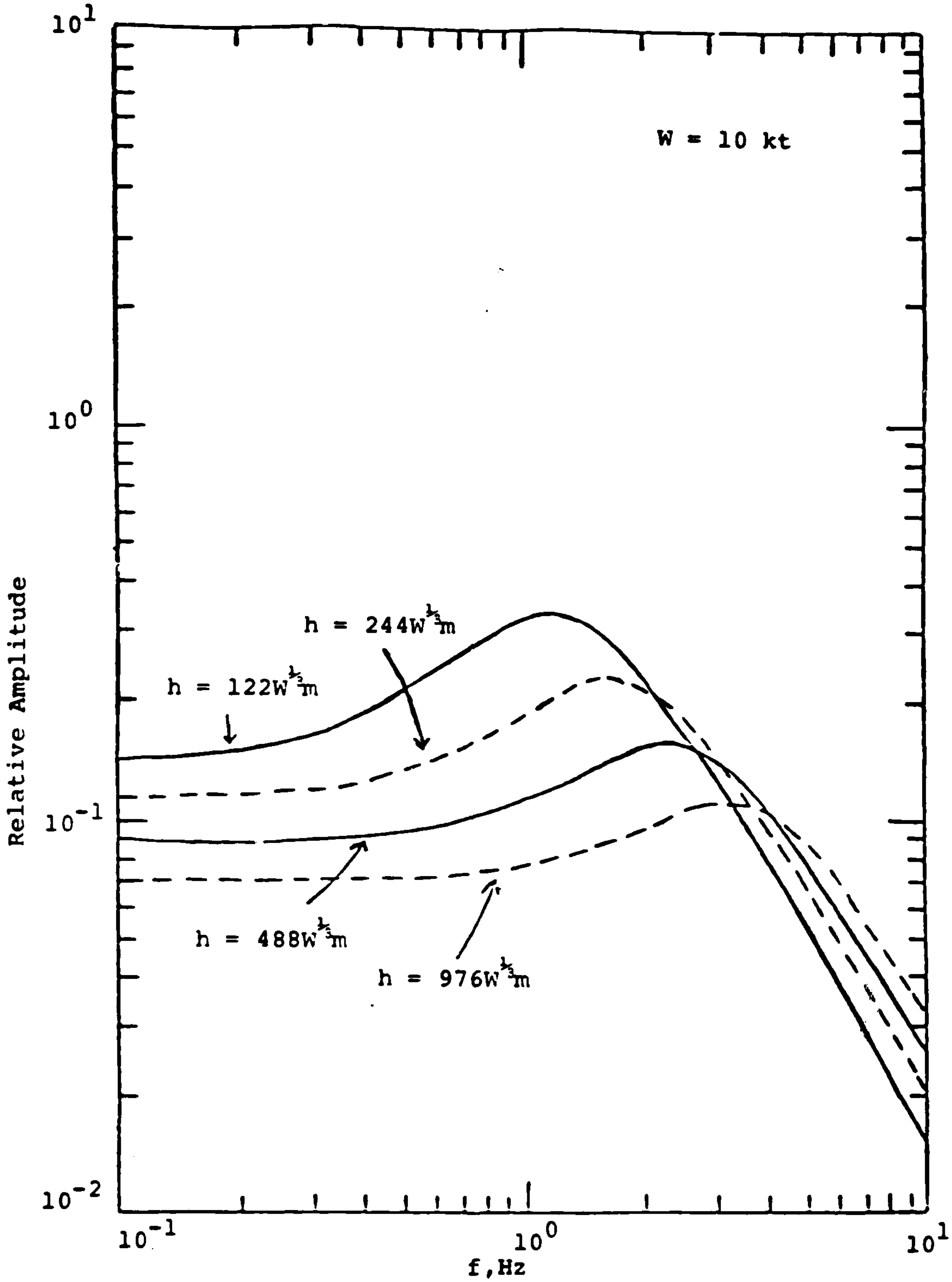

Figure 4. Far-Field Displacement Spectra as a Function of Source Depth, Salmon Salt.

Murphy

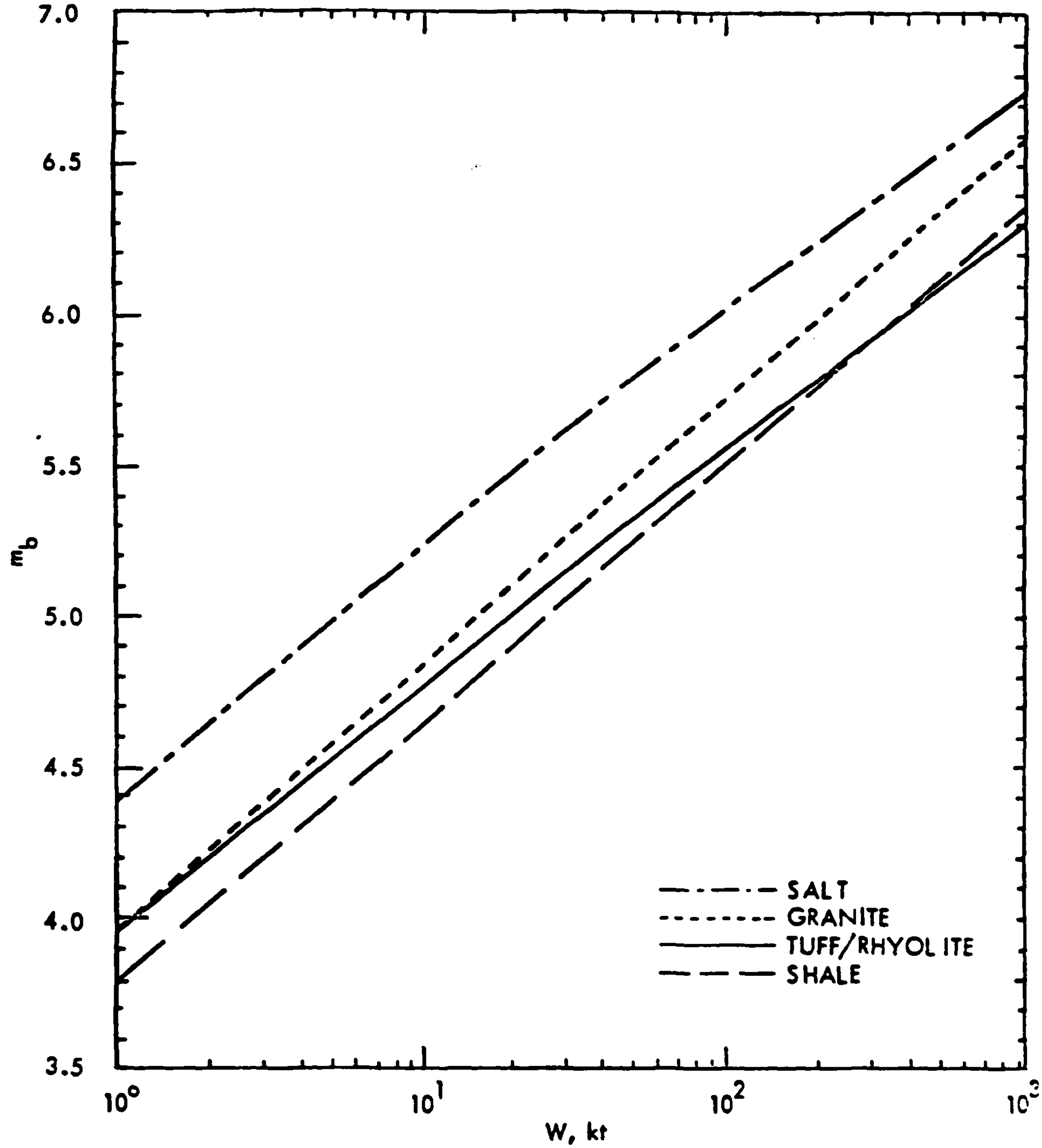

Figure 5. Theoretical Western U.S. m_b/Yield Curves, $h = 122W^{1/3}$ m.

curves has been set such that the theoretical tuff/rhyolite m_b value at W = 100 kt, h = $122W^{1/3}$ m is 5.54, in agreement with the average observation. As might be expected from the spectral comparisons shown in Figure 1, the predicted m_b/yield curves for granite, wet tuff/rhyolite and shale are quite similar, differing by less than 0.2 magnitude units over the yield range from 1 to 1000 kt. The salt curve, on the other hand, is offset above the other three by an amount which depends on yield and reaches nearly 0.5 magnitude units for yields around 10 kt. At first glance, this seems to be inconsistent with the fact that the observed m_b value for Salmon fell very close to the empirical m_b/yield curve for explosions in wet tuff/rhyolite emplacement media. However, Salmon was deeply overburied and, at least according to the Mueller/Murphy scaling model, would be expected to have a significantly lower m_b value than it would have had if it had been detonated at the normal containment depth of $122W^{1/3}$ m typical of the wet tuff/rhyolite sample. It can be seen from Figure 5 that all four curves are nearly linear in this representation and they can, in fact, be approximated very closely (i.e. within 0.02 units m_b) over the yield range from 1 to 1000 kt by the following correlation equations:

$$m_{b_{salt}} = 4.41 + 0.78 \log W$$

$$m_{b_{granite}} = 3.95 + 0.88 \log W$$

$$m_{b_{tuff/rhyolite}} = 3.96 + 0.78 \log W$$

$$m_{b_{shale}} = 3.77 + 0.86 \log W$$

By way of comparison, the observed Western U.S. m_b/yield curves for granite and tuff/rhyolite are:

$$m_{b_{granite}} = 3.79 + 0.91 \log W$$

$$m_{b_{tuff/rhyolite}} = 3.92 + 0.81 \log W$$

Thus, the slopes of the theoretical and observed m_b/yield curves for these two media are in excellent agreement.

The effect of source depth of burial on the theoretical Western U.S. m_b/yield relation is illustrated in Figure 6 which shows a comparison of the wet tuff/rhyolite m_b/yield curves obtained by assuming constant scaled depths of 122, 244 and 488 $m/kt^{1/3}$ respectively. It can be seen that the level of the curve decreases with increasing depth as would be expected from the spectral examples shown previously in Figure 4.

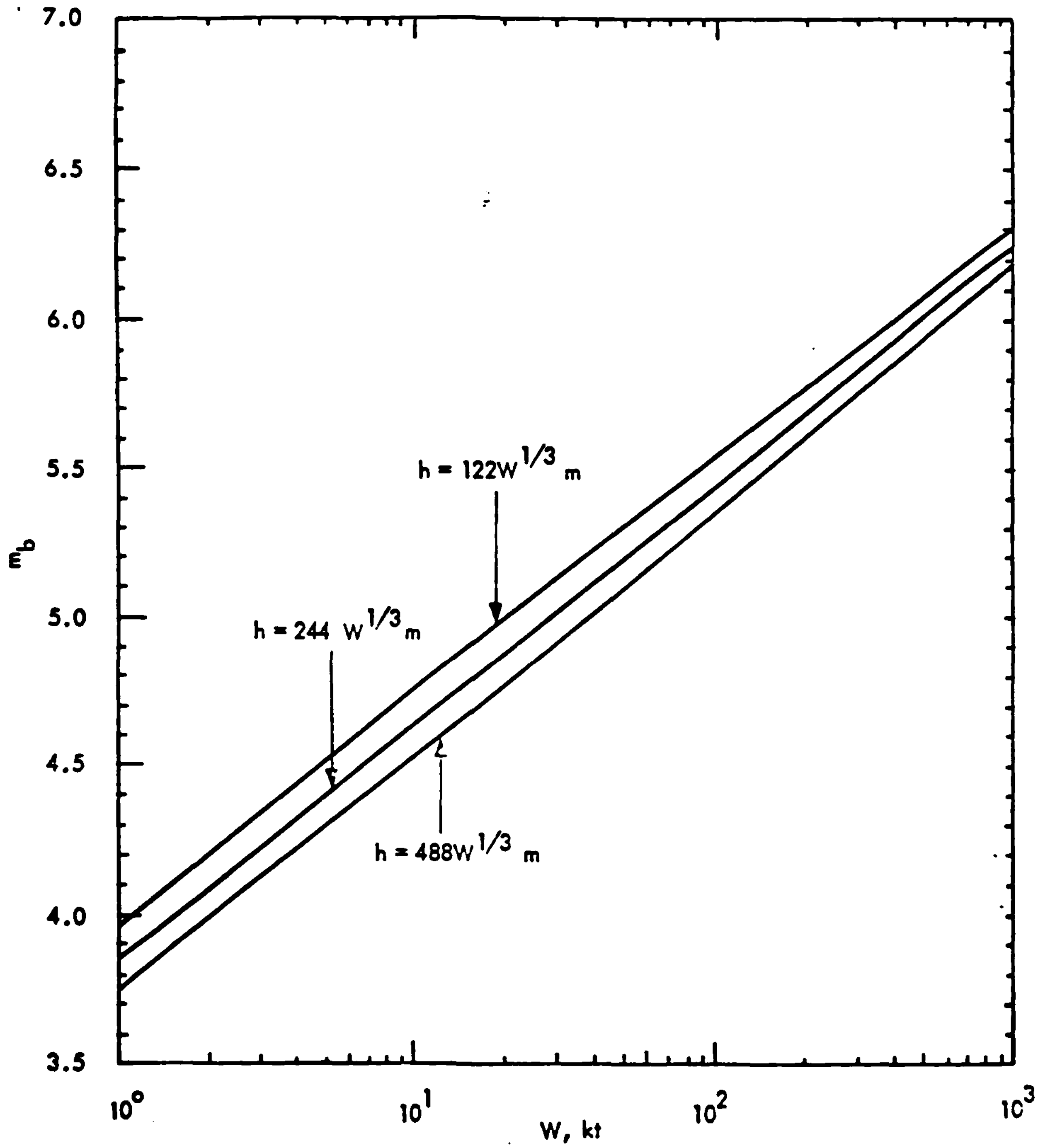

Figure 6. Example of the Effect of Source Depth of Burial on Theoretical Western U.S. m_b/Yield Curves, NTS Tuff/Rhyolite Emplacement Medium.

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TOPICS 4 AND 10

SOURCE COUPLING FOR BODY WAVES AND SURFACE WAVES

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Introduction

In this summary we present the results of a deterministic technique that predicts the seismic coupling for both body waves and surface waves. The basic element of the technique is a computer model of near field, nonlinear stress wave propagation which calculates the ground motion at arbitrary distances from the explosive source. For seismic coupling predictions the distances at which the calculation is monitored are always chosen to be outside the nonlinear region. This elastic ground motion forms the basis for estimating seismic coupling. 16

Within the computer model are descriptions (constitutive relations) of the response of the rock environment to stresses varying from a few megabars in pressure down to the elastic level. The most critical constitutive relations affecting seismic coupling are those involving irreversible pore collapse, tension failure and effective stress.

The next section of this summary presents the normalization of these constitutive relations involving comparisons between calculated and observed ground motion from explosive sources. In the remaining sections we present calculations of seismic coupling as a function of rock type and depth of burial.

Model Normalization

For estimating seismic coupling, the monitored stations must be in the elastic region. Therefore, the model must

evenutally be able to accurately propagate an elastic wave. Figures 1 and 2 (Cherry, et al., 1973) compare the analytic and computer model solutions to an elastic disturbance generated by an exponentially decaying pressure load applied to the inside of a 10 m cavity. The model is capable of accurately simulating the propagation of a small displacement elastic disturbance.

A constitutive model for irreversible pore collapse was presented by Cherry, et al. (1973). The pressure loading and release states obtained from this model for a partially saturated tuff are shown in Figure 3. Riney, et al. (1973) used this model to predict the ground motion from the Mine Dust HE shot, a 1,000 pound nitromethane explosion detonated May 10, 1972 at NTS Area 16. Figure 4 shows a comparison between the computer model, run on May 8, 1972, and the particle velocity recorded from the shot. The agreement shown in this figure is typical of comparisons at other distances. These results provided us with a great deal of confidence that realistic ground motion predictions can be made in weak rocks where the dominant mechanism for stress wave attenuation is the removal of air filled porosity.

The basic features of the tension failure model were presented by Cherry, et al. (1975) and used by Rimer, et al. (1979) to match the surface spall and slap down phases from the Piledriver event. Figure 5 compares calculated and observed particle velocities 368 m from SGZ for this event. This comparison indicates that the tension failure model contains the physics necessary to model spall effects from a nuclear event. As a result we proceed with a two dimensional parameter study to determine the effects of yield and depth of burial on body waves and surface waves in NTS grandiorite. Preliminary results from this study will be presented later in this summary.

Two features are included in the material strength portions of the constitutive model that are not usually present in calculations performed by other investigators. These include the dependence of material strength on both the

third deviatoric stress invariant (Cherry and Petersen, 1970) and pore fluid pressure (Cherry, et al., 1975; Rimer, et al., 1979). These features have allowed the model to accurately calculate ground motion in saturated and partially saturated rocks having large values of material strength when tested dry in the laboratory. Figures 6 and 7 compare the observed and calculated ground motion from the Piledriver event at shot level. This type of agreement is only possible if the rock environment is assumed partially saturated and the effect of pore fluid pressure is included in the material strength portion of the model.

Near field ground motion measurements are sparse and often inadequately address the low frequency content of the ground motion responsible for body wave and surface wave coupling at teleseismic distances. Therefore, as an additional aid for model normalization, we conducted laboratory experiments to obtain high quality measurements of rock motion from an explosive source (Cherry, et al., 1977). The measurements were taken on the surface of specially prepared concrete cylinders and the source was 0.25 gm of PETN. Figure 8 compares the experimental data with two calculations, with and without tension failure. A characteristic of tension failure is a peaked RVP spectrum, caused by a discontinuity in tangential stress at the linear-nonlinear boundary. The calculation with tension failure in the material model is in better agreement with the data.

Finally, a definition of what we mean by "elastic behavior" is now appropriate. We define elastic behavior as the absence of irreversible pore collapse, tension failure and yielding. In addition, the stress-strain relation is obtained from single values of bulk modulus and shear modulus. We have not found it necessary to include rate dependent effects in the model. However, near field ground motion data in salt suggest that salt's material strength may be dependent on inelastic strain energy.

Parameter Study in One Dimension

Cherry, et al. (1975) conducted a one-dimensional parameter study to determine the dependence of teleseismic magnitudes on the nonlinear behavior of the near source rock environment. They calculated $\psi(\infty)$ for systematic changes in material properties and computed the corresponding change in magnitude (Δm), where

$$\Delta m = m^i - m^k = \log\left[\frac{\alpha_i \ \psi_i(\infty)}{\alpha_k \ \psi_k(\infty)}\right] .$$

Figures 9, 10 and 11 show the effect of air filled porosity, maximum material strength and overburden pressure on Δm. These were shown to be the most sensitive parameters in the model.

Seismic Coupling at NTS

Bache, et al. (1975) used this computer model to explain the relative differences in body wave coupling between various testing areas at NTS. The observed teleseismic data is shown in Figure 12.

Material properties used for the equivalent elastic source calculations were obtained from both laboratory tests on appropriate rock samples and CEP reports. The RVP spectra computed for each testing area is shown in Figure 13. The reasons for the differences in these sources are as follows:

1. The ratio of the spectral peak to the zero frequency limit ($\psi(\infty)$) increases with increasing material strength. Therefore, both Piledriver and Pahute Mesa rhyolite, having the longest values of material strength, show highly peaked spectra compared to the other three areas.

2. Both the Piledriver and rhyolite calculations used the same material strength. Piledriver couples approximately twice as well due to lower overburden pressures.

3. The Area 12 material couples three to five times better than Yucca Flat tuff and Pahute Mesa rhyolite for frequencies up to ten Hz.

4. The Area 12 tuff couples higher than Yucca Flat wet tuff due to lower air voids and overburden pressure.

5. The Yucca Flat dry tuff couples low due to the high air filled voids and high strength assumed for the site.

6. Pahute Mesa rhyolite couples lower than Area 12 tuff due to high strength and high overburden pressure.

These equivalent sources were propagated to teleseismic distances. Synthetic seismograms were computed and compared to the data. As shown in Figure 14, the calculations match the data quite well indicating that the one-dimensional source calculations are accounting for the robust features controlling body wave coupling at NTS.

A similar analysis has been performed by Bache, et al. (1978) for surface waves using data from stations at Tucson and Albuquerque. They concluded that the $\psi(\infty)$ values obtained from the source calculations are within acceptable limits to those inferred from the data.

Parameter Study in Two Dimensions

Rimer, et al. (1979) conducted a two-dimensional parameter study to determine the effect of yield and depth of burial on surface wave and body wave magnitudes. The rock environment was NTS fractured granodiorite. Near field data from the PILEDRIVER event (61 KT, 460 m) was compared to the results of the PILEDRIVER calculation (Figures 5, 15 and 16). The conclusion was that the agreement was good enough to warrant a systematic investigation of the degradation of pP by spall and the resulting effect on seismic coupling. The yields (W) and depth of burial (DoB) comprising the study are given in the following table.

W (KT)	DoB (m)	Scaled DoB (m(KT)$^{1/3}$)
20	400	147
20	1000	368
61	460	117
150	1000	188

Each calculation was monitored on a cylindrical surface in the elastic regime. These results were then analytically continued to the far field in order to obtain estimates of m_b and M_s.

The surface wave calculations (Figure 17) show 0.3 magnitude units difference between the one- and two-dimensional PILEDRIVER calculations and 0.6 magnitude units difference between the shallow and deep two-dimensional calculations of the 20 KT sources. Agreement between the one- and two-dimensional calculations is good for the two deep shots.

The body wave magnitudes are shown in Figure 18. the variation of these magnitudes with yield and depth of burial is much reduced from that found for surface waves. It appears

that the "b" phase is adequately modeled by a one-dimension simulation for the depths and yields considered here.

These results are preliminary in that the analytic continuation procedures are still being tested and the physical basis for the variations shown here has not been fully addressed. However, it does appear that two-dimensional simulations can improve our understanding of the effect of "spall" on seismic coupling and hopefully permit a more detailed match of the short period seismogram.

Seismic Coupling in Salt

Realistic decoupling scenarios in salt can be developed only after apprcpriate free field ground motion data in this rock is understood. The SALMON event (Perret, 1968) provided a large subset of this data. These data exhibit a number of puzzling features, the most important being a small amplitude "elastic" precursor which is not consistent with laboratory strength measurements or the overburden pressure at shot depth.

Here we present a possible explanation of this precursor which depends on the SALMON shot environment being saturated. If we assume saturation, then the precursor emerges. Its amplitude depends on the assumed saturated strength.

Figure 19 shows a comparison between the calculated and observed radial ground motion at a distance of 278 m from the SALMON event. The peaks have been aligned by shifting the time axis. In the calculation we assumed that the salt was totally saturated.

In the data the "elastic" precursor has a peak velocity of approximately 0.4 m/sec while the precursor in the calculation peaks at 1.2 m/sec. Therefore, the assumed saturated strength for salt was too high by about a factor of three. We should note that there is no strength data for saturated salt.

Our assumed strength was 77 bars, obtained from an extrapolation of triaxial compression data to zero mean stress. There is no reason to expect this extrapolation accurately represent the strength of salt at low stress states.

In addition the width of the calculated velocity pulse is about a factor of two broader than the data. Therefore, the assumed saturated strength was low by at least a factor of two during that portion of the velocity pulse which follows the precursor.

This conflict between the material strengths associated with the precursor and that following the precursor can be resolved if salt is assumed to work harden after the saturated strength is attained. The physical explanation for the work hardening may be an increase in the effective stress, and conversely a decrease in pore fluid pressure, caused by dilatancy.

It is interesting that salt apparently requires a constitutive model different from those used for the rocks at NTS. They all have one feature in common however, namely that seismic coupling is controlled by low strength states, i.e., those that are between the tensile strength and the unconfined compressive strength. This reduction in strength has been attributed to pore fluid pressure and effective stress. Therefore, the degree of saturation at shot depth and the location of the water table are critical seismic coupling site properties. In addition, it is important that laboratory strength data be obtained for critical rock types at stress states below unconfined compression.

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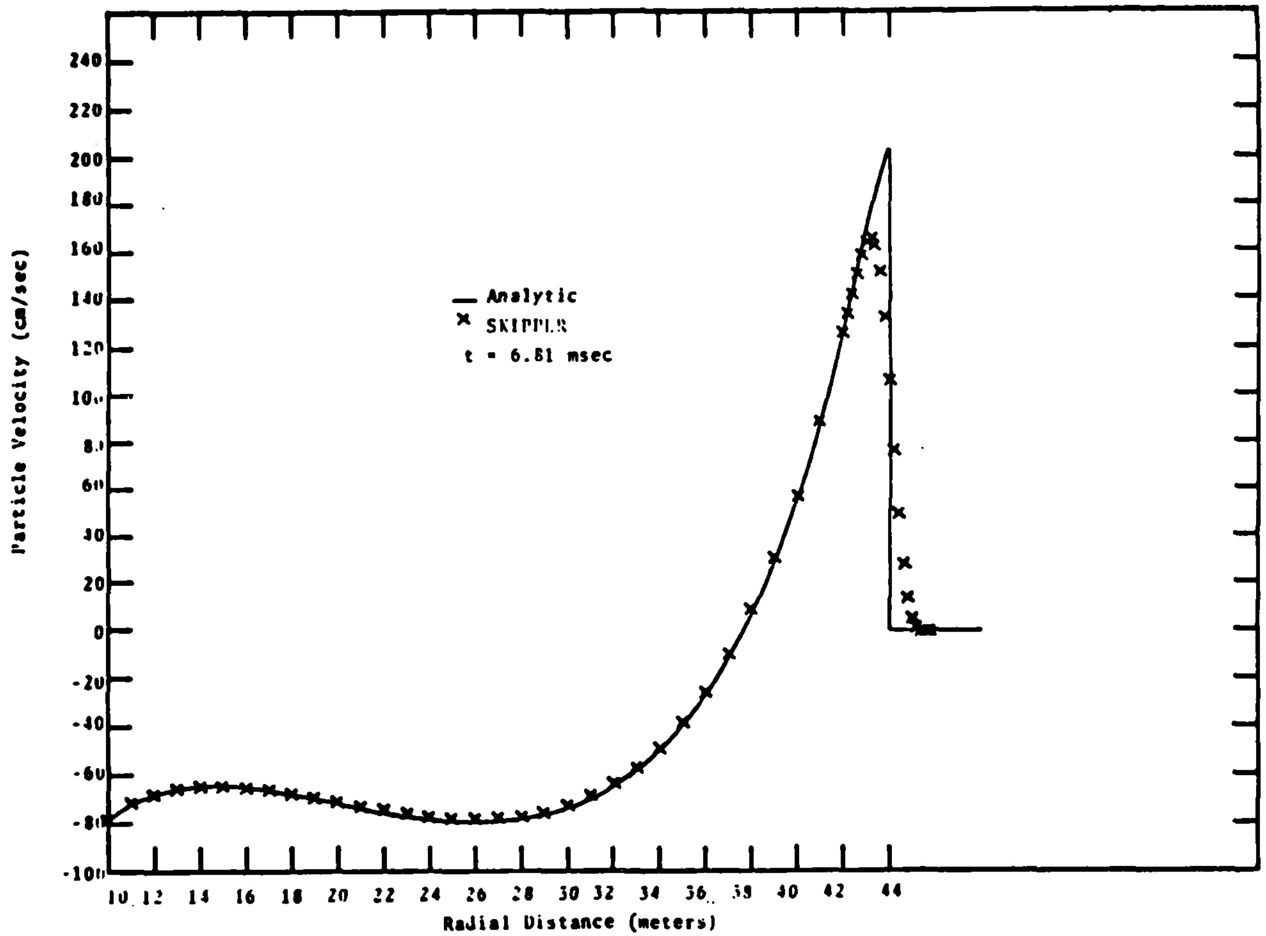


Figure 1. SKIPPER – Analytic comparison.

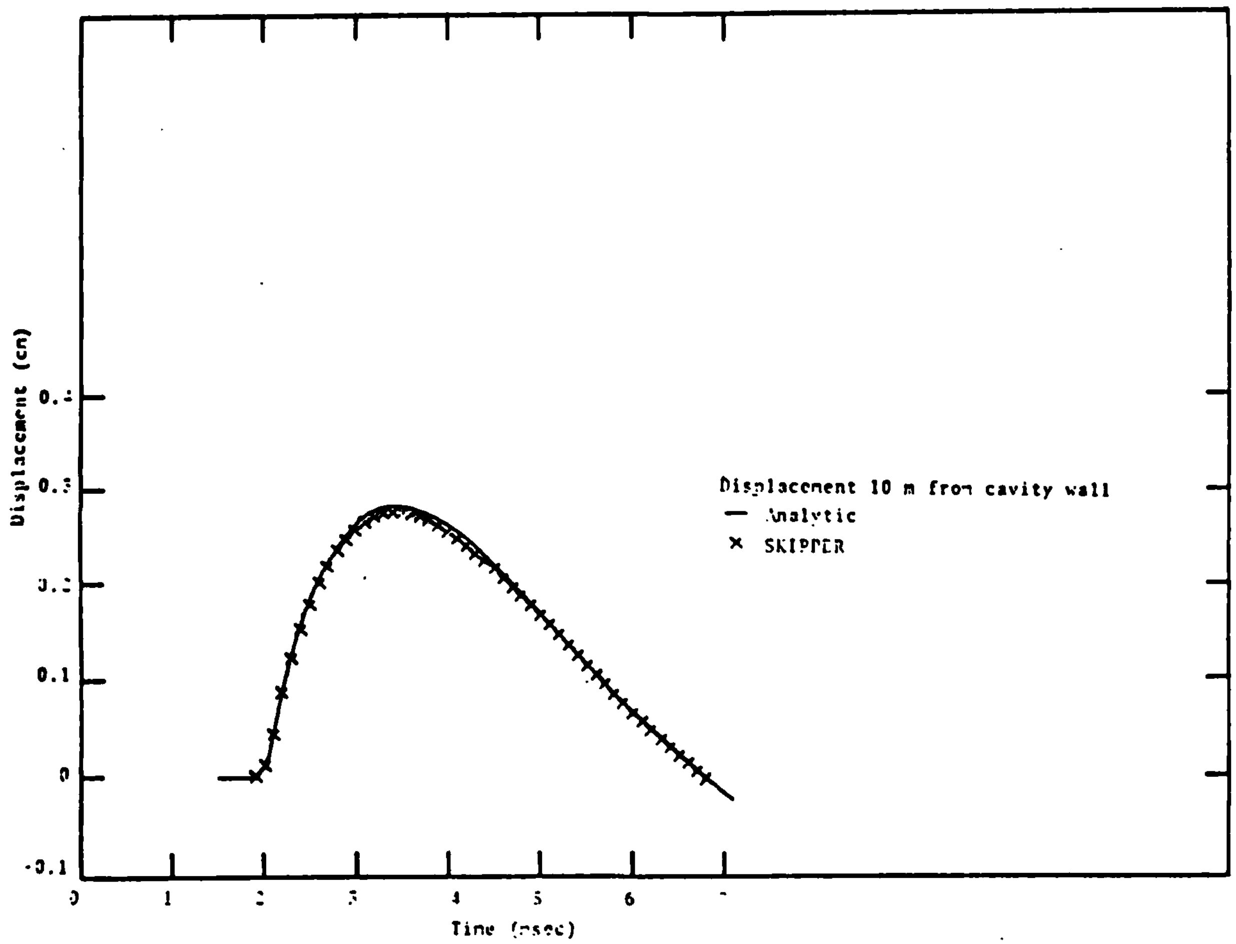


Figure 2. SKIPPER — Analytic comparison.

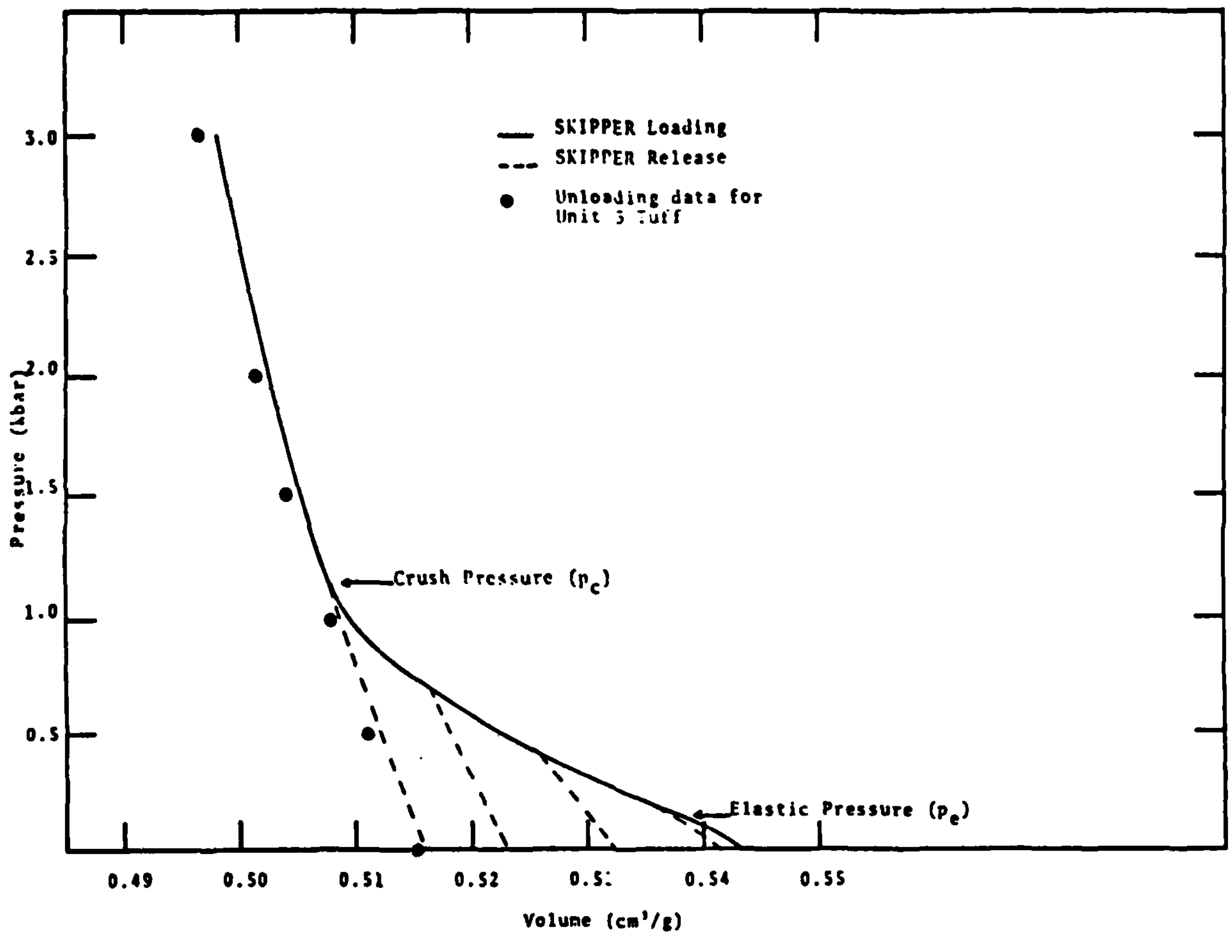


Figure 3. Loading and release P-V curves for partially saturated tuff ($f = 0.17$, $\phi_0 = 0.05$).

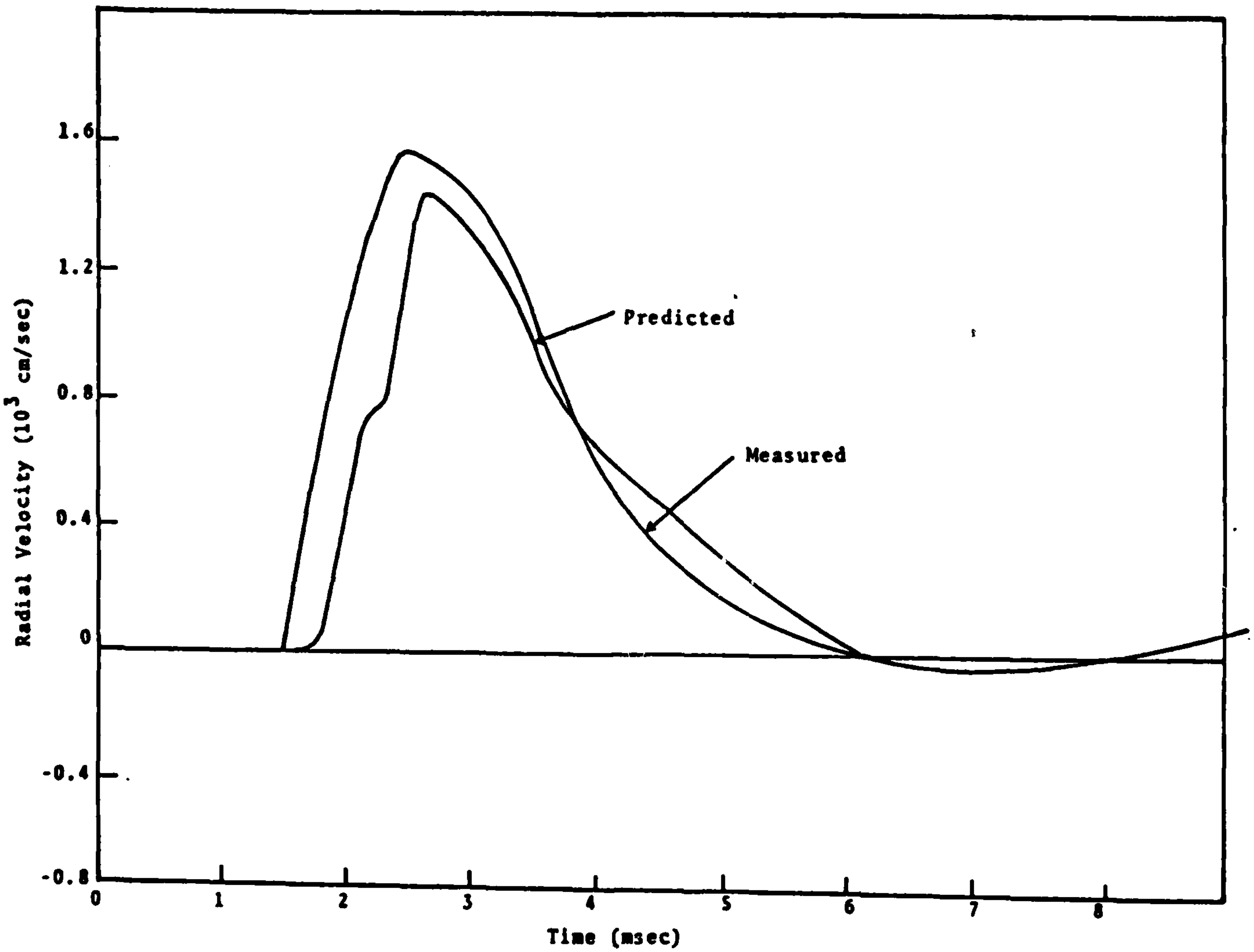


Figure 4. Predicted (May 8 Run) and measured (ATI) radial velocity histories at R = 11.95 feet and 12.09 feet respectively.

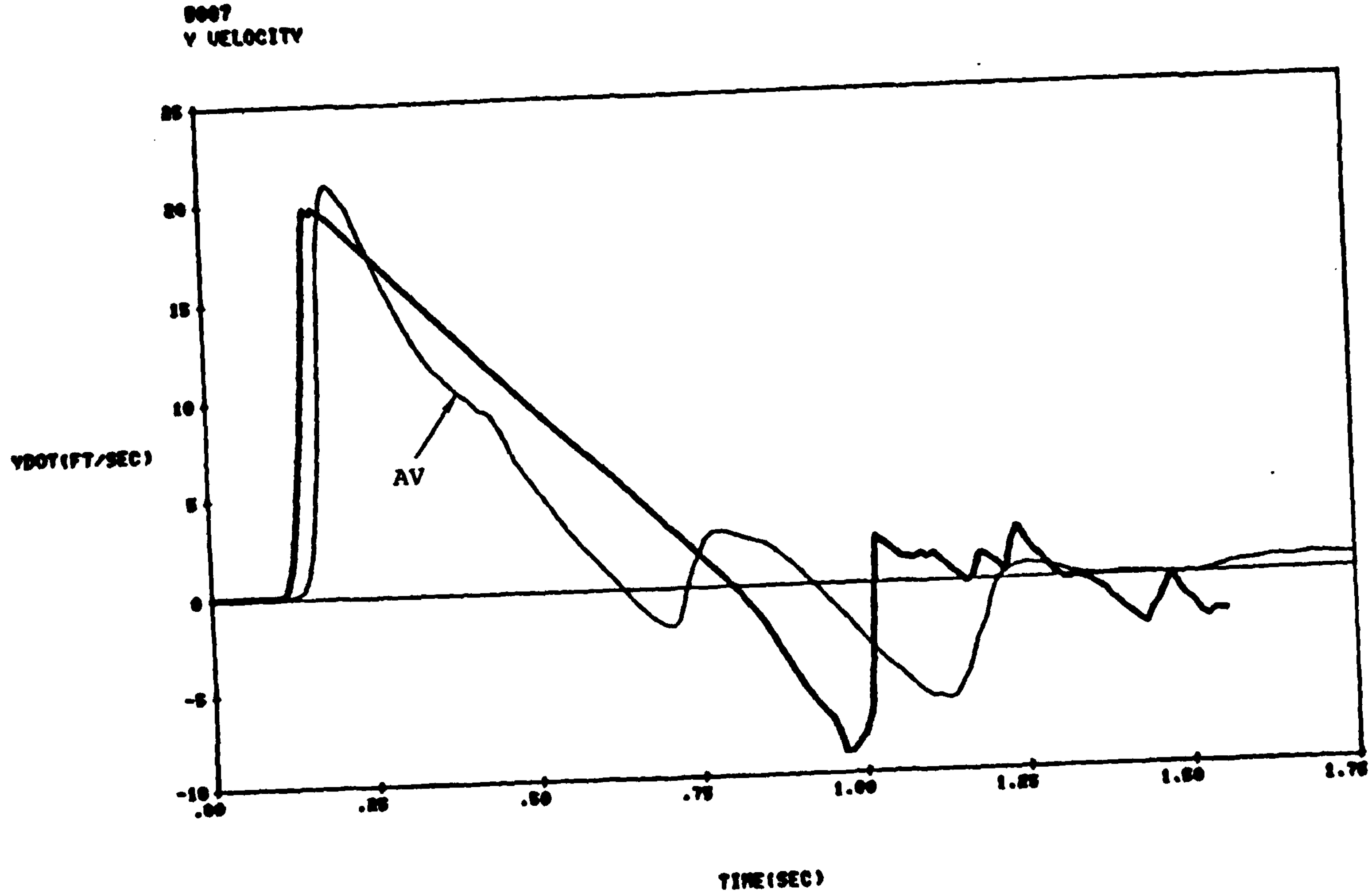


Figure 5. Measured and calculated vertical velocities at free surface station 9007 (horizontal range = 368 m).

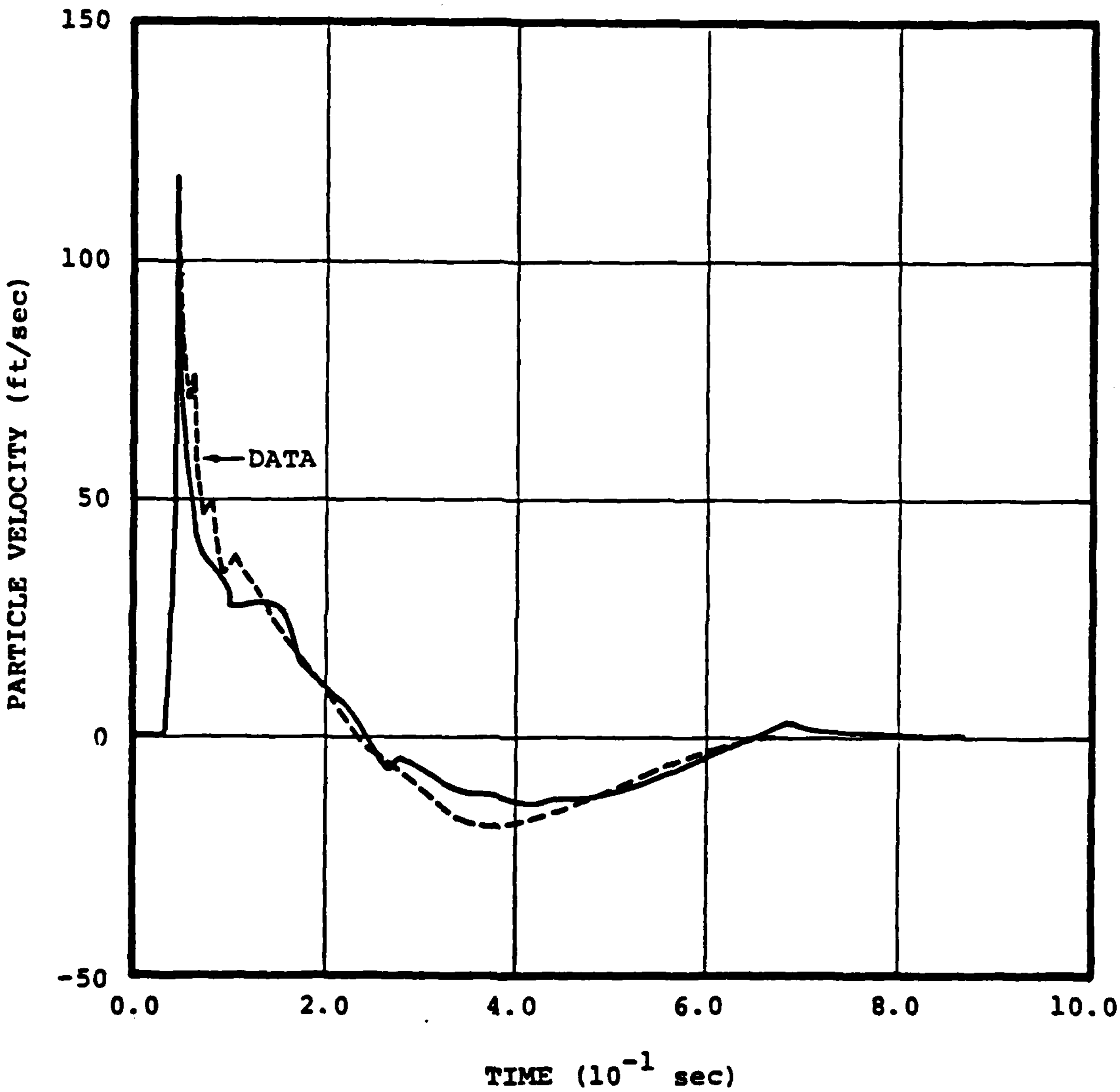


Figure 6. Comparison between velocity gauge data and one-dimensional calculation at Perret shot level station B-SL, X = 204 m, Y = 0.

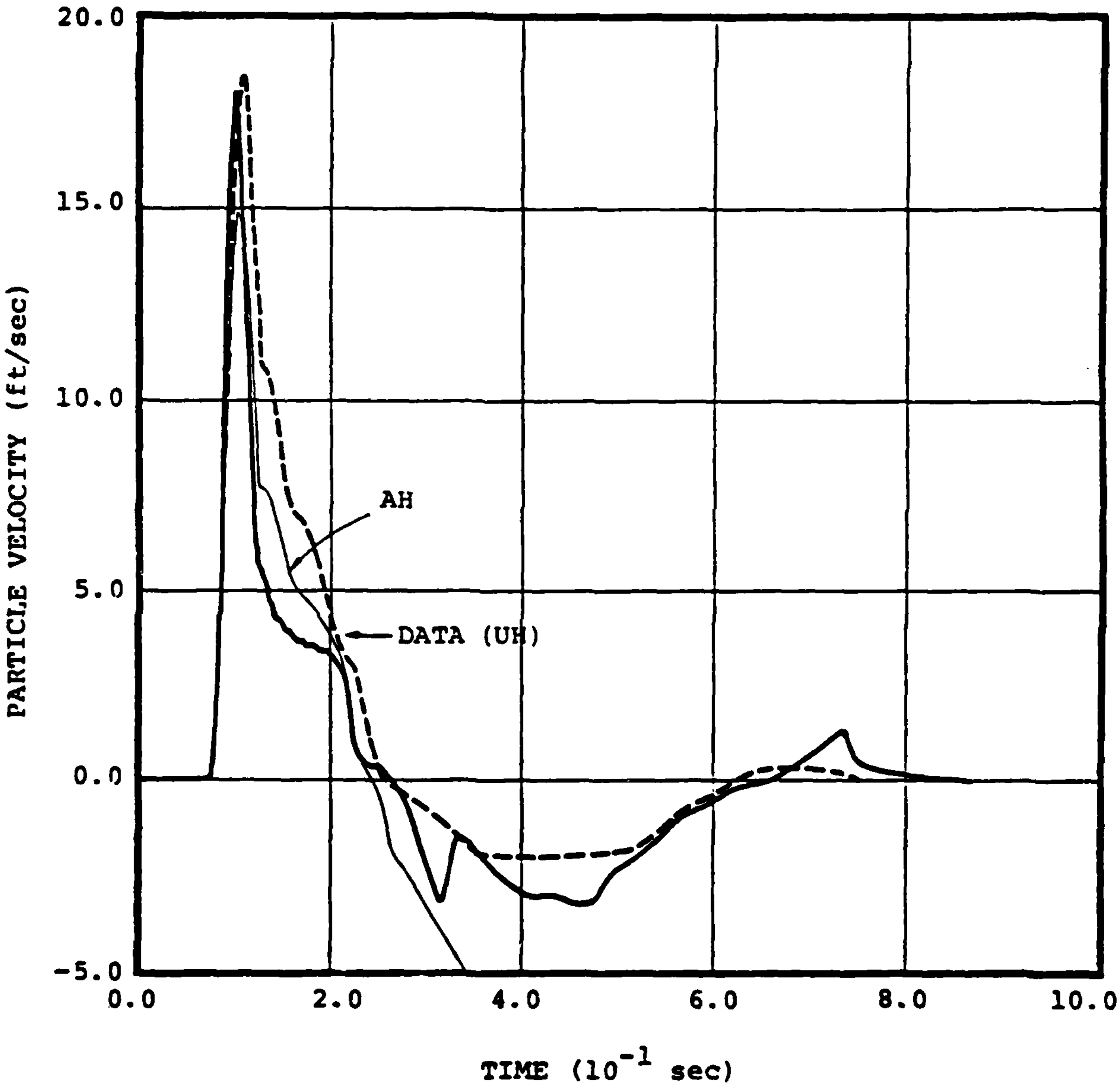
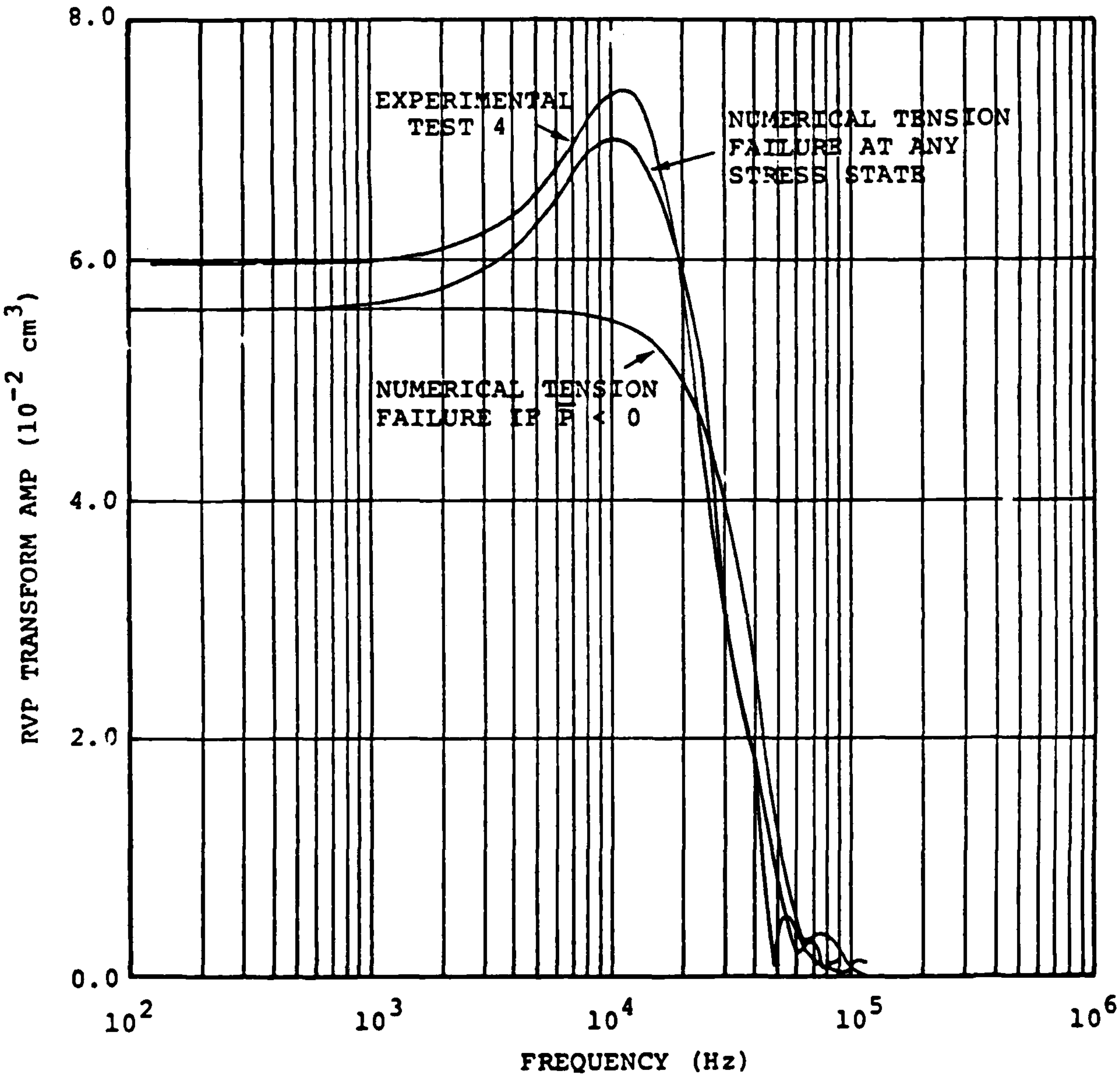


Figure 7. Comparison between velocity gauge data and one-dimensional calculation at Perret shot level station 16-SL, X = 470 m, Y = 0.



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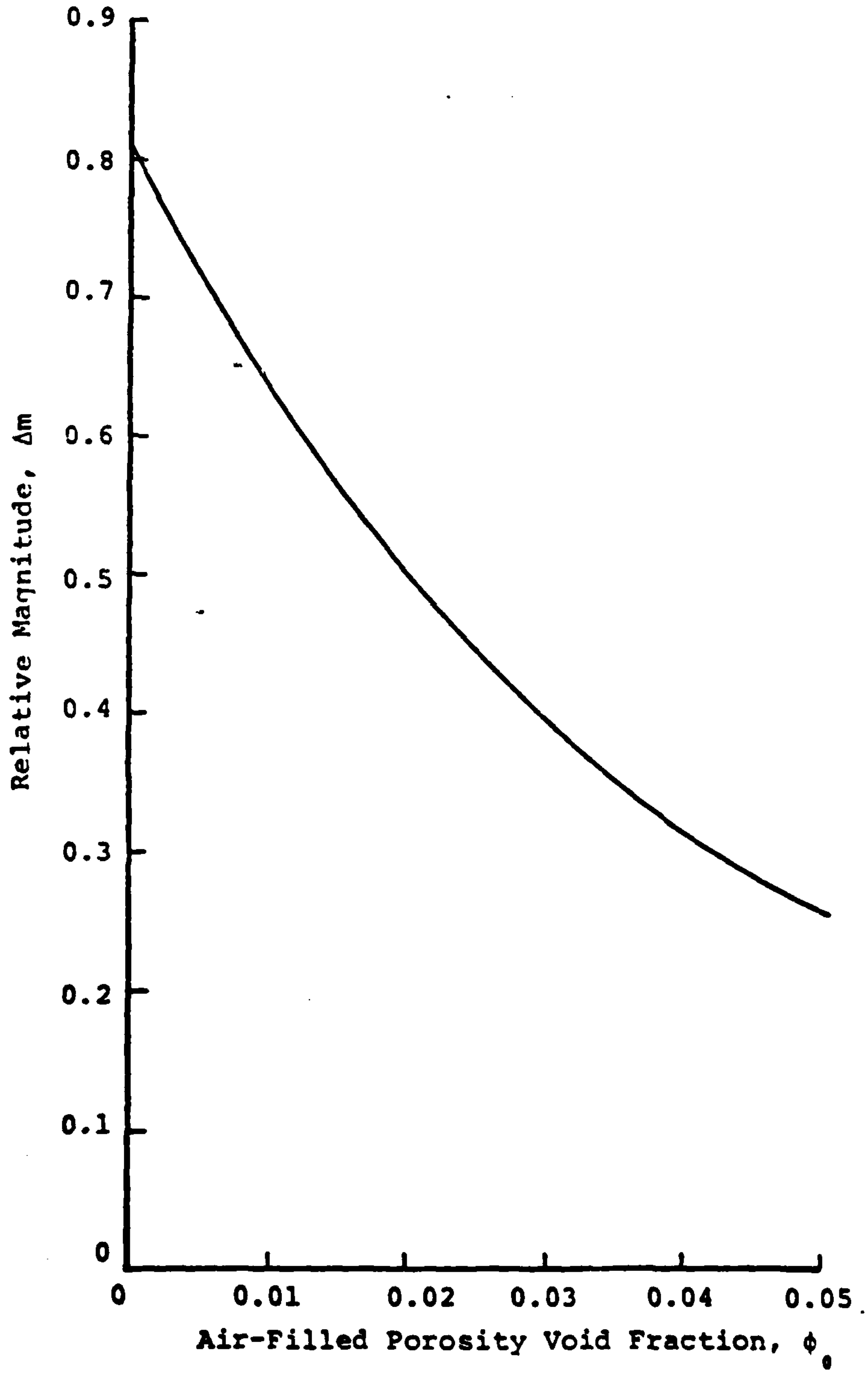

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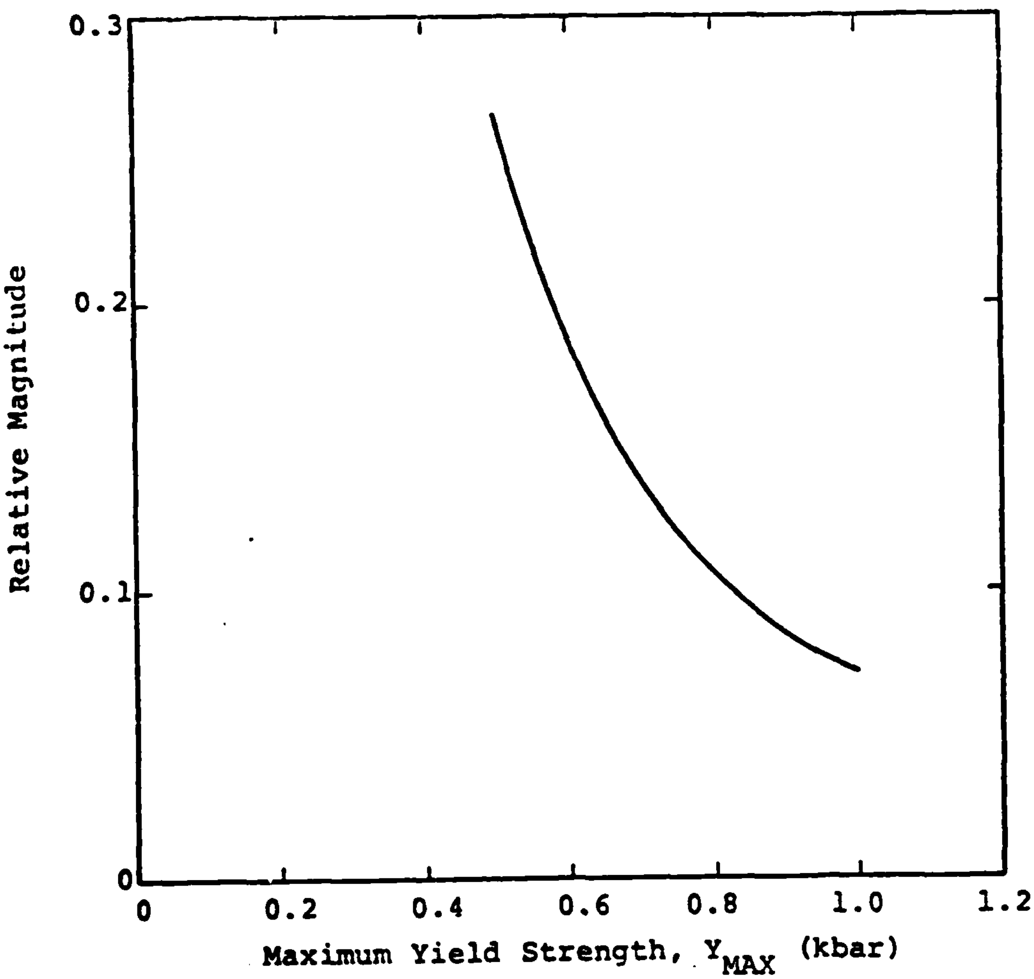
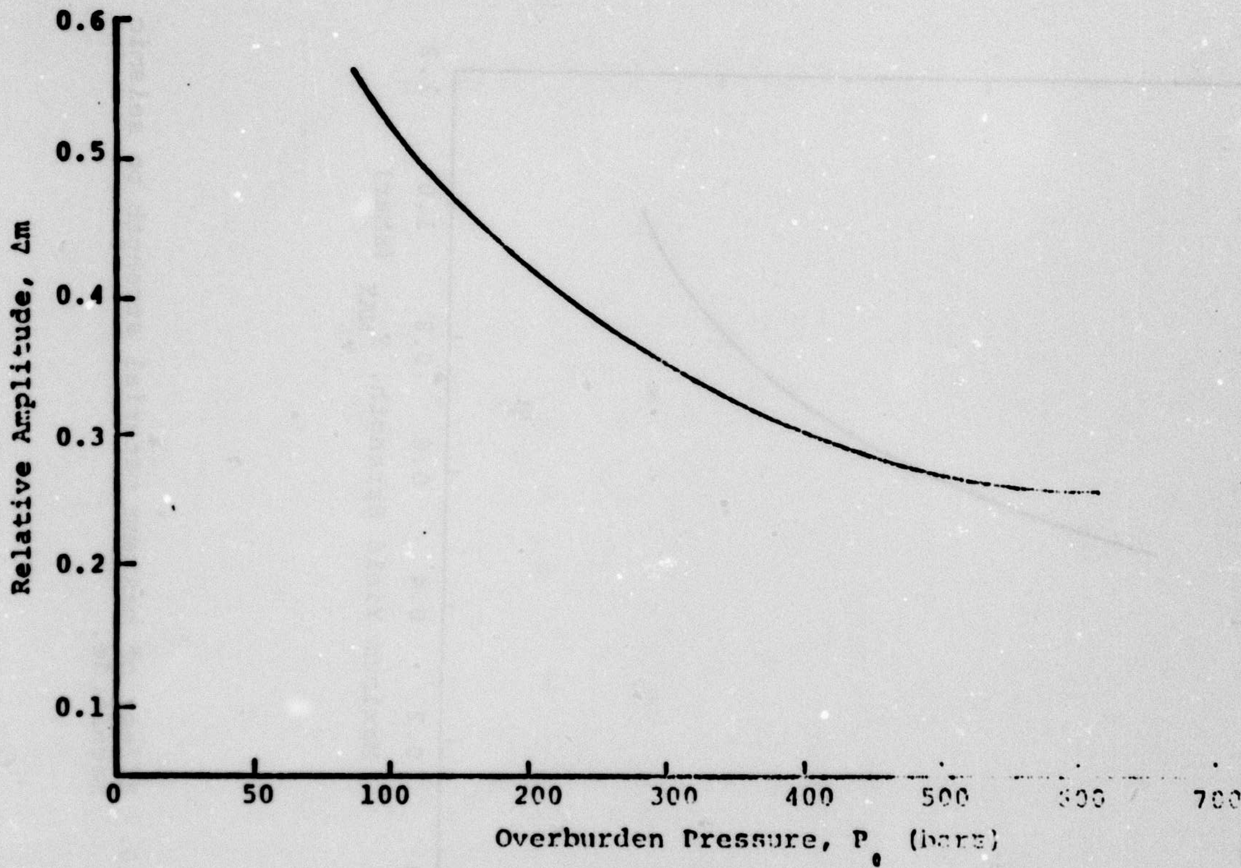


Figure 10. Effect of maximum material strength on seismic magnitude.



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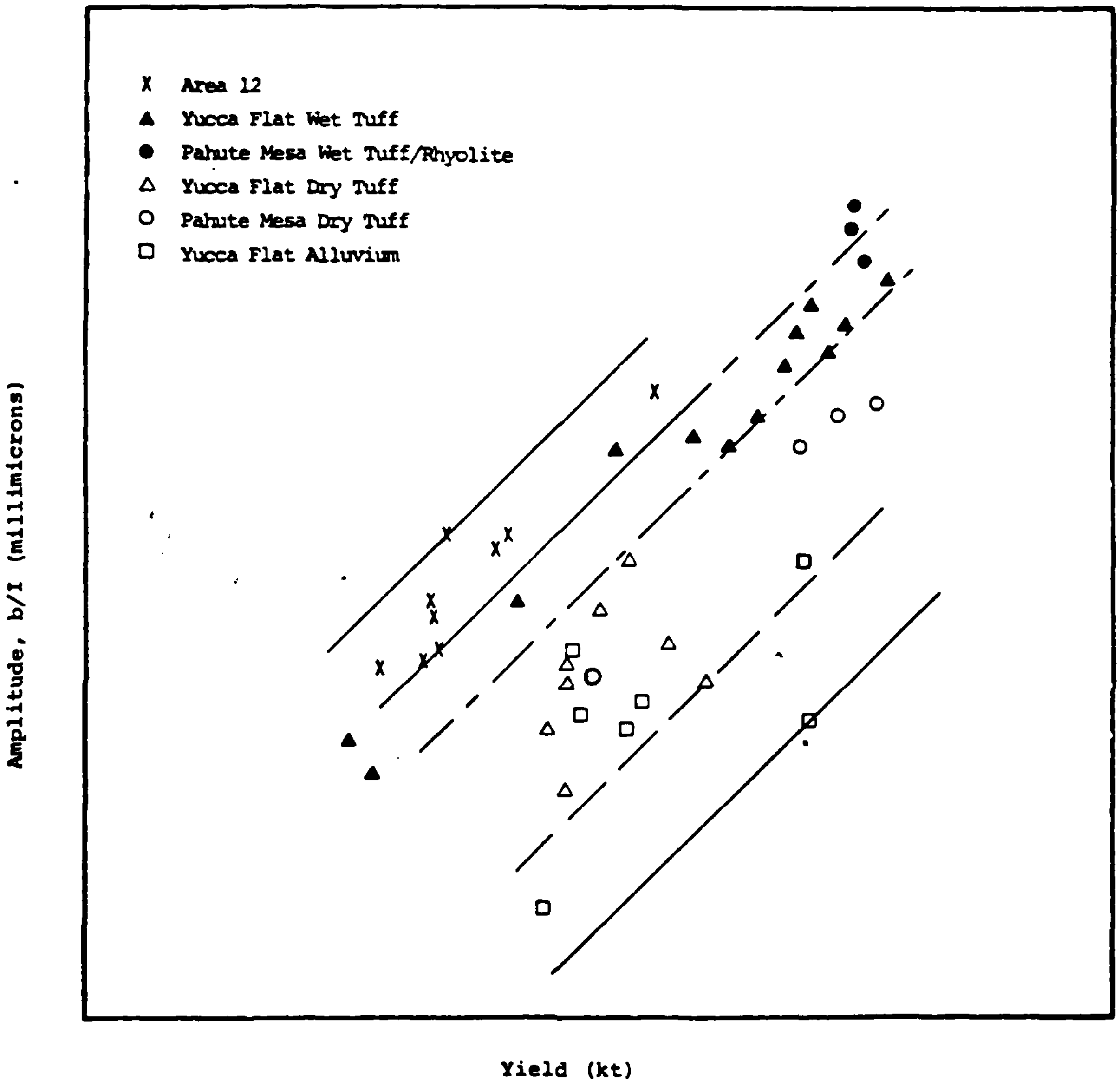

Figure 12. Amplitude of the b phase, corrected for instrument response, as recorded at a single seismograph station in the teleseismic field. The plot is log-log with three cycles on each axis. The recorded events are separated into groups of superficially common source material characteristics as indicated in the legend. The lines are of unit slope.

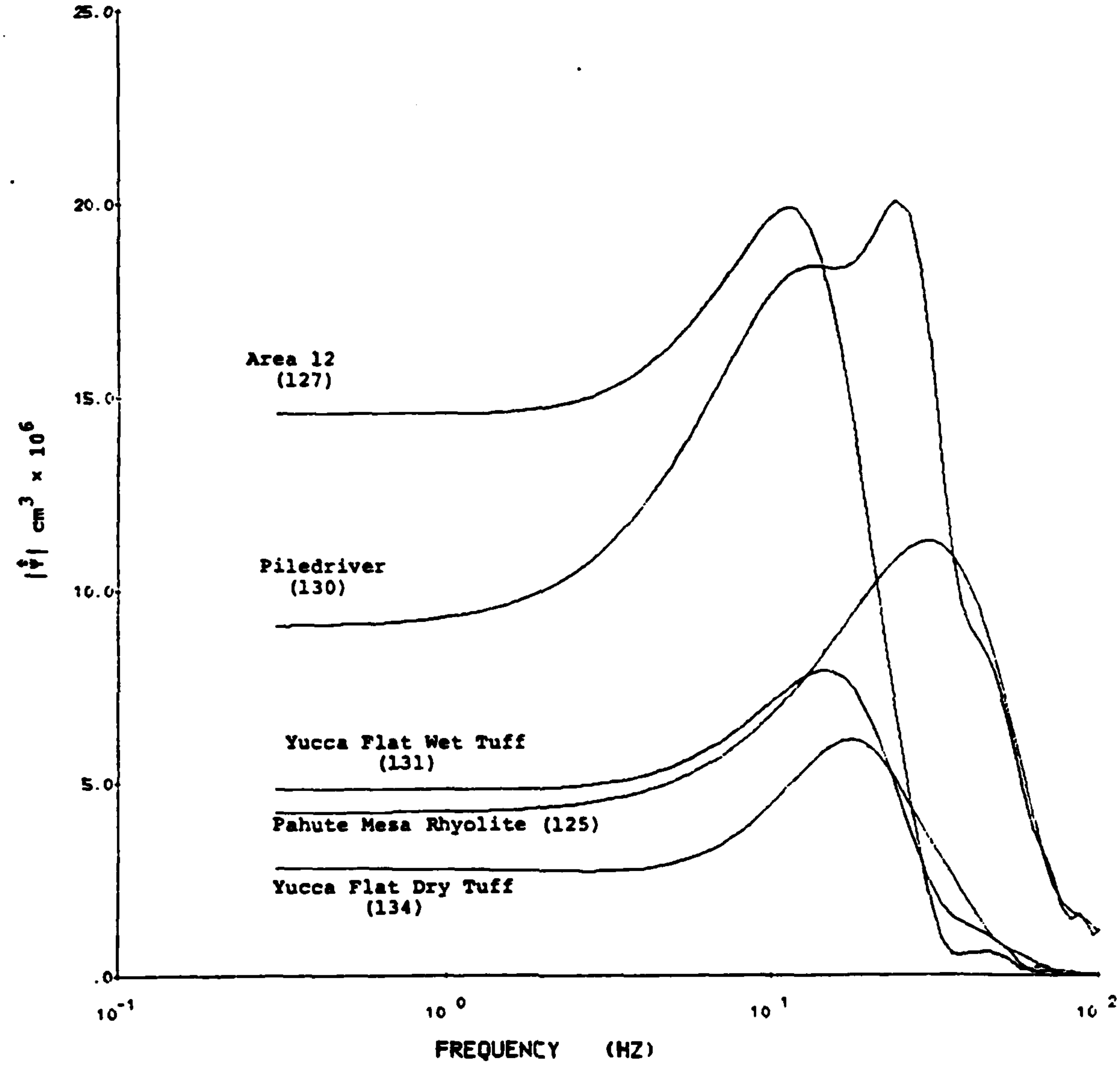

Figure 13. Source spectra for the indicated test areas which best match the teleseismic data. The device yield for each calculation was 0.02 KT.

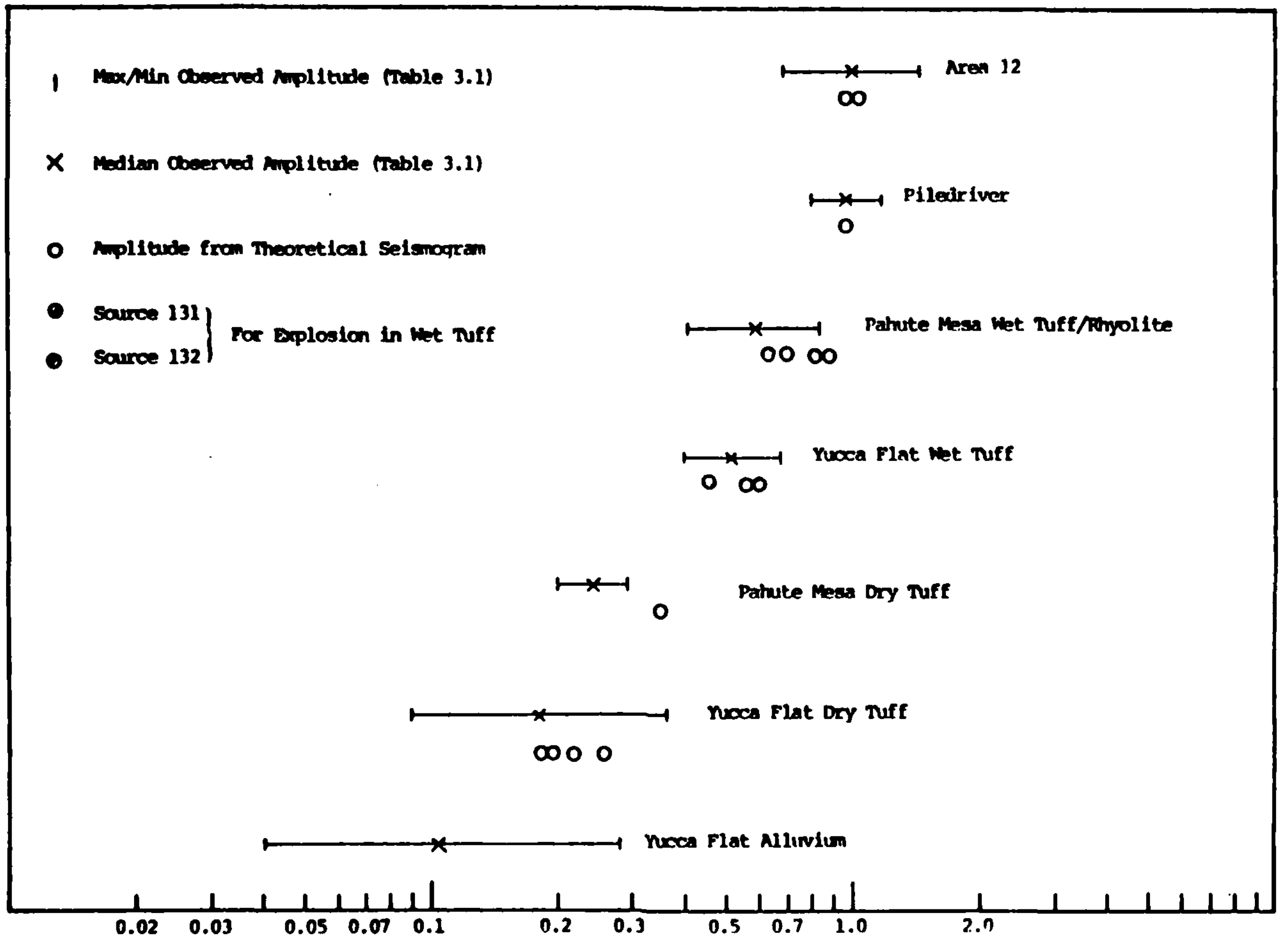

Figure 14. Comparison of theoretical and observed relative coupling between NTS events. The observed data is directly from Table 2.1. The theoretical amplitudes, $|A_b|$, have been normalized to the median for the theoretical Area 12 events.

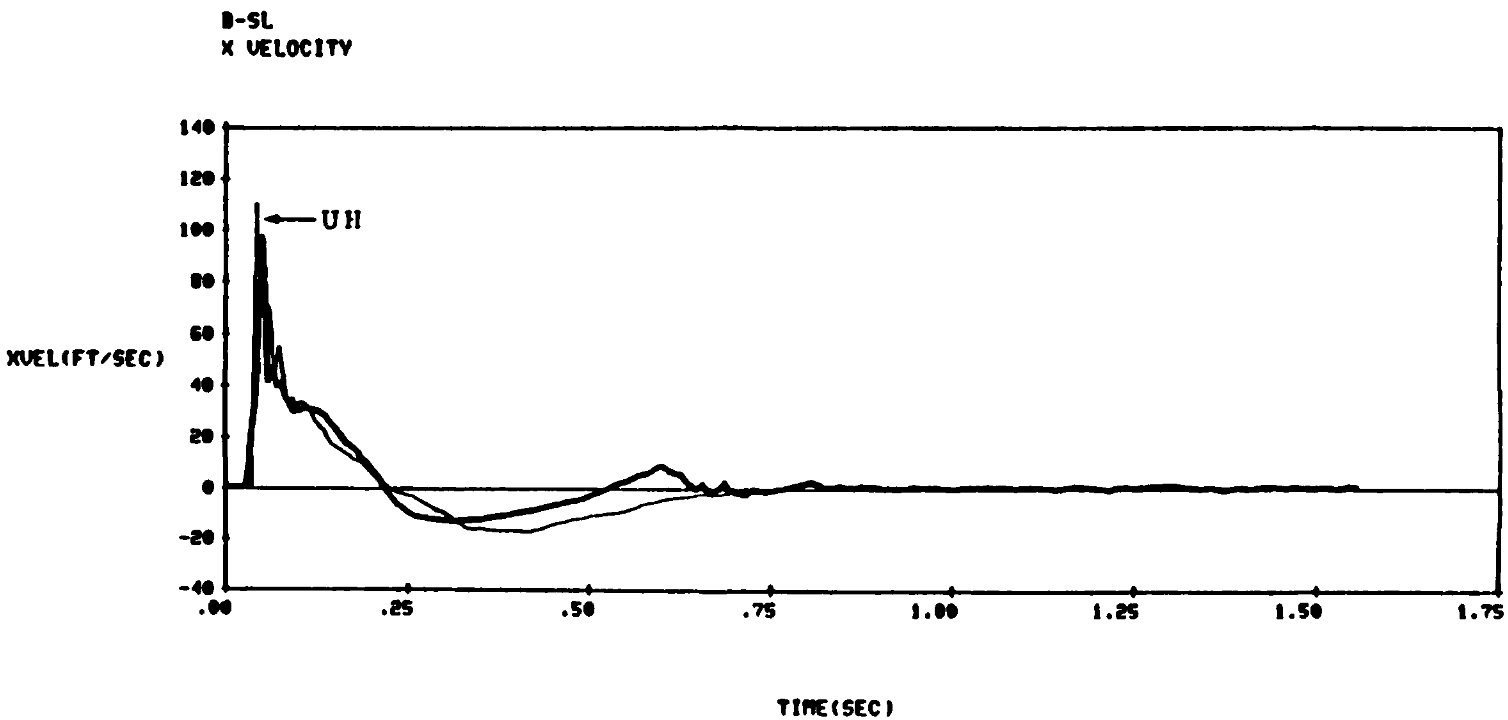

Figure 15. Measured and calculated horizontal velocities at shot level Perret station B-SL (range = 204 m).

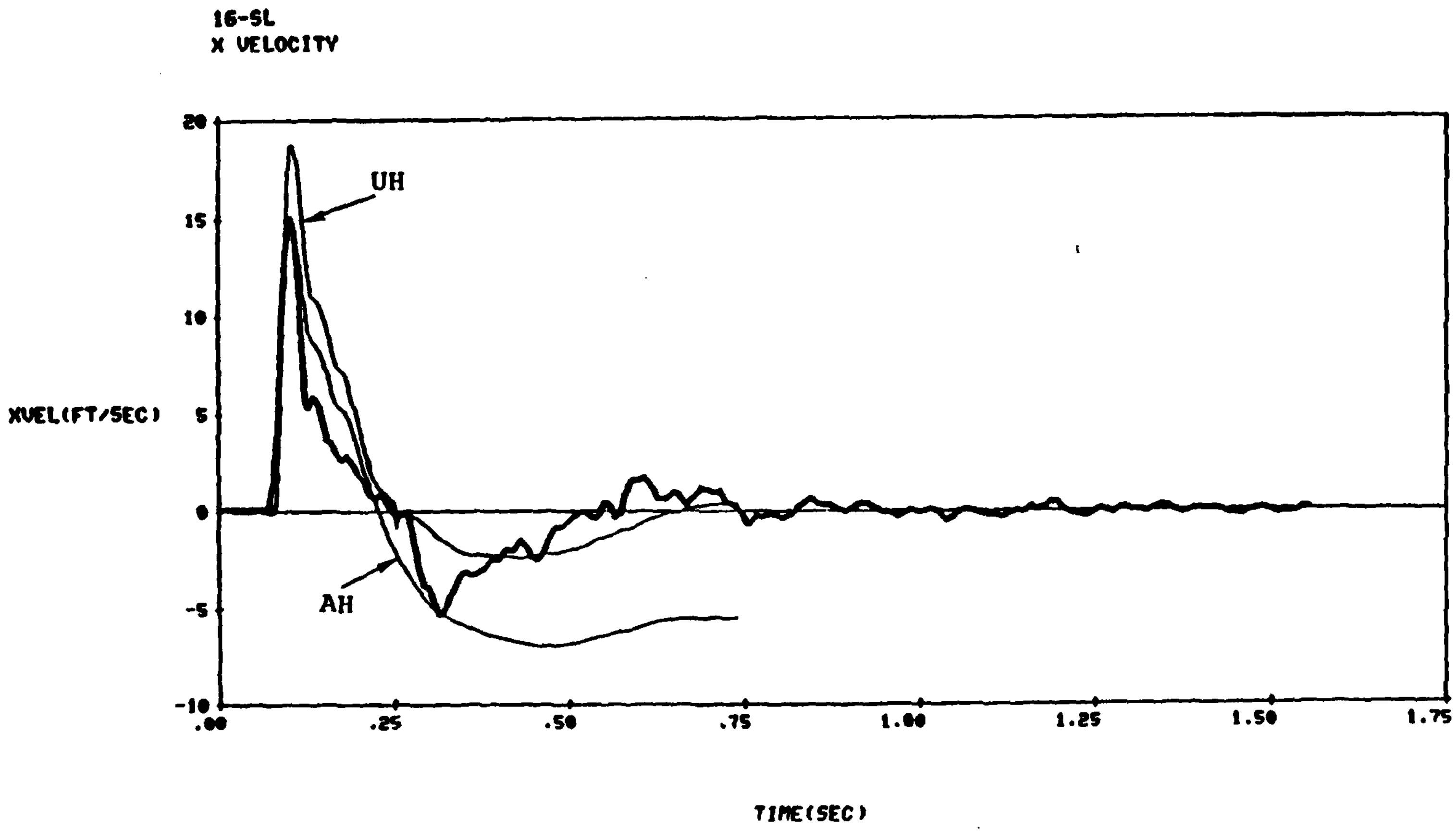

Figure 16. Measured and calculated horizontal velocites at shot level Perret station 16-SL (range = 470 m).

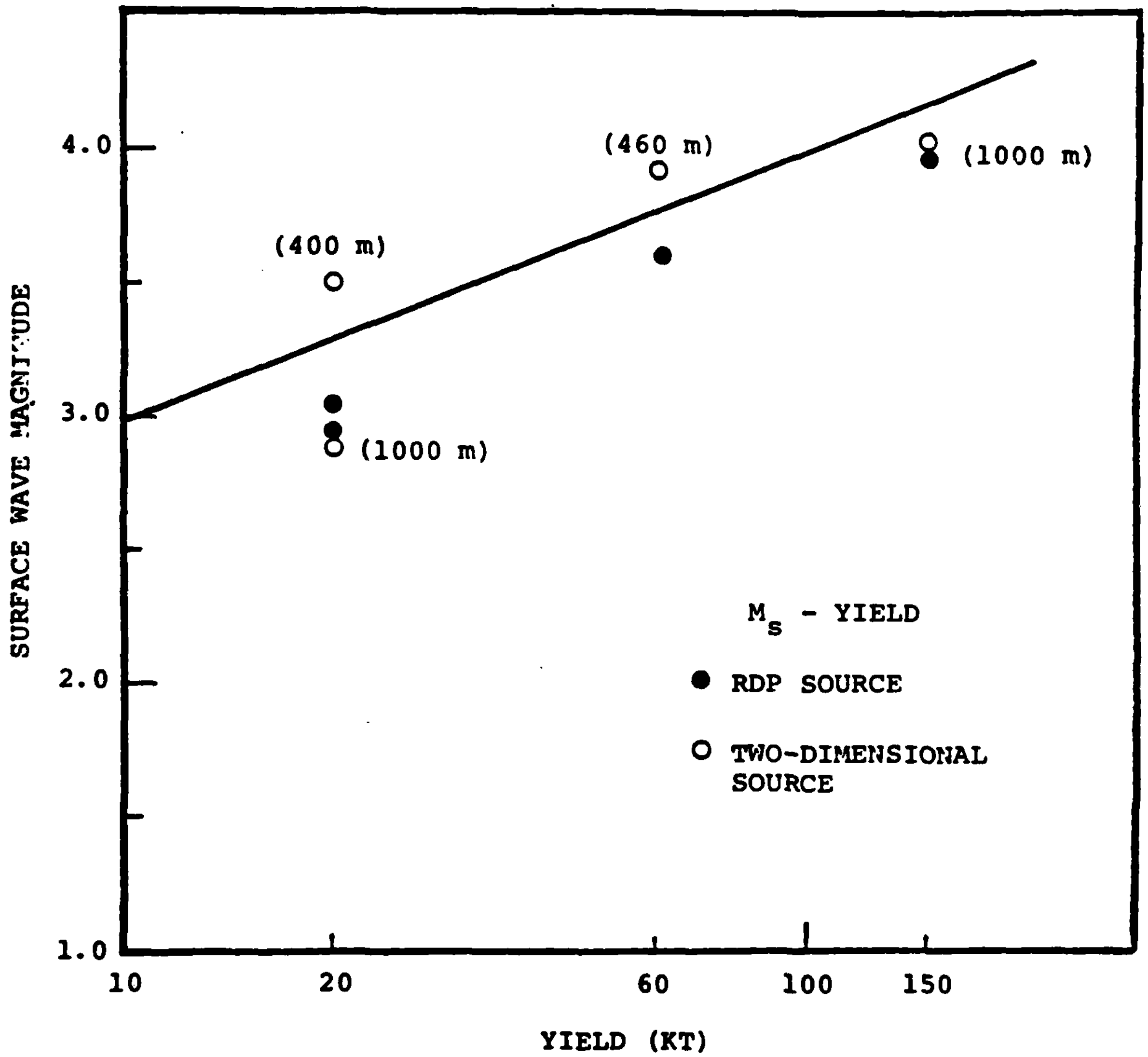

Figure 17. M_s versus yield for NTS granodiorite.

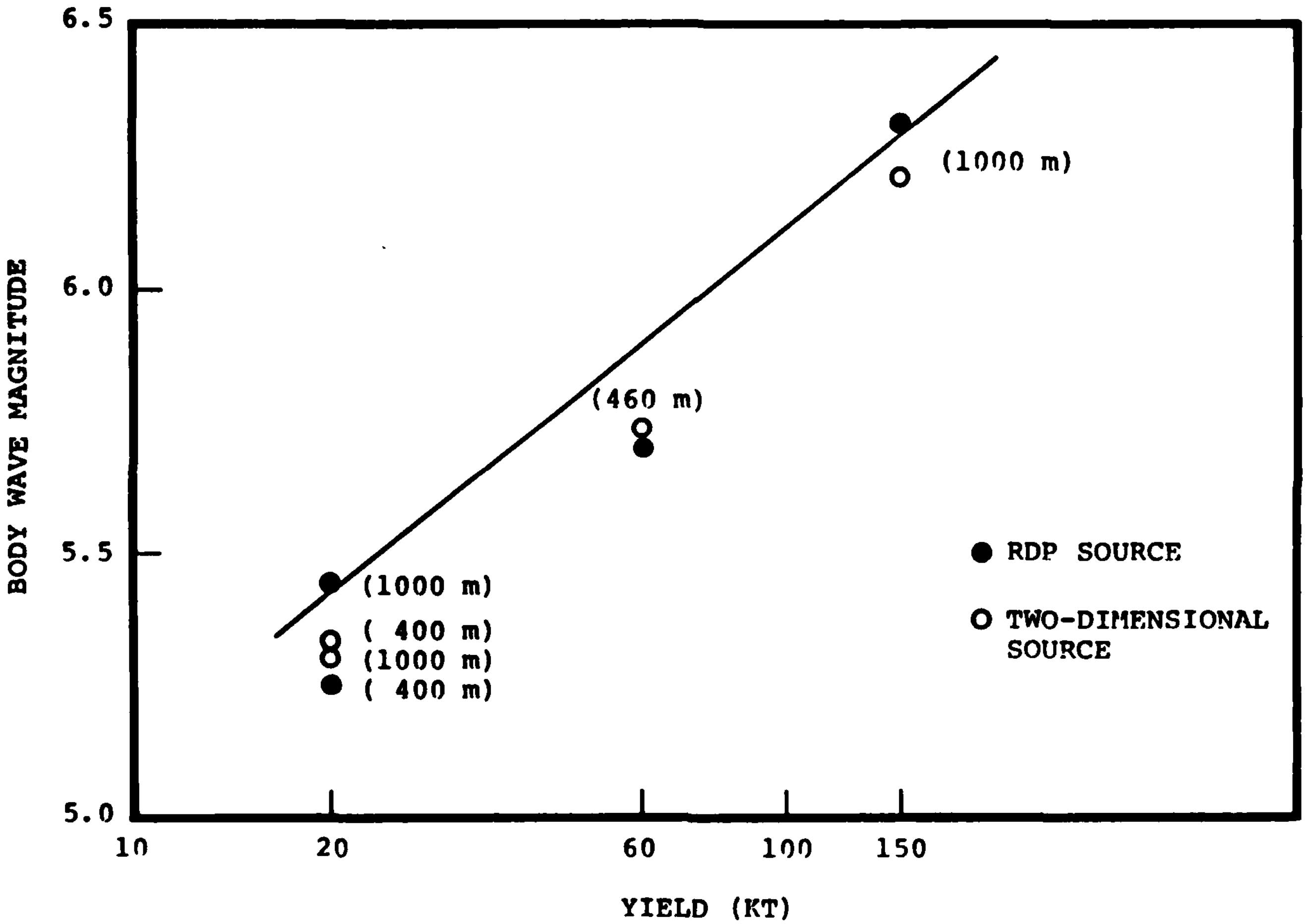

Figure 18. m_b versus yield for NTS granodiorite.

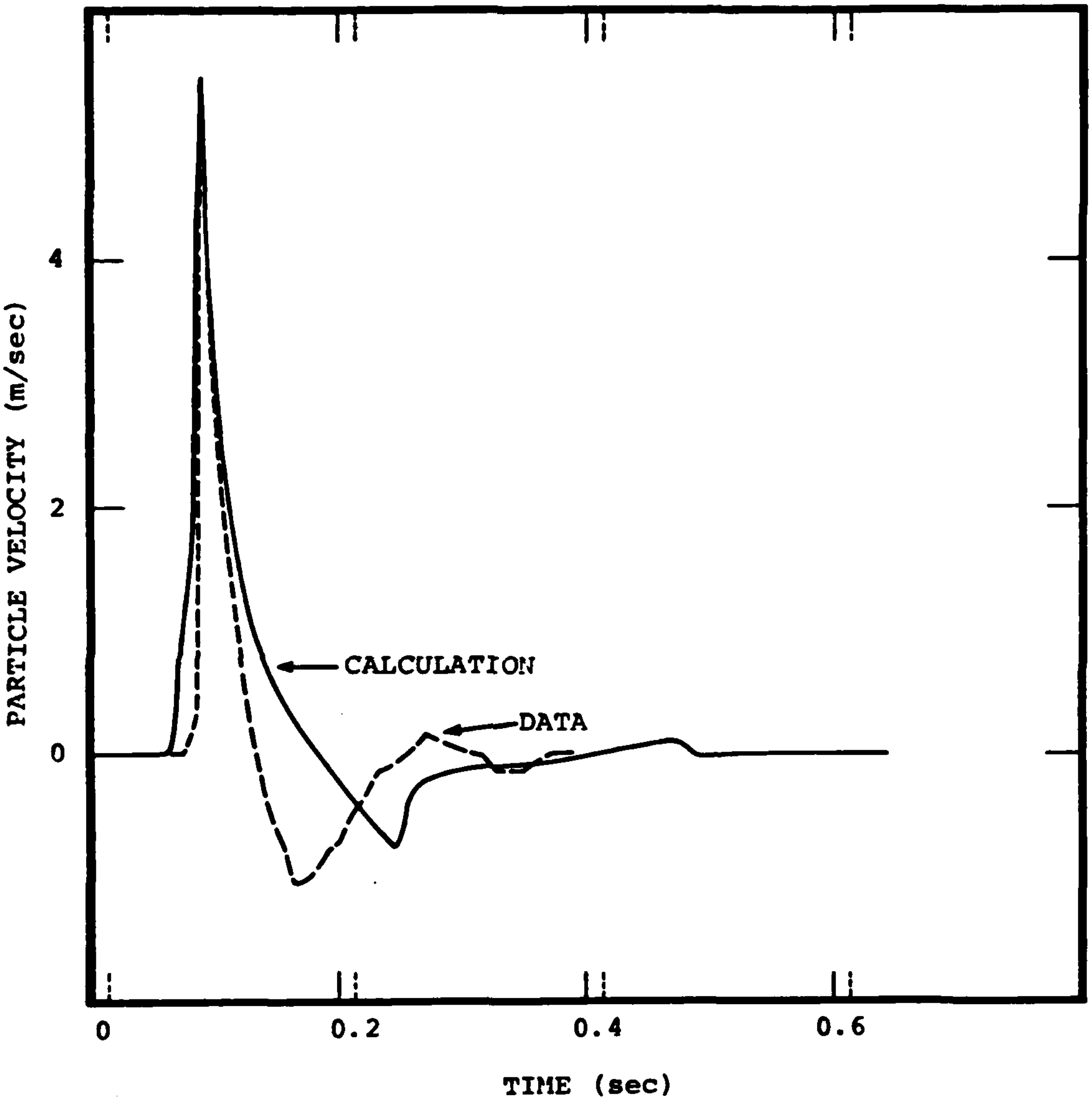

Figure 19. SALMON ground motion at 278 m.

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Estimation of Body Wave Magnitude

The main problem with "m_b" is that it is a rather nebulous parameter; simply, it is a function of the largest peak-to-peak amplitude in the first few seconds of P wave motion with adjustment for the period of the arriving phase. The parameter m_b was adapted from the need to systematically order the size of earthquakes. The measure itself has inherent impreciseness as the measure is not related to the physics of the source, but is the largest constructive interference of waves originating at the source, source region, path, 17
receiver region, and receiver. To relate the m_b to the seismic yield, all effects not due to the source must naturally be corrected. These effects are estimated by sophisticated techniques utilizing the wave equation and an earth model to generate synthetic seismograms. The m_b data measurements are made from digital seismograms - where sophisticated techniques could easily be applied, only the "simple" m_b measurement is made. It seems that the most primitive and least understood part of the process is the meaning and utility of the m_b measure in regards to the source strength.

Site Specific Propagation Effects

Waveform complications have been observed at a number of WWSSN and SDCS stations which may be ascribed to site specific propagation effects. Among these stations are RKON, COL, MSO, LON, ATL, GOL, BLA, and GSC. The studies of the Yucca Flats stations at the Nevada Test Site indicate that many waveform complications characteristic of the stations situated on a sedimentary basin may be explained by simple

elastic wave propagation effects within the basin. This is in direct opposition to researchers who have insisted that waveform complications must be treated exclusively stochastically.

To conclude, the nature of waveform complications due to the receiver site are not well understood in general, nor have known site complications at many WWSSN and SDCS stations been explained or correlated meaningfully with the local geology or other geophysical parameters.

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FACTORS AFFECTING THE ESTIMATE OF THE m_b/YIELD BIAS FOR SOVIET UNDERGROUND NUCLEAR EXPLOSIONS

In order to determine the magnitude of the m_b/yield bias for underground nuclear explosions, we must first understand what the critical controlling parameters are and what the uncertainties may be in our knowledge of each parameter. We must then effect a synthesis of these diverse factors, which tend to be entwined in feedback loops, into a coherent picture. The accuracy of our initial assumptions, the trade-offs between different factors, and the limitations of our measurement techniques must be examined.

The single most important seismological parameter in this process is attenuation. This is also perhaps the most difficult geophysical parameter to determine empirically or to understand theoretically. It will, however, be the focal point of our discussions of the m_b/yield bias question. Our discussions of seismic attenuation will include both anelastic dissipation and elastic attenuation due to elastic scattering and propagation effects which influence seismic amplitudes. We will attempt to integrate the various t* and Q measurement studies, across a wide frequency range (4 hz body waves through long period surface waves) with amplitude and m_b studies into a coherent and consistent estimate of the yield bias. The following text outlines some of the considerations which will be included in this analysis.

Other important factors which influence our ability to accurately estimate this bias and, hence, the yields of Soviet explosions are such

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phenomena as spall phases, instrumental responses, uncertainties in seismic source descriptions, and the effects of changing lithology and depth of burial. In the following discussion we will touch upon each of these. In our forthcoming reports a more thorough treatment will be given to all of the above.

One of the features of seismic attenuation which has recently received renewed emphasis is that of the frequency dependence of attenuation. It now seems clear that this is an important effect within the seismic band of interest in discrimination and detection research. As such, it clearly must be properly included in our analyses. While this in part complicates our procedure, it offers the somewhat compensating advantages of unifying many of the divergent conclusions and results reached in recent studies. The functional form of the dependence, the variation of that dependence with pressure and temperature and, therefore, with depth in the earth, regional variations due to differing lithology and states of material need to be carefully analyzed. Studies of the physical mechanism of seismic attenuation may provide insight into the problem of predicting attenuation characteristics of sites where direct measurements are lacking.

Another area of significant impact on the estimation of yield bias is the determination or selection of appropriate source descriptions for underground nuclear explosions. This aspect of the problem has been approached in the past from three directions: large scale numerical modeling, theoretical mathematical models, and empirical determinations. Questions remain with each approach. This is a serious problem since severe trade-offs can exist between attenuation and source function.

We will try to quantify the uncertainties that currently arise from this situation and their influence upon bias estimates. The accuracy of numerical simulations in modeling the explosion process and the properties of and interactions with the surrounding material need to be considered and is currently underway. The determination of the radius of non-linear material behavior is also of significance.

A third area of concern, closely related with the attenuation concerns mentioned earlier, are regional differences. This is more than just differences in anelastic properties and the selection of analog regions. Regional differences can produce varying propagational effects on seismic energy and potentially mask the anelastic properties. Near-source and near-receiver environments are important in near-field, regional, and teleseismic applications. The problem of propagation to regional distances, while important in some detection applications, seems beyond the scope of the current report.

Consideration must also be given to evaluating the methods and techniques we employ in measuring and estimating the parameters discussed above. Both time domain and frequency domain methods have been employed and have, at times, yielded divergent results. It is important to evaluate the strengths and weaknesses of each approach. At least in part, the frequency dependence of attenuation may unify these results. With respect to time domain measures, results should be interpreted in terms of what portion or portions of the record have been utilized. The usual m_b measurement, for example, may be flawed by not always measuring a consistent phase from observation to observation. It is almost certainly wiser to try to utilize more of the information contained than merely a

single peak-to-peak amplitude. However, a very large body of research exists based on m_b-type observations which must be carefully considered in our overall synthesis. This also includes such modified scales as Marshall, Springer, and Rodean's m_Q technique.

Other considerations include instrumental errors, depth of burial, and spall phases. Incorrect instrument calibration or poorly determined instrument response could seriously affect observation conclusions. Indeed, Langston feels that with respect to WWSSN stations at least, much of the observed amplitude scatter can be attributed solely to inaccurate instrumental gains. Careful selection of high quality sites should minimize or eliminate this problem. Bache and others have examined spall phases and have presented a very reasonable case that they can be important contributors to the observed waveforms, particularly in terms of m_b-type measures. This effect can be expected to depend not only on yield but also depth of burial and surrounding lithology making it potentially a very complex contributor.

Despite the abundance of questions raised in the preceding discussion, the picture is not overly bleak. We do know quite a bit about these phenomena and the magnitude of their effects. We can also realistically expect to continue to refine that knowledge and hence our ability to estimate bias in seismic yield determinations.

Sierra Geophysics, Inc.

SGI-R-80-017

REVIEW OF MAGNITUDE/YIELD ESTIMATION

PRELIMINARY REPORT

G. R. MELLMAN

R. S. HART

TOPICAL REPORT

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PREFACE

This report reviews and summarizes DARPA-supported research on attenuation and magnitude/yield bias estimation. Our review of this program is continuing and additional comments, and an increased breadth of scope, will be presented in a later report. This report concentrates on intrinsic attenuation and on m_b bias measurements. It is not our position that m_b-type measurements represent the best characterization of explosion yield. However, until more advanced and more robust techniques have been developed for this purpose and applied to a significant database, our conclusions must be based on existing data and methods.

REVIEW OF MAGNITUDE/YIELD ESTIMATION

PRELIMINARY REPORT

I. INTRODUCTION

This report presents an analysis of the current state-of-the-art in the estimation of magnitude and yield biases for the evaluation of underground explosions. Our understanding of the highly complex phenomena involved, and of the trade-offs existing between different parameters affecting these results, is the subject of continued research and study and thus our abilities in this area can reasonably be expected to be refined in the future. However, it is important now to take a critical look at the current state of knowledge. From that point, we can assess both the strengths and weaknesses of our existing monitoring efforts and also which directions for future research show the most promise for improving our estimation procedures. To that end, this report will first detail the individual parameters or effects which influence our estimation techniques, included in the evaluation is a review of previous relevant work. This is followed by a synthesis of that information, recognition of current uncertainties, our current conclusions as to the magnitude/yield bias, and some brief recommendations for future research programs.

II. TECHNICAL DISCUSSION

The order of topics as presented here is arbitrary and some of the phenomena discussed affect several topical areas. We have organized this portion of the discussion into five topics.

2.1 Near-Source Effects. Near-source structure can have a substantial influence on outgoing seismic energy. Strong variations in amplitude, waveform, and frequency content of outgoing, short period P waves have been observed for a number of closely spaced underground explosions (see, for example, Alewine et al., 1977). Similar variations have been noted from a single event recorded at different azimuths (e.g., Hadley, 1979; Hadley and Hart, 1979). It has been demonstrated that these variations result from strong structural variations within the near-source region. The observed phenomena appear to fall into two categories. The first may be typified by the experience with the Piledriver event at the Nevada Test Site. Figure 1 illustrates some of the observed short period P waves, the ray paths to the WWSSN stations, and the principal geological features in the immediate test area. The effect to note here is the very dramatic degradation in observed amplitude for observations in which the outgoing ray path intersects the Boundary Fault to the east-northeast of the event site. The waveforms at these stations are also substantially more complex than those observed at other azimuths. This is attributed to a strong structural interaction of the short period P waves with the fault zone (which marks a significant impedance contrast). Such an effect is dependent on both azimuth and take-off angle (distance). Clearly, if magnitude or yield estimation were conducted for this event over only a small azimuthal window to the east-northeast, the explosion yield would be badly underestimated. This error can be avoided by using data recorded over a wide azimuth range

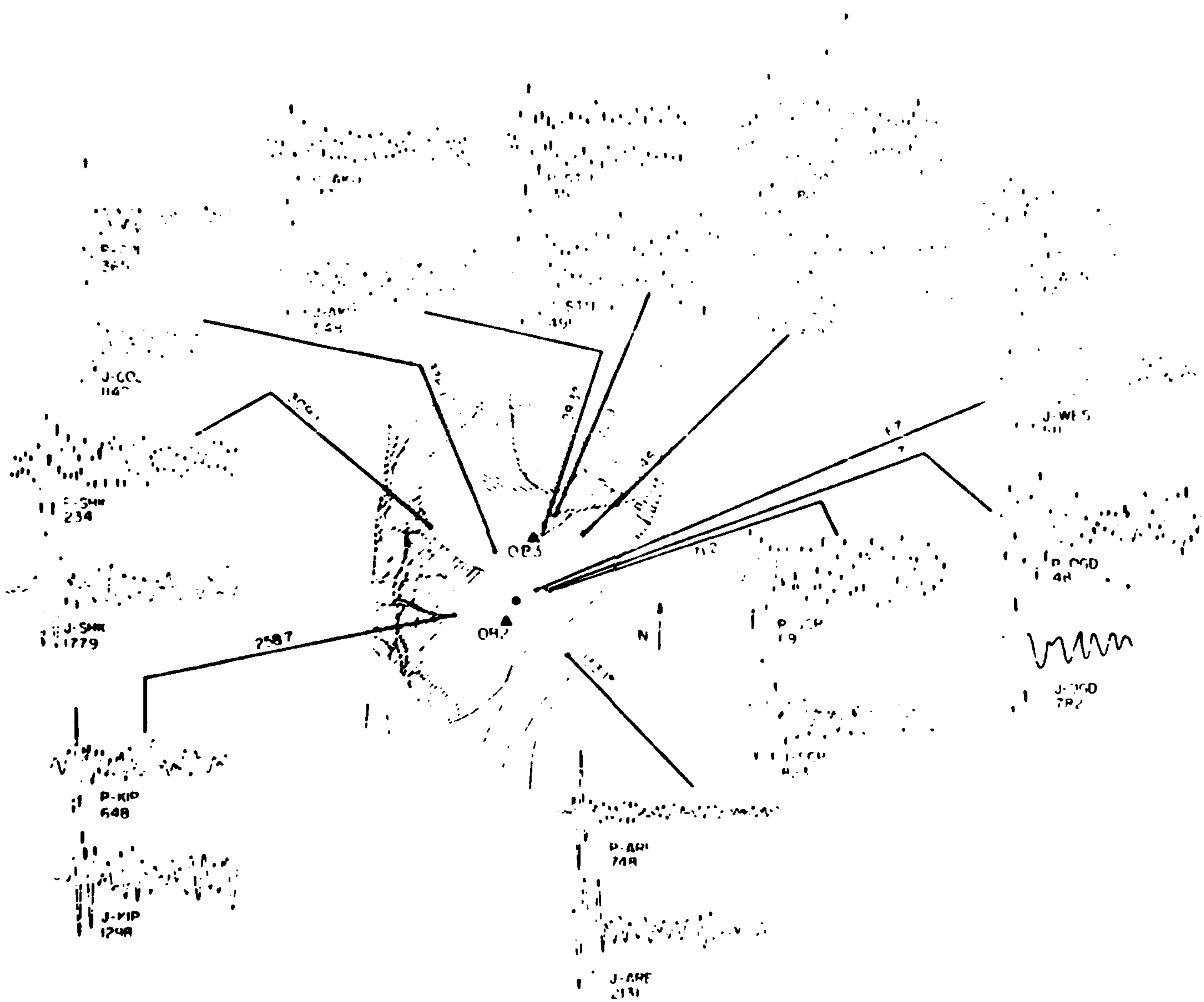

Figure 1 - Comparison of the waveforms and amplitudes from Piledriver (P) and Jorum (J). Generalized map shows the Climax Stock, location of the SDCS stations OB2NV and OB3NV and the location of Piledriver (dot north of OB2). Lines radial from Piledriver show the great circle azimuth to the WWSSN stations. Excluding stations TOL, WES, SCP, and OGD, the average amplitude ratio is 0.30. Amplitude and waveforms for these four stations, for Piledriver relative to Jorum, are both attenuated (average ratio = 0.08) and distorted. These data suggest that the deep structure of the Climax Stock, along a northeast azimuth from Piledriver is significantly altering outgoing seismic energy.

and using some seismologic judgment as to which observations to trust. This determination can also be substantially assisted by utilizing even relatively crude knowledge of the structural geology in the area of an underground explosion.

A more difficult near-source condition exists for events in an area such as Yucca Flats at NTS. Yucca Flats is a relatively complicated geologic feature. The long trough-like structure is offset by several large block-faults in the basement. Except for events on the edge of the basin, the azimuthal variations due to structure-induced propagation effects generally do not show the relatively simple perturbation of the Piledriver case. We are forced in this example to perform global averages of the data. The size of this effect on observed m_b is probably less than $\pm$ .2 m.u. (magnitude units) (Alewine et al., 1977). It is important then to carefully examine the data from events outside of the United States for evidence of highly complex source region structure. This type of environment produces complicated waveforms and thus careful examination and modeling of the seismic data can be a reliable diagnostic tool. It seems, at this time however, that the first type of near-source effect, as exemplified by the Piledriver experience, is more likely to occur. This sort of interaction, as has been noted above, can be readily taken into account in the analysis.

Some progress has been made in the development of analytical techniques to model wave propagation in a laterally varying structure (Hong and Helmberger, 1978). This method, glorified optics, is a geometric optics solution and does not include diffraction effects. The method cannot model sharp variations in structure. In very complex and very rapidly changing structure, the method can become quite expensive; nevertheless, it has proved a useful tool in some applications. Hart et al. (1979) used this technique on a fairly

simple model of Yucca Flats and demonstrated that rather small lateral changes in source location (1-2 km) or in source-to-station azimuths can produce m_b variations of ±.15. This result further emphasizes the need for wide azimuthal coverage in the case of a shot medium of the complexity of Yucca Flats.

One highly significant result derived from the observational and theoretical studies of the NTS data is that the structure of greatest importance to the waveform and to m_b-type measurement is located in a rather narrow region surrounding the explosion site (Hart et al., 1979). Structures more than 1-2 kilometers from the working point are not strong contributors to the waveforms and amplitudes of the first few seconds of the short period P-wave. This fact is important in remote analysis of underground explosion sites.

2.2 Site Specific Receiver Effects. This section is closely related to the preceding discussion and will summarize observational results for a number of studies that deal primarily with attenuation measurements. The frequency dependence of seismic attenuation is treated explicitly in a later section and will not be dealt with directly here.

The incorporation of local δt^* station corrections into the discrimination process can be very important. However, because of the currently unresolved frequency dependence of the attenuation operator, the relation between δt^* and δm_b is highly uncertain. Hence the δt^* measurements cannot be used at this time to predict δm_b. It should be emphasized that near-station attenuation resulting from variations in the local crust and upper mantle are included in our definition of receiver functions. A great deal of effort has been expended, especially by Der and his co-workers to isolate and identify the variations in attenuation associated with different seismic observatories. These studies have quantified these variances in terms of δt^*_α and t^*_α. The overall observed differences are not great (Der, 1977), generally less than

0.25 seconds in total variation between shield and basin and range-type locations.

The importance of near-receiver effects on short period seismic energy should not be neglected in magnitude or yield estimation. Several empirical and theoretical studies have shown that non-planar structure beneath a station can significantly affect observations at that site (e.g., Burdick and Langston, 1977; Langston, 1977a, b; Ishii and Ellis, 1970; Rogers and Kisslinger, 1972). Several other studies have examined incoming short-period seismograms recorded at the SDCS array located at Yucca Flats (Der et al., 1979a; Hart et al., 1979) and have tried several techniques to model those very complex seismograms. One interesting result of those empirical studies was that the structure itself could produce a slight slope in the receiver function spectra across the frequency band of 1-4 hertz. This slope corresponds to a $\delta t_{\alpha}*$ of about 0.1. The large amplitude differential between the YF stations and the OB2NV at Climax Stock can be attributed almost entirely to amplification resulting from the lower seismic velocities in the Yucca Flats alluvium. Indeed, the single most important receiver structure parameter in terms of short period seismic amplitude or m_b measurements is sediment amplification. Der et al. (1979b), Butler (1979) and several others have demonstrated theoretically that the presence of even a relatively thin low velocity layer will produce amplification factors of up to 1.8. Der and his co-workers (1979b) found a highly significant correlation between m_b and crustal impedance (Figure 2). In studies of short period P waves from nuclear tests and earthquakes (Butler, 1979; Butler et al., 1979) recorded at seismic stations in the continental United States, some stations have been shown to exhibit large azimuthally-dependent variations in amplitude (up to factors of 2) and

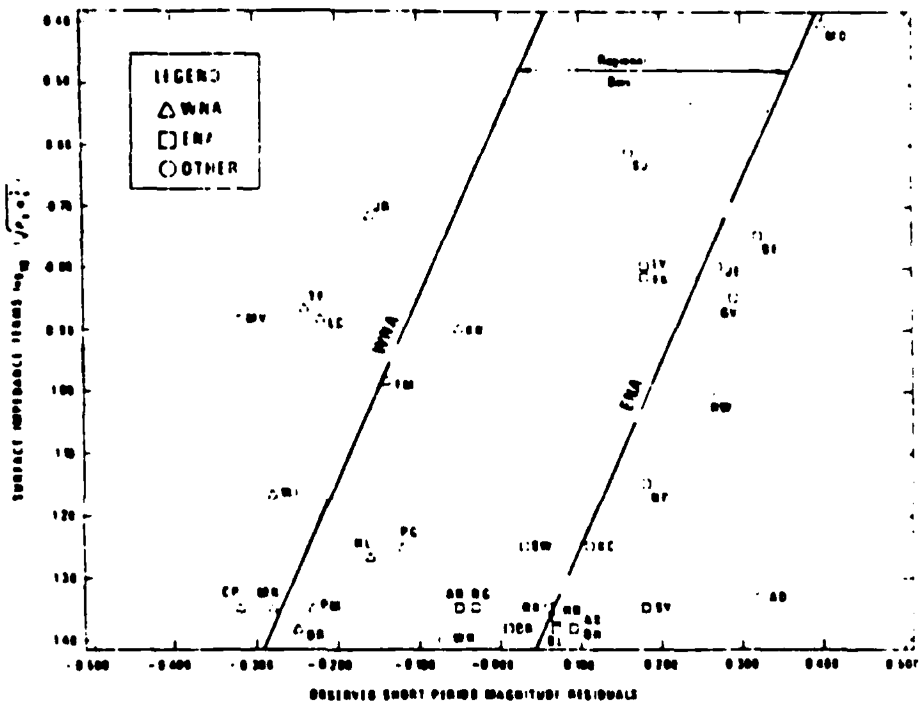

Figure 2 - Surface acoustic impedance terms plotted against short-period magnitude residuals of Booth et al. (1974). Regression lines for the ENA and WNA are drawn separately. (from Der et al., 1979b)

significant complexity. For example, the amplitude response (as measured by the "b" phase value on the short period seismogram) at several WWSSN stations (e.g., AAM, ALQ, ATL, BLA, and GOL) varies by as much as a factor of 2 depending upon the azimuth of the incoming seismic phases (Butler et al., 1979). The azimuthal variation observed at RKON in the same study was even greater (2.8) although that result is based on a limited number of observations.

The net effect of these observations is to raise a warning flag with regard to conclusions drawn from stations whose receiver function is inadequately determined. Clearly the accuracy and reliability of yield estimates is a function of the resolution of the receiver functions, or station corrections, used in the determination. If we are careful and select data from stations whose receiver function has been determined and the necessary corrections applied, receiver function variations should cease to represent an important contribution to yield bias errors.

2.3 Absolute Q Measurements. Of all the factors that influence amplitudes of seismic waves, probably the most difficult to reliably estimate is attenuation. This is due primarily to the trade-off that exists between attenuation effects and source or propagation effects. Indeed, even when source and propagation effects are eliminated, there still remains a virtually unresolvable tradeoff between losses by anelastic attenuation and losses by scattering.

Despite these problems, a number of determinations of the effective Q for the earth have been made over the frequency range of ~.001 to 5 hz. At the low frequency range, free oscillations and surface waves data have been used. At higher frequencies, body wave data are employed. Since our primary interest in this report is yield bias, we will concentrate primarily on body wave measurements. Moreover, we will concentrate on continental, rather than oceanic, source-receiver pairs.

At very long periods, free oscillation data are used to estimate attenuation. The Q's of individual modes may be estimated either by observing the decay time of narrow band filtered records or by measuring the broadening of spectral peaks. It is possible by combining the computed eigenfunctions of the unattenuated modes with the individual mode Q's to determine the Q structure as a function of depth. Studies of this type have yielded attenuation versus depth models SL 8 (Anderson and Hart, 1978) and QKB (Sailor and Dziewonski, 1978).

In order to compare results of free oscillation Q studies with results of body wave studies, it is convenient to compute $t^*_\alpha \equiv T/Q_\alpha$ for ray paths corresponding to distances of 30° to 90°. Here T is the P wave travel time and Q_α is the average P wave Q over the ray path. Though both SL8 and QKB produce variations in t^*_α with distance, the average value over the range 30°-90° is slightly greater than 1 sec. It should be noted, however, that this represents a whole earth average and is thus dominated by oceanic structures which exhibit somewhat lower Q's than those typical of continental structures.

Attenuation measurements for surface waves are generally made using a spectral ratio technique. By computing the spectral ratio of two stations on a single great circle path from the source or a multiple traverse of a great circle path at a single station and correcting for elastic effects, it is possible to determine an attenuation coefficient as a function of frequency. Then, from knowledge of the elastic eigenfunction, Q as a function of depth may be determined. It should be noted that source effects are eliminated through choice of stations on great circle paths and use of spectral ratios.

Numerous authors have obtained Q models based on surface wave studies. However, most of these have been based on primarily oceanic paths. One

example of such a model is MM8 of Anderson, Ben-Menahem and Archambeau (1965). This model predicts t^*_α values of ~.9 sec. for distances of 30°-90°. Another global average model, that of Mills and Hale (1978a), produces Q values about 20% below those of MM8 on the average. This would correspond to $t^*_\alpha \simeq 1.1$ sec., which is slightly more consistent with the SL8 value.

Lee and Solomon (1978) have modeled surface wave attenuation for continental paths. In general, their models, which extend only to a depth of ~ 350 km, produce significantly lower Q's than do the models previously discussed. In particular, their model for the western U.S. gives an average $Q_\beta \simeq 40$ ($Q_\alpha \simeq 90$) for the upper 350 km. of the mantle. This value is less than one-half the averaged values of the other studies and would produce $t^*_\alpha \approx 2.5$ sec at 15°, the approximate distance at which the geometric ray bottoms at 350 km. Causes for the discrepancy between this model and other surface wave Q models are not immediately obvious. However, the extremely low Q values obtained in this study make the results somewhat suspect.

Surface wave and free oscillation data thus seem to give $t^*_\alpha \approx 1$ sec. for globally averaged models. While such studies are relatively insensitive to source effects, errors may be introduced by the presence of scattering and lateral inhomogeneities. Further, all studies have assumed frequency independent attenuation in the surface wave-free oscillation band. The presence of frequency dependence in this frequency range would have a very definite effect, particularly on the surface wave models since depth dependence of Q is inferred from frequency dependence in these models.

The problems encountered in making absolute Q estimates using body waves are somewhat different from those encountered in surface wave and free oscillation studies. In body wave studies, it is still necessary to eliminate the effects of both source and path. However, this is less easily

done for body waves than for surface waves due to greater uncertainty in earth structure at high frequencies and the fact that the apparent earthquake source time functions change not only as a function of azimuth but as a function of distance as well. By restricting distances considered to 30°-90°, it is possible to largely eliminate elastic path effects other than geometric spreading. We will therefore first consider studies performed in this distance range, together with several other studies where path effects may be easily taken into account.

In certain special cases, it is possible to use spectral ratio methods to eliminate source effects for body waves without making the assumption that the source is known. Two such cases are ScS_n-ScS ratios and P-PP or S-SS ratios.

In the case of ScS_n-ScS ratios, as long as the source-receiver distance is not too large, ScS amd ScS_n emerge from nearly identical locations on the focal sphere. The ScS_n/ScS spectral ratio then may be seen to give the attenuation factor for the appropriate number of surface-to-core bounces, together with the geometric spreading factor and the core and surface reflection coefficients. If SH polarized waves are used, or if the ray parameter is sufficiently small, the reflection coefficients at the core-mantle interface and free surface are essentially unity.

Several studies using long period ScS_2/ScS ratios have been done. These include Kovach and Anderson (1964), who obtained $Q_{ScS} = 600$ for South American events and stations, and Yoshida and Tsujiura (1975) who obtain $Q_{ScS} = 290$ for the Sea of Japan.

The most recent, and probably the most reliable, study is that of Sipkin and Jordan (1979). Using HGLP and WWSSN-LP data, and a stacking and inversion technique that improves the signal-to-noise ratio over more standard spectral

ratio methods, they find a Q_{ScS} = 155 for the Western Pacific in the .006 to .06 hz band. For continental data, they infer an average Q_{ScS} = 225, with a somewhat higher value Q_{ScS} = 285 for South America. Regional Pacific Ocean values varied from a low of Q_{ScS} = 140 to a high of Q_{ScS} = 200.

Much of the discrepancy between Q_{ScS} studies appears to be due to the different methodologies used. As Sipkin and Jordan (1980) point out, the presence of incoherent high frequency noise in ScS_n tends to bias results of the simple spectral ratio and averaging procedure. This noise is rejected by the technique employed by Sipkin and Jordan and the similar technique used by Nakanishi (1979). In particular, analysis of the same data by Best et al. (1974) using the spectral ratio method and Jordan and Sipkin (1980) yield values of Q_{ScS} = 300 and 200, respectively, while results of Nakanishi in general agree extremely well with Sipkin's results.

While Q_{ScS} is not directly translatable into t* due to differences in the manner in which these rays average the earth, some idea of t* values consistent with Q_{ScS} studies may be obtained by comparing measured Q_{ScS} values with those computed from free oscillation models. Both SL8 (Anderson and Hart, 1978) and QKB (Sailor and Dziewonski, 1978) predice Q_{ScS} = 230-240. The somewhat lower values of Q_{ScS} obtained by Sipkin and Jordan thus would imply a t^*_α of slightly greater than 1 sec. which is predicted by SL8. For average inferred continental values of Q_{ScS} a $t^*_\alpha \approx 1$ sec. seems consistent.

A similar approach to that used on ScS/ScS_n pairs may be used on P/PP or S/SS pairs. That is, we choose two stations at the same azimuth but with the second station twice the distance from the source as the first. The PP phase at the second station follows exactly the same ray path as the P wave at the first station. Thus, the PP phase at the second station will be related to the P phase at the first station by a free surface reflection

coefficient, a geometric spreading factor, a Hilbert transform, and an effective attenuation operator for the path from the first station to the second.

Butler (personal communication) has used underground nuclear explosions in the Aleutians and Novaya Zemyla as source and WWSSN stations in North America as receivers. Rather than examine spectral ratios, he has used Futterman (1962) or Minster (1978a, b) Q operators and elastic transfer functions to model in the time domain PP at the second station from the P waveform at the first station. Because of this, his results are principally dependent on the absolute amplitude effects of the attenuation operator rather than the relative frequency dependent effects on which spectral ratio methods depend, although waveform matching considerations do provide some sensitivity to the frequency dependent effects within the instrument passband. Preliminary results indicate t^*_α = 1 - 1.3 sec for both long and short period P/PP pairs.

Several factors may bias the results of this last type of study. These include the effects of near receiver structure near the stations, and the effect of structure near the bounce points. The magnitude of these effects is not known at this time. It should also be pointed out that the Hilbert transform relation between P and PP is valid only when the turning point of the P wave occurs in a region of gradual velocity change. Upper mantle triplications should therefore be avoided.

The phase pairs mentioned above are the only cases useful for the investigation of the mantle where spectral ratios eliminate source variations for the phase involved. In the case of other phase pairs, such as P/PcP, differences in takeoff angle may have a significant effect. Kanamori (1967) used such pairs to study core reflections and to produce a mantle attenuation model. Using the Tonto Forest Array and a spectral ratio method, he determined t^*_α values in the range 1 - 1.5 sec. However, no correction was made in this study for the differing radiation pattern between the P and PcP rays.

Where a second phase is not available to remove source effects, log spectral slope methods may still be used if some source estimate is available. Such a source estimate may be derived from near source measurements, or it may be derived from some source model. Most, but not all, such studies have used high frequencies, from ~ .5 - 4 hz, since the exponential decay of the attenuation operator with frequency reduces the error caused by uncertainty in the source spectrum at high frequency.

Der and McElfresh (1976a) have used the log spectral slope method at high frequency, together with close-in source spectral measurements and scaled source estimates for underground nuclear explosions to estimate t^*_α for a number of test sites in the U.S. For source-receiver distances of 30° or greater, they obtain t^*_α = .2-.5 sec. These values generally segregate into two classes, with the lower values corresponding to the shield-shield type paths and to the higher values western U.S.-shield type paths. These results are typical of a number of such studies at high frequency. These include studies by Frazier and Filson (1972) using NORSAR to observe NTS events (t^*_α = .4), and studies by Noponen using NORSAR and earthquake sources corrected for $1/\omega^2$ falloff (t^*_α = .2-.4).

Spectral slope estimates have been used at longer periods by Sipkin and Jordan (1979) to estimate Q_{ScS} at periods where multiple ScS data is unavailable. Q_{ScS} estimates were made for both the long period WWSSN instrument band and the short period band. Source estimates used were a delta function and a simple Brune (1970) type source with corner frequency of .16 hz. Q_{ScS} values for these two sources ranged from 400-1000 for the long period band (.1 hz-.5 hz) and 1000-2000 for the short period band (1 hz-2.5 hz). These values are considerably larger than the estimates of these same authors for frequencies less than .01 hz.

Sipkin and Jordan (1979) use still another measurement technique in their Q_{ScS} estimates. This is the energy ratio approach in which the integral of the squared seismogram is computed over a specified interval. This is done for both the long and the short periods. Each integral is a measure of the power through that instrumental pass band. Then, given the assumed source spectrum, and using Parseval's theorem, an attenuation operator may be found such that the energy ratio of the synthetic matches the energy ratio of the data. Using this method, Q_{ScS} of 300 and 500 are determined for the two assumed source functions.

The energy ratio method is, in fact, a time domain analog of the spectral slope method. However, rather than weighting all frequencies equally, the energy ratio method weights most heavily those frequencies for which the actual recorded response is largest on each instrument. This provides a certain degree of noise stability, but at the expense of much of the dynamic range of the individual instruments.

The largest single source of error in any of the spectral slope methods is the uncertainty in the source. This includes both errors in the source time function and errors in the effect of near source structure. When underground nuclear explosions are used, source errors may also arise from ignoring the effects of near field and non-linear terms on the source estimate.

Near receiver structure may also be a source of bias. Der e[illegible] 1977) have considered a number of plane layered structures and in general s[illegible] that no appreciable bias exists. Hart et al. (1979) show that for actual computed crustal transfer functions for a station in a sedimentary basin relative to a station on competent rock, a bias in t^*_α of .1-.15 sec. exists for the .5 to 4 hz frequency band. This difference is small, however, and is considered an extreme case.

All methods using source estimates that we have examined so far have made use of the relative behavior of the attenuation operator as a function of frequency, rather than the absolute amplitude decrease associated with the attenuation operator. Several studies have used estimates of the time functions of underground nuclear explosions at NTS to determine Q by matching the absolute amplitude and early portion of the waveform at teleseismic distances.

Trembley and Berg (1972) used a computed time function to match close-in and teleseismic records at NPNT in the time domain. In this manner, they obtained $t_0^* = 1$ sec. Bache et al. (1975) used a reduced displacement potential (RDP) calculated assuming a non-linear rheology to estimate $t_\alpha^* \approx 1.05$ sec for an average amplitude and wave shape observed at a number of WWSSN stations. Hadley (1979) has studied a number of the same events using estimates of the RDP derived from modeling of near field records. The value obtained for t_α^* was 1.3 sec. We note that with this value it was possible to fit not only short period amplitude, but the early portion of the short period waveform and long period-short period amplitude ratios as well.

Once again, the principle cause of error in these t* measurements is related to errors in effective source estimation. In particular, the entire difference between the results of Hadley and Bache is caused by differing methods of estimating the RDP. The decision as to which result is indeed more correct is critically dependent on the radius at which displacements are well described by linear elasticity. This is an important issue not only for Q estimation, but for magnitude-yield and source scaling results as well, since many of these results are dependent on close-in observations. At this time, the importance of non-linear effects on Hadley's source estimates, which are based on recordings made ~ 8 km. from the source, is not known.

Another possible source of bias in these estimates is the use of linear elasticity to describe the pP arrival in the far field. Studies by Blandford (1976), Blandford et al. (1977), and others indicate that, in many cases, the amplitude of pP is considerably reduced compared with what would be predicted by linear theory. We note, however, that this is highly dependent on material properties and depth of burial. Also, we anticipate that long periods at teleseismic distances will be less affected than short periods. Thus there exist significant tradeoffs between the amplitude, frequency content and arrival time of pP and the estimated t*. Der (personal communication) has shown that this could result, in the worst case, in significant overestimation of t*. At this time, however, it is not possible to assess the probable size of this error.

In the methods discussed above, source estimates have been based on near field observation or calculated from some simple source models. If we assume that all attenuation losses occur in shear, so that $t^*_\beta = 4\ t^*_\alpha$, and that source time functions for P and S waves are the same, it is possible to determine source and Q estimates directly from far field observations. Burdick (1978) used a long period-short period amplitude ratio method, similar to the energy ratio method described previously, to estimate t* for four deep South American events observed in the continental U.S. on WWSSN instruments. In this method, a Futterman operator is applied to P-waveforms to produce predicted S-waveforms. In this manner, long period-short period amplitude ratios for S waves may be predicted as a function of t*. The t* value that produces the best fit to the data is then chosen. This method produces $t^*_\alpha = .7$ sec for these deep focus events, which corresponds to $t^*_\alpha \simeq 1$ sec for surface focus. A similar method, using energy ratios, was used to estimate Q from east coast WWSSN stations using sS and sP phases from the Borrego Mountain earthquake. In this case, a $t^*_\alpha \simeq 1.3$ sec was obtained.

If long period nodes are avoided, energy ratios should be quite good in eliminating source radiation pattern effects. The greatest possibility for error appears to lie in the assumption that P and S wave time functions, or upgoing and downgoing S wave time functions, are identical. If strong directivity is present, this will certainly not be the case. Additional sources of error include contamination of individual phases by other phases produced by near source inhomogeneities, and the effects of near-receiver crustal structure.

As noted earlier, attempts to obtain Q estimates for body waves where significant elastic path effects are present, such as upper mantle distances, are considerably more difficult than studies of data in the 30°-90° range. This is particularly true at short periods. The problems that arise are not only problems of frequency dependent propagation effects, but also problems of identifying the path that any given arrival followed in the earth. Despite this, a number of studies have been done using data from 1°-30°.

Archambeau, Flinn and Lambert (1969) have attempted to simultaneously estimate velocity and Q structure for profiles extending from NTS. They estimate velocity structure from travel times and amplitudes. This Q structure is determined simultaneously with velocity structure in order to fit the amplitude data. With this method, a Q model with $t^*_\alpha \approx .5$ sec was produced for 30°-90° distances.

Unfortunately, the methods used to determine synthetic amplitudes and amplitude derivatives were geometric ray theory and classical head-wave theory. The shortcomings of the former in the neighborhood of caustics are well known. Classical head-wave theory, too, has significant problems if gradients exist above and below the interface. Amplitudes at high frequencies may also be drastically affected by the presence of a gradient,

rather than a sharp discontinuity. These problems, together with the large degree of scatter in the amplitude data and the large tradeoffs that exist between Q and structure in predicting amplitudes cast serious doubts on the accuracy of this Q model.

Der and McElfresh (1976) have attempted to use the high frequency spectral slope method to estimate absolute Q's for several underground nuclear explosions in the U.S. These include Salmon, Gnome, Mast and Knickerbocker. No attempt is made in this study to model propagation effects or to identify the travel paths of individual phases. Instead, it is assumed that propagation effects will not produce a trend over the .5-4 hz frequency band. While this may be true, we note that the structure introduced to the log spectra, by multiple arrivals, particularly at the long period end, will probably add to the scatter of the computed t* estimate. Somewhat more serious, particularly with regard to the Salmon study, is the fact that for stations at distances less than 15°, records at the same distance but different azimuths do not appear to have similar envelopes. This probably indicates major differences in the velocity structure for those azimuths. One possible candidate for such a structure is a thin high velocity lid overlying the low velocity zone for certain azimuths. Large high frequency arrivals, which are largely unattenuated since they do not penetrate the low Q zone (however small it may be), are produced by models with this feature, and may in fact be first arrivals at distances of 15°-17°. At distances greater than ~17° envelopes are more coherent from azimuth to azimuth and it seems likely that azimuthal differences in frequency content are, in fact, diagnostic of true attenuation differences.

In the preceding discussion, we considered a number of methods for obtaining estimates of absolute Q, together with some of the effects that

might introduce bias into the results. In general, the largest source of error in most methods was in imprecise knowledge of the source. Indeed, without some outside physical constraints on the source, there exists a nearly total tradeoff between source and attenuation. The quality of the results of individual studies are therefore often directly dependent on the validity of the source assumptions employed.

In attempting to assess the compatibility of the results of the various studies that have been presented, the question arises as to what degree different methods might provide different answers for the same data set. A partial answer to this question is provided by Sipkin and Jordan (1979). Using the same set of source assumptions and the same data, they calculated Q values from long and short period spectral slopes and from long period-short period energy ratios. The two methods give quite similar results, as one would hope. This gives the investigator some confidence in his ability to choose the method best suited to the requirements of his particular data and still get answers consistent with other studies. As noted earlier, time domain methods are particularly useful in situations where noise or spurious arrivals may be present. This includes hand-digitized data, where digitization noise and record skew may introduce large effects. Frequency domain methods, on the other hand, take better advantage of the dynamic range of the instrument.

If we take the values of the preceding section at face value, the question still arises as to whether a single Q model can be found from this data, or whether actual incompatibilities exist between measured values of Q. One fact that emerges almost immediately is the requirement for a frequency dependent Q. At the long period end, free oscillations, surface waves and long period body waves all produce values of $t^*_\alpha = 1$ sec.,

it would be impossible to observe 3-4 hz. energy for reasonable source models. On a number of occasions such energy is clearly observed. This is a fact that has been widely appreciated by those working with spectral methods and widely ignored by many working with time domain methods. Use of narrow band filters indicate that this high frequency energy arrives at the time of the first P-wave arrival and is not the result of windowing effects (see, for example, Hart et al., 1979).

Several authors have examined the frequency dependence of Q from a data standpoint. Sipkin and Jordan (1979) have shown frequency dependent Q_{ScS} may be explained by a single relaxation model (Liu et al., 1976). These results are shown in Figure 3. Lundquist (1979) has shown that body and surface wave Q data may be explained by a double absorption band model. Figure 4 shows the dependence of t* with frequency for this model.

It should be noted that with a frequency dependent t*, estimates based on the relative changes in amplitude in some frequency band will generally give a biased estimate of the absolute attenuation over that band. This is simply a statement that, for a function of non-constant slope, an estimate of the slope at a point does not determine the value of the function at that point. In general, if t* is a decreasing function of frequency, spectral slope and energy ratio methods will tend to underestimate t*. Thus, the apparent large discrepancy between short period absolute t* measurements of Bache et al. (1975), and Hadley (1979) and the spectral slope t* estimates of Der et al. (1976a) may be considerably smaller than appearances would first indicate. We also note that this effect would mean that high frequency t* estimates made using spectral slope methods are not necessarily equivalent to absolute amplitude measurements such as m_b even when all source and elastic propagation effects are known exactly.

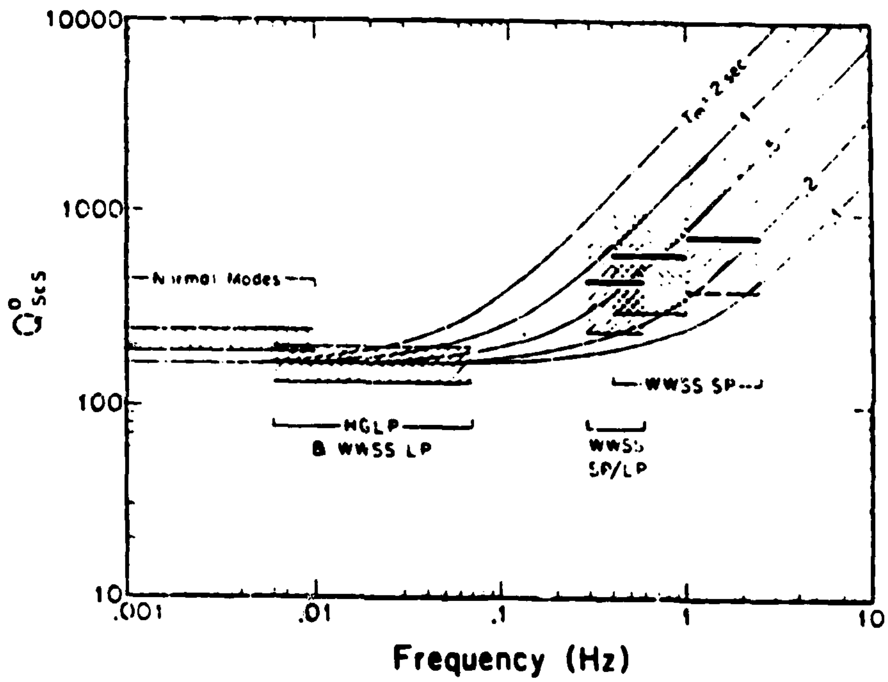

Figure 3 - Summary of constraints on Q^{o}_{ScS}. Lower bounds on WWSS estimates derived by assuming $f_c = \infty$; best estimates are indicated by heavy solid lines, derived by assuming f_c = 0.16 Hz. Solid curves computed from the absorption band model for $\tau_M = 10^4$ sec, and the τ_m values shown. (From Sipkin and Jordan, 1979.)

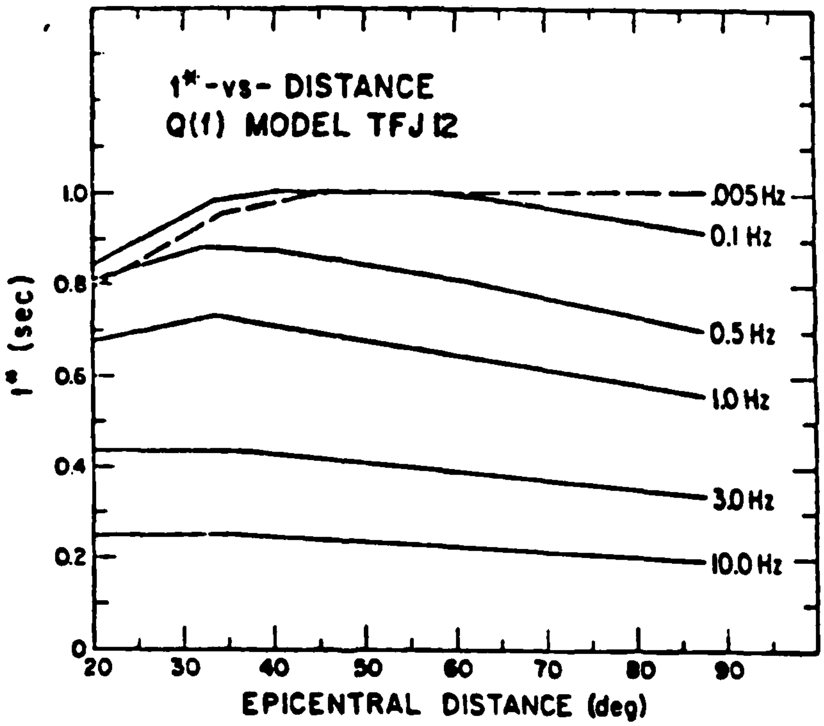

Figure 4 - t*-vs-distance for selected frequency. The velocity model is QM2, and the Q model is the double absorption band model, TFJ12. t*(TFJ12) converges to t*(AFL) at 3 Hz and converges to t*(SL8) as frequency decreases. (from Lundquist, 1980)

The introduction of frequency dependence also has implications for the study of lateral variations in Q. There are now several additional parameters that may be varied as a function of position, including the location of the edge of the absorption band. There is thus no reason that one station cannot have a lower t* than a second station at short periods and a higher t* value at long periods.

Studies of long period ScS by Sipkin and Jordan (1980) indicate that long period Q does, in fact, show regional variation. Such variations have been shown to occur at high frequencies by Der at al. (1976a). Whether the location of the absorbtion band shows systematic changes dependent on the absolute level of either high or low frequency Q is not known at this time.

2.4 Relative Q Measurements and Regional Variations. While absolute Q measurements are highly desirable for certain types of studies, such as relating near field and regional measurements to teleseismic measurements or performing waveform inversions for modeling sources or structure, relative Q measurements between regions or between individual receiving stations are important for a number of purposes. One such purpose is the calibration of individual source sites to a different source site where magnitude-yield relationships have been reasonably well established.

The principle reason for performing relative Q measurements, rather than absolute measurements, is that it is much easier to reliably eliminate source effects. In terms of spectral slope estimates of relative Q, source effects may be eliminated by dividing the spectrum at a receiver by the spectrum of some reference station. We note that, while such a procedure would at first appear to be effective in removing the source time function indpendent of source orientation, some problems are encountered for near-nodal stations. This is partly a statement that faults are not purely linear

features and partly a statement that different frequencies "see" different size regions of the focal sphere. Thus, stations which exhibit long-period nodal behavior for an event often still show large short-period arrivals. Spectral ratios for this type of station would therefore show an anomalously high Q value. Problems may also be encountered with shallow earthquakes due to complications in the spectrum introduced by possible size changes of surface reflected phases between stations.

The two basic methods used to estimate relative attenuation are the spectral ratio method, which we have just discussed, and the amplitude ratio method. Amplitude ratio methods include m_b studies, with and without corrections for predominant period, and studies of first peak to first trough amplitudes, the so-called "b" phase. Amplitude ratios contain a host of information in addition to just attenuation. They are also strongly affected by variations in radiation pattern and by near receiver structure. Radiation pattern effects may be avoided by considering sources which show small variation over the azimuth in question or by averaging a number of measurements for sources with random orientation. Receiver structure problems cannot be avoided in these measurements, although b amplitude measurements would be less affected than would m_b.

Several studies of m_b residuals have been done for the continental U.S. (Evernden and Clark, 1970; Booth et al. 1974). The results of the study of Booth et al. (1974) is reproduced in Figure 5. As may be seen, an east-west magnitude bias appears, on the order of .3 m_b units. This general trend appears in the Evernden study as well.

Der (1977) has attempted to explain this magnitude bias by attenuation differences between the eastern and western U.S. Using a high frequency spectral ratio method, he has determined δt^*_α for a number of LRSM stations,

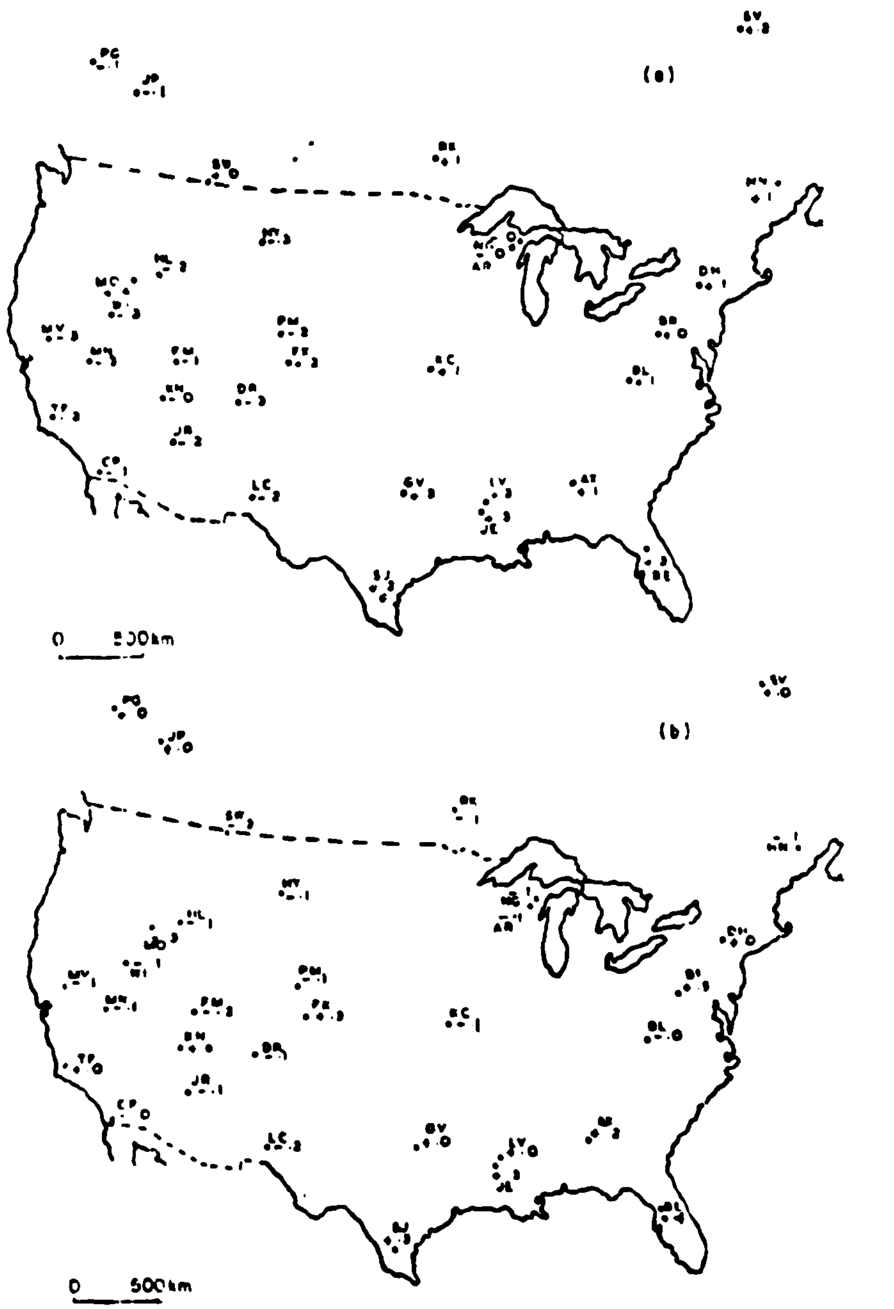

Figure 5 - m_b station corrections for long periods (a) and short periods (b) (from Booth, et al., 1974).

using data from both teleseismic and upper mantle distances. We note that although much of the data is at ranges of 17^{o} and less, where, as noted earlier, propagation effects may have a serious effect on t^*_α measurements, data at larger distances generally support an east-west δt^*_α difference of .2 sec. This difference in t* is sufficient to explain the observed magnitude bias of .3. As pointed out by Der (1977) and Butler et al. (1979), this general east-west bias occurs in a number of other geophysical parameters, such as heat flow, travel time anomalies, and Pn velocities. Additional studies by Der at al. (1979b) indicate that when m_b is corrected for crustal amplification effects, the separation between eastern and western station groups is more apparent.

While a simple east-west dichotomy is certainly appealing, several other studies indicate that the actual situation is somewhat more complicated. Solomon and Toksöz (1970) used a spectral ratio method to study δt^* at long periods. The sources used were deep South American earthquakes, and the receivers were WWSSN long period instruments. Results of this study are shown in Figure 6. While a general east-west trend still exists, there now appears to be sizable low attenuation regions on the west coast and a high attenuation region in the northeast. It should be noted that the total variation in t^*_α observed in this study is greater than 2 sec. even when their extra factor of π is removed. This seems somewhat unrealistic, although their δt^*_α and δt^*_β estimates are generally consistent.

Burdick (1978), in his long period-short period amplitude ratio study to determine absolute Q's using deep South American events, failed to find any resolvable difference between eastern and western U.S. stations. However, as a majority of the eastern stations used in this study were on the east coast, this result may not in fact be drastically different than the Solomon and Toksöz (1970) result.

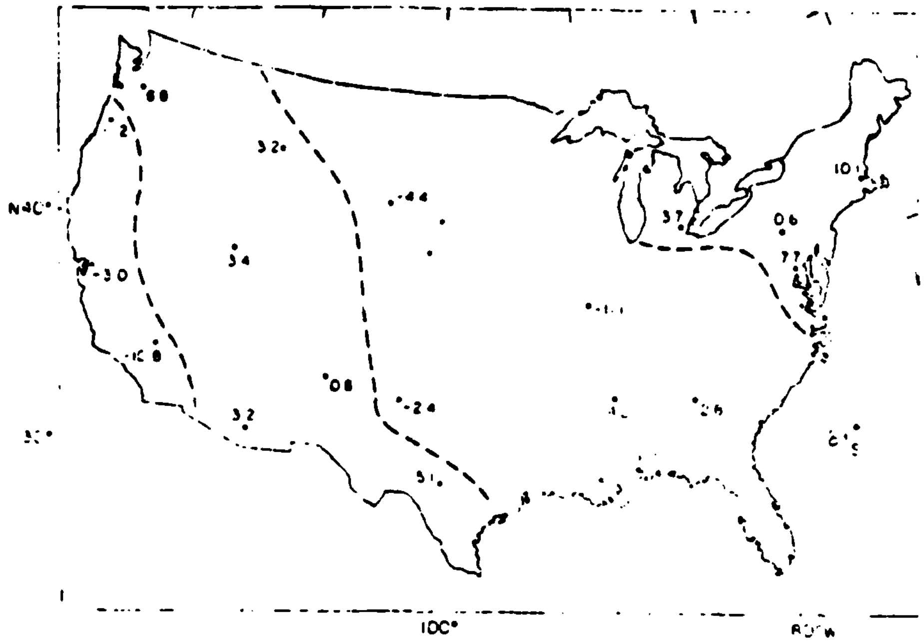

Figure 6a - Lateral variation of S wave differential attenuation at U.S. stations. δt_S* is the average of attenuation measurements from two deep earthquakes in the Peru-Brazil border region. (From Solomon and Toksoz, 1970)

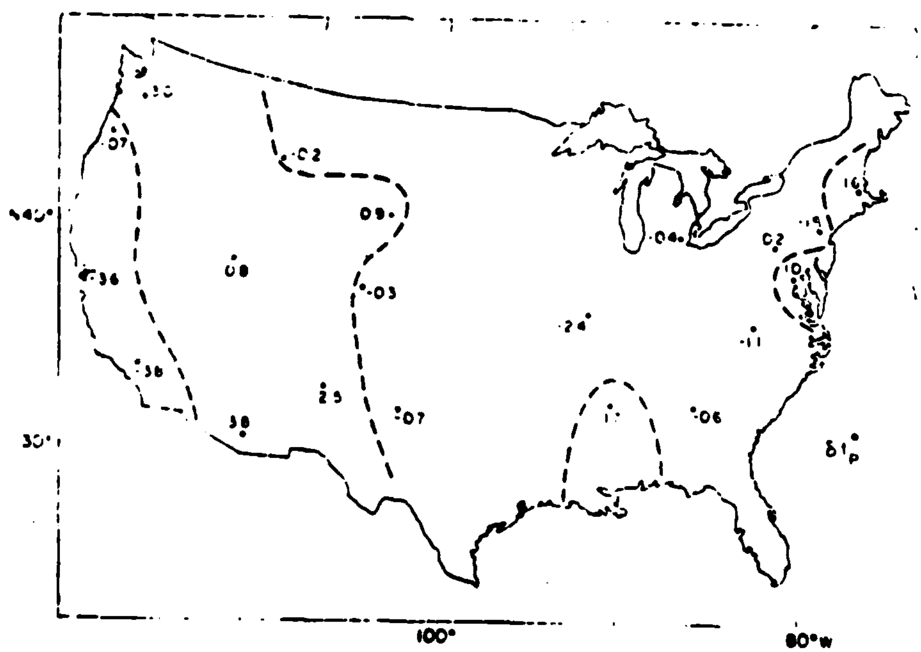

Figure 6b - Lateral variation of P wave differential attenuation at U. S. stations. δt_P* is the average of attenuation measurements (Teng, 1968; Mikumo & Kurita, 1968) from two deep earthquakes in the Peru-Brazil border region. (From Solomon and Toksoz, 1970)

Butler et al. (1979) has studied b amplitudes of short period P-waves at WWSSN and selected SDCS stations using Soviet underground explosions and simple, impulsive, deep earthquakes. His results are shown in Figure 7. While a general east-west difference still exists, we note that the lowest amplitudes are associated with the Rocky Mountain front, not with the Basin and Range. The central U.S. still shows large amplitudes, but now the east coast and west coast, including portions of the Basin and Range exhibit the same intermediate amplitudes. In addition, a fairly strong azimuthal dependence is shown at a number of stations. While this study does not include the effects of crustal amplification, it is not expected that this will affect the overall pattern appreciably.

It should be noted that, while the Butler et al. (1970) study and the Der (1977) study agree in large part, the east and west coast patterns observed by Butler appear to largely destroy the good correlations observed by Der with numerous other geophysical parameters.

Studies by Der (1976) of δt^*, using the high frequency spectral ratio technique for several Novaya Zemyla underground nuclear explosions, give low t^* values for several stations in the Colorado Plateau. These same stations show much larger t^* values for upper mantle distance with sources to the east and southeast.

The impression given by these studies is that anomalously low Q regions appear to be somewhat more localized than would be indicated by a simple east-west dichotomy. This would, in fact, explain a large portion of the discrepancy that appears to exist between the Der and McElfresh (1976b) study and both the Der Novaya Zemyla study and the Butler study. Rays at upper mantle distances would tend to see an averaged upper mantle path which,

SHORT PERIOD P WAVE AMPLITU' E ANOMALIES

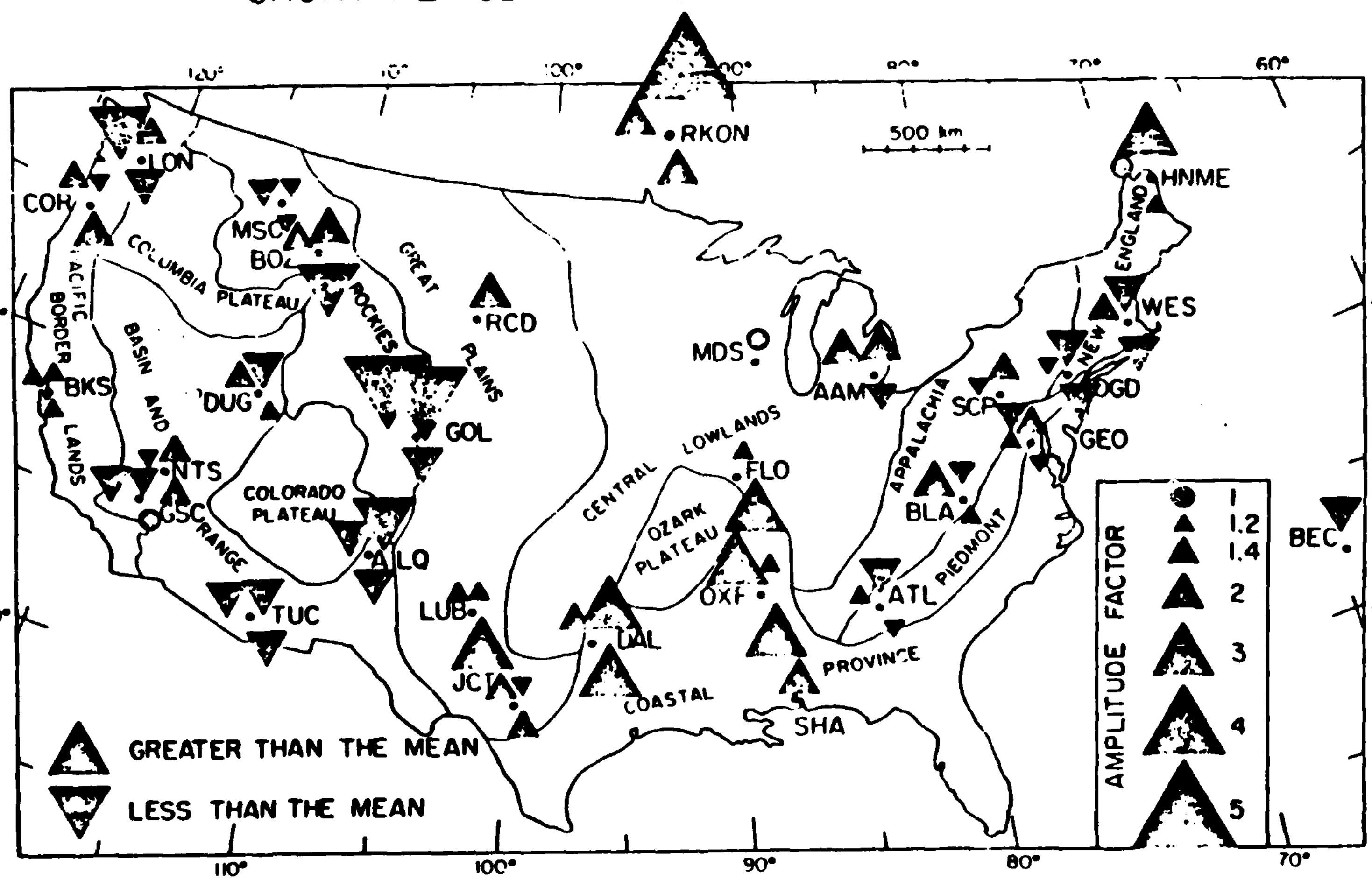

Figure 7 - Short period P-wave amplitude anomolies are plotted relative to station location and approximate physiographic provinces within the United States. Upward pointing triangles are values greater than the mean, downward pointing triangles are values less than the mean. A circle indicates a mean amplitude. The amplitude factor relative to the mean is noted at the lower right. The position of each triangle relative to each station, indicates azimuth: north, northwest and southeast.

for stations in the western U.S., would include large segments of the low Q region that Butler associates with the Rocky Mountain region. This would extend the apparent low Q region farther than if teleseismic data alone were used, due to the steeper takeoff angle of rays at teleseismic distances.

While m_b residuals appear, on the average, to correlate well with high frequency (.5-4 hz) δt^* for the U.S., several studies exist which cast some doubt on the applicability of this method to data in other regions and to specific sites within the U.S.

Von Seggern and Blandford (1977) have used regional δt^*'s calculated at LASA to attempt to reduce scatter in m_b-M_S. In highly attenuating regions, it would be expected that m_b would be reduced relative to M_S which, due to the longer periods involved, would be considerably less affected. However, correcting m_b values using δt^* derived corrections does not seem to reduce the scatter in m_b-M_S. Such an effect could, conceivably, be caused by systematic variations in source spectrum from region to region. An alternative explanation involves rapid changes in Q (f) near 1 hz. for a number of regions, with the location of the corner changing from region to region. If this is indeed the case, then the δt^* determined from spectral slope methods at high frequencies would not necessarily be a good predictor of the absolute attenuation level at 1 hz., which is what determines the m_b bias.

Der et al. (1979a) have recently completed a detailed study of relative m_b and high frequency δt^* for a number of sites at NTS, two additional sites in the southwestern U.S., and the SDCS stations RKON and HNME. Sources used were all located at distances of 30°-85°. Magnitudes were corrected for crustal amplification effects where applicable. The results of this study may be seen in Figure 8.

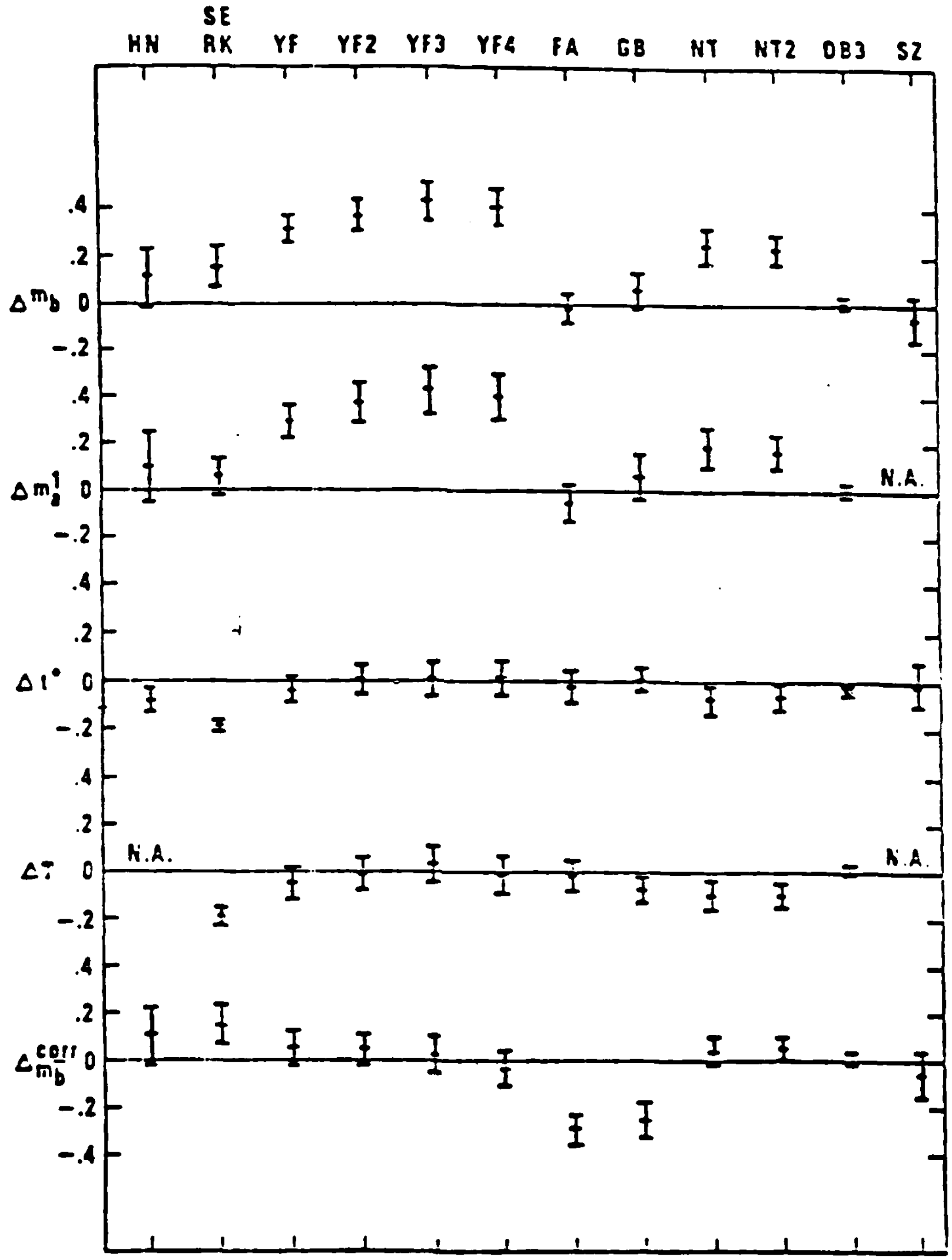

Figure 8 - Magnitude, δt^*, and period residuals for selected SDCS stations relative to station OB2. δm_b^{corr} is magnitude residual corrected for surface impedance. From Der, et al. (1979a).

We note that the δt^* difference between RKON and the southwestern SDCS stations is .15-.25 seconds and that the difference between HNME and the southwestern stations is 0.0-0.1 seconds. This would indicate an m_b differential of ~.3 between RKON and the southwastern sites. The actual corrected m_b for sites FANV and GBMN is in fact .3 lower than RKON. The NTS sites, however, are uniformly only .1-.15 m_b units below RKON. This indicates that magnitudes at NTS are not only large by .2 m_b units compared with the western average, but that, at least in this case, δt^* is once again not a valid predictor of magnitude variation. Thus, if one were to base yield estimates for events in shield areas on NTS results, a different answer would be obtained than if the correction were based on actual magnitude differences measured at the relevant test sites. As in the previous case, one explanation for these discrepancies is to have a different "corner" for Q (f) at the NTS sites and at RKON, thus effecting the absolute amplitudes at these stations but not necessarily the spectral slopes.

2.5 Source Effects. Our understanding of the seismic source function associated with an underground explosion is central to our abilities to estimate the magnitude or yield bias. The depth of burial and medium characteristics in the source region can strongly influence seismic radiation. The analytical representation, or source function, for an underground explosion is of great importance in separating attenuation from the source strength. This is of course the classic problem of seismology, the separation of source from path or propagation effects. Thus we must examine our understanding of the source in order to evaluate the estimation of potential yield bias.

The effects upon seismic radiation due to changes in depth of burial are intimately tied in with the properties of the material surrounding the

working point. Moreover, such effects are also yield dependent. These effects have been studied theoretically by Systems, Science, and Software and by Pacific-Sierra Research Corporation and Applied Theory, Inc. (e.g., Perl et al., 1979). One of the more significant results illustrates that coupling of the source to the surrounding medium and interaction with the free surface are very sensitive source depth. For example, calculations for wet sandstone show that m_b varies by .38 as the depth of burial is varied from 200 meters to 550 meters. However, if the m_b measure is replaced by an m_a magnitude (which utilizes the "b" phase amplitude), then for depths greater than 200 meters, the magnitude is essentially depth independent. (This phenomenon was noted qualitatively in previous reports. For example, Bache et al., 1976.)

The presence and influence of spall phases is closely related to the above discussion. A series of numerical investigations of spall phases have been conducted by Bache (report at the V.S.C. Research Conference, December 1978). These studies demonstrated that a spall phase could produce an enhancement of the second downswing of the short period P-wave seismogram, a phenomena which has often been observed. This can produce larger m_b values but would not bias an m_a measure. Studies have been conducted (Sobel, 1978) to explicitly identify spall phases by examining the variation of near-field records as a function of depth. They concluded that while spall phases were very evident from the shallow near-field recordings, it was absent on the deeper near-field records, indicating that it should not be an important contribution to teleseismic seismograms. Other near source environment parameters, such as coupling efficiency between the source and the surrounding medium, are beyond the scope of this current report.

III. CONCLUSIONS

In the preceding section we have discussed the significant phenomena or parameters effecting our abilities to estimate the yields of underground explosions. The sections on absolute and relative Q measurements made up the bulk of that discussion since we feel that Q or attenuation is the single most important contributing aspect to regional magnitude/yield biases and to our uncertainties in accurately estimating explosion yields. It is in this area that we feel significant future research should be pursued and in particular, emphasis should be placed upon studies of the frequency dependence of seismic attenuation in the frequency band of .1 to 5 hertz. The effécts of frequency dependence has probably led in large part to the discrepancy between the conclusions of different inventigators and to the resulting controversies on m_b and yield bias. This reserach program should be directed at determining the general nature of this frequency dependence, at resolving the local frequency dependence for specific source and receiver sites, and at developing better techniques to measure seismic Q and source yields, methods in which the frequency dependence is explicitly taken into account.

Despite the uncertainties in the data and gaps in our knowledge discussed above and resultant need for additional research, we can make a reasonable estimate of the magnitude or yield bias between various regions based upon current understanding. We feel strongly that, at this time, this estimation should be based upon studies of m_b variations rather than δt^* evaluations. Most δt^* studies have been conducted with spectral analysis techniques and, as such, are probably affected by intrinsic frequency dependence. As a result, the spectral techniques and time domain methods such as energy ratios between frequency bands do not provide data on absolute amplitudes. The m_b studies are, on the other hand, biased more toward the frequency band of 2 to .5 hertz

in which the effect of the frequency dependence of Q should be diminished. Moreover, since seismic yield estimates are currently based on m_b, we feel that the m_b or amplitude measures should in large part provide the basis for our evaluation.

At this time, we feel that the average magnitude/yield bias between the Soviet test sites and the Nevada Test Site is +.1 to +.15 magnitude units. We have been strongly influenced in reaching this conclusion by the studies of Der et al. (1979a) (see Figure 8 in section 2.4) on m_b and t* variations between the SDCS stations OB2NV, YFNV, HNME, and RKON, by the studies of Butler (Butler, 1979; Butler et al. 1979) on amplitude differences between the various Soviet test sites, by the reported geological similarity between regions near the station HNME and test sites inside the Soviet Union, and by the initial studies of the frequency dependence of seismic attenuation (e.g. - Hart et al., 1979). It is possible to increase this bias by a factor of 2 if either the observing stations or the underground explosion is located in a particularly anomalous site such as typified by the Faultless or Gasbuggy sites. Both of these tests were located in areas of extremely high heat flow, Faultless at the Battle Mountain High and Gasbuggy at the Rio Grande Rift. Seismic reflection studies in the Rio Grande Rift, for example, identify magma bodies at fairly shallow depths. We note, however, that the critical seismic stations now used in yield determinations are not located in such areas and also that none of the usual Soviet test sites appear to be in areas similar to either the Battle Mountain High or the Rio Grande Rift.

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THE CHANGING RESULTS ON ATTENUATION OF P WAVES

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INTRODUCTION

The topic of attenuation of seismic waves is difficult to review at the current time because many long established ideas are giving way in the face of new results. The values of the seismic quality factor, Q, have been revised substantially in recent gross earth models. There are now firm indications that Q varies significantly with frequency as well as depth in the earth. The premise that teleseismic body wave amplitudes will generally be systematically low in the western U.S. has been shown to be untrue for the specific case of Soviet nuclear tests. Apparently a much more detailed zoning scheme is required for meaningful results. It has been suggested that receiver site characteristics vary so rapidly that attempts to predict them by tectonic region should be abandoned altogether. These issues are far from resolved at the present time, and the only way to review them fairly is to give several key results substantiating each of the major viewpoints.

THE ABSOLUTE LEVELS OF ATTENUATION IN THE EARTH

When it became clear recently that the dispersion due to attenuation must be accounted for in modeling gross earth data a new series of comprehensive velocity and Q models were put forward. They were designed to be appropriate for 1 hertz data, and they were based on both normal mode and body wave data. All varieties of seismic observations of the earth were considered in constructing these models. The two best known are the SL models of Anderson and Hart (1978) and the QBS model of Sailor and Dziewonski (1978). In many ways, there are the standard Q models accepted by the community. Some researchers prefer to parameterize attenuation of body waves in terms of $t^* = T/Q$ where T is the travel time and Q is the average Q along the raypath. The values of t^* for SL8 are tabulated by Hart and Anderson, and they are very close to 1 for P waves (4 for S waves) at telesemis distances. The Sailor and Dziewonski t^* values are comparable though slightly higher. The classic studies which initially showed that t^* at 1 hertz was about 1 were those of Carpenter and Flinn (1965), Carpenter (1967) and Teng (1968). More recent investigations which give similar results include the ScS-ScS_n studies of Jordan and Sipkin (1977)the P-PP studies of Butler (personal communication) and the relative attenuation studies of Burdick (1977). In each of these investigations the uncertainties of source excitation have been eliminated by experimental design. An alternative approach is to assume a model for the source and to apply corrections based on the model. Studies of this type include the work of Sipkin and Jordan (1979), Lundquist (1977) and Der and McElfresh (1976). These studies all tend to give much lower values for t^* above 1 hertz which

is one of the major points of controversy. However, some of the discrepancy could be caused by the inaccuracy of the source models used. Some of the earthquake sources were simple single corner frequency formulations. Others did not contain any free surface corrections at all even though the sources were very shallow.

FREQUENCY DEPENDENCE OF Q

An alternative way of reconciling the discrepant measurements of t* is to turn to frequency dependent Q models. Such models have been discussed in the past by Gutenburg (1958), Solomon (1972), Anderson and Minster (1979) and Sipkin and Jordan (1979) among others. These authors have all recognized that, for physical reasons, one would expect attenuation to have a frequency dependence related to the mechanism responsible for it. Futterman (1962) recognized that Q must depend on frequency to even preserve causality. Important theoretical contributions on propagation of seismic waves in media with frequency dependent attenuation have been made by Liu *et al*. (1976), Minster (1978a,b) and Chin (1980). A very useful mathematical model for attenuating solids discussed by all these authors is the standard linear solid. Simple expressions for the Q operator for this medium are easily derived. Some time ago Attewell and Ramana (1966) presented evidence that Q was not dependent on frequency over a very broad range. Der and McElfresh (1976, 1977) found no evidence in spectral ratio data that Q varies in the high frequency range of .5 to 5 hertz though they did suggest that it varies at longer periods. The more recent study of Sipkin and Jordan (1979) supports the opposite viewpoint. These latter contributors suggest on the basis of ScS data that Q begins to

increase dramatically with frequency above one hertz. If true, it could explain why 5 to 10 hertz energy is observed so commonly.

REGIONAL VARIATIONS IN ATTENUATION

One of the long standing patterns of regional attenuation which is of particular importance to the magnitude-yield problem is the east-west trend in the U.S. Evernden and Clark (1970) and Booth, Marchall and Young (1974) concluded that amplitudes measured at stations west of the Rocky Mountains front were about 3 times smaller (.5 magnitude units) than those in the east. Similar claims were made by Der, Masse and Gurski (1975) and Solomon and Toksoz (1970). Butler (1979a,b) found that this was not generlly true for the combined data for all of the Soviet nuclear tests. Some amplitude patterns which do exist are related to the source rather than received. The whole concept of attributing amplitude variations to receiver characteristics has been brought into question. Butler worked exclusively with explosion data, so he circumvented the difficulties with radiation pattern encountered by previous researchers. Obviously his data set is also most directly related to the yield determination problem. Der et al. (1979) revised the estimate of the magnitude differential caused by attenuation down to only .3 magnitude units. Burdick (1978) could not resolve any higher S wave attenuation in the western U.S. though his resolution of differences in t* for P was only about ±.2 seconds. In recent analyses of SDAC data it has been found that eastern U.S. station HNME is only .1 magnitude unit higher than the NTS station OB2NV in the west. Station RKON is again only .15 units higher than the test site station. A .1 differential in magnitude units corresponds to only a

.1 second differential in t^*. The overall effect is certainly much smaller than the original estimate of .5 magnitude units. It is possible at this point to more finely divide the U.S. into regions, but we are rapidly approaching the point where we have as many regions as stations. Herrin pointed out in a presentation at the 1979 AFTAC review that regional corrections may simply not be practical in the yield determination problem. Source corrections for tests inside the Soviet Union cannot be estimated without seismic data from each new test site.

AREAS FOR FUTURE RESEARCH

Some of the major questions about the nature of P wave attenuation in the earth have obviously not been answered at this point. Directions for future researach are on the other hand fairly clear. More measurements of the absolute value of t^* are needed at all frequencies. Emphasis should be placed on those experiments in which source uncertainties can be eliminated such as P-PP or ScS-ScS_n studies. Attempts must be made to detect the frequency dependence of Q directly rather than by comparing the results of one type of constant Q experiment to another in a different frequency band. The importance of scattering as a mechanism for attenuation and its frequency dependence must be established. Attempts to relate amplitude variations to tectonic regions should perhaps be continued at some level. It must be recognized, however, that most previous attempts to do so have been too simplistic. Future research must take into account the fact that major changes in signal amplitude and spectrum at a station are related to many factors besides inelastic attenuation under the station.

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STATE OF THE ART ASSESSMENT: REGIONAL ATTENUATION EFFECTS ON P WAVES

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The totality of geophysical evidence to date shows that large lateral variations exist in the properties of the Earth's upper mantle. The laterally inhomogeneous parameters include seismic velocities, temperature as derived from heat flow and conductivity measurements, chemical composition and anelasticity. Regions of the upper mantle under tectonic regions, rift zones, mid ocean ridges and some regions behind island arcs are characterized by low upper mantle seismic velocities, high heat flow and high attenuation of seismic waves. The Basin and Range province of the western United States also exhibits all these characteristics. On the other hand, the upper mantle under shields (and possibly under old ocean basins) is characterized by high velocities, lower temperatures and low anelastic attenuation (11,12). 21

The anelastic attenuation parameters Q_α or t^*_α are also frequency dependent. The efficient propagation of high frequency P energy (up to 5 Hz) over great distances through the Earth and the relatively low Q values found at lower frequencies ($f < .1$ Hz) makes the conclusion that Q_α is frequency dependent inevitable (1,3,7,18,22).

Besides the regional variation of Q_α the gross mechanism of attenuation is also of interest. Most available evidence indicates that losses occur in shear deformation and that $t^*_s \sim 4t^*_p$, implying that the attenuation effect is more severe on S waves (8, 25). There are some reports that some losses also may occur in compressional deformation, but the proportion of losses in compression is relatively small (21).

Since the body wave magnitude m_b, a parameter important in detection and discrimination of earthquakes and explosions, is determined in the .5 - 5 Hz band (mostly in the .5-2 Hz band) the knowledge of regional variations of upper mantle Q_α is very important.

Measuring attenuation in the short period band in the mantle under a given station is complicated by many factors. Anelastic attenuation introduces a first order, drastic reduction of high frequency content of seismograms. These changes

in frequency content are easily measurable in the fre˜ ‹ency domain. It has been shown that other disturbing factors on shapes of ˆ spectra are of second order, much less important in size when compared to the effect f t^*. Such disturbing factors are the local crustal amplification, local focusing, multipathing, and surface reflections above the source, to name a few. In order to separate the effect of t^* from other effects, it is necessary to use both spectral and amplitude information as well as other geophysical diagnostics. The evaluation of the degree of anelastic attenuation should be based on several parameters, such as

a) Relative P magnitude anomalies (corrected for estimated crustal amplification and any radiation patterns)

b) P wave spectra shapes

c) Relative short period S wave amplitude anomalies (corrected for crustal amplification and any radiation patterns)

d) S wave spectra shapes

e) Other geophysical diagnostic parameters such as:

  Travel time delays (P and S)
  Heat flow (unreliable)
  Mantle conductivity

In our view a and c by themselves are not suitable for measurement of attenuation under a given site, due to the effects of azimuthally dependent local focusing, imperfectly known local structures, topography and other factors on absolute amplitudes. It has been shown that such effects can cause perturbations in magnitudes of several tenths of magnitude units. Besides, there is no simple way by which t^* can be related to time domain amplitudes. Studies of P amplitudes and spectra at arrays showed that while amplitudes are subject to locally and azimuthally varying systematic biases, spectral shapes are remarkably stable (19). It appears, therefore, that criteria b and d should even be given more weight than a and c.

A region of special interest for P wave attenuation studies is the contiguous United States. Since most available magnitude-yield information comes from this region, it is important to know how lateral variations of attenuation may effect any conclusions drawn from the magnitude-yield data. It is certain that broad regional variations of attenuation exist under the United States. The most

contrasting areas are the Basin and Range province and the north central United States that is essentially a geophysical shield (2,5,6,7,8,9,12,17). All five diagnostic criteria listed in the previous paragraph are satisfied to conclude that the mantle under the Basin and Range attenuates all traversing seismic waves more severely than under the shield areas of the United States (2,3,5,6,7,8,9, 10,13,14,15,16,17,20,23,24,25). Besides these extremely contrasting regions one can discern other, finer, regional variations in the attenuative properties of the mantle under the United States. The northeastern United States is characterized by somewhat higher attenuation than the shield, the mantle under some parts of the western United States outside Basin and Range probably is less attenuating than under the Basin and Range proper (9,4,25). There are also some indications of a high Q strip in the mantle along the Pacific Coast. All these details require further study.

An important question for yield verification is the position of NTS within this general picture. Our studies show that criteria a and b are satisfied and that P wave amplitudes and spectral ratios at RKON (a shield station) and NTS differ in a statistically significant (99% conf)level. The available data shows a significant decrease of high frequencies in S waves at NTS relative to RKON. (Criterion d). The available S wave data is not sufficient to test criterion c. Comparison with other WUS stations, the sites of FAULTLESS, GASBUGGY, RIO BLANCO and RULISON explosions showed that the spectral content in P waves at all these locations is deficient in high frequency energy relative to RKON (on the 95% confidence level).

P travel times at NTS are about 1.5 second late relative to RKON after correcting for elevation difference (Part of criterion e). All these data indicate that the mantle under NTS is highly attenuating. The RKON-NTS t_p^* differential around 1 Hz is estimated to be about .19$\pm$.02 sec. (95% confidence limits). Data from two stations in the northeastern United States (HNME and IFME) show that they occupy an intermediate position between NTS and RKON with respect to t_p^*. This agrees well with the general regional picture presented above.

There is a need for much more research to understand the precise manner in which Q_κ depends on frequency, depth and attenuation mechanism in the various regions of the Earth. The results of such research will help us to better understand the dynamics of tectonic processes and the development of the oceans and continents.

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State of the Art
Regional Attenuation Effects with
Special References to the Nevada Test Site

Robert R. Blandford
Teledyne Geotech

Summary

I believe that the evidence is overwhelming that some form of attenuation exists in the earth and varies to such a degree that m_b can be affected by up to 0.5 magnitude units or possibly more, and that the spectral ratio between 1 and 4 Hz can vary by a factor of 100 or more. With respect to an explosion detonated in the Canadian Shield or similar region as compared to that detonated in the Basin and Range; I would expect on the average a difference in m_b of about 0.4 m_b. For the special case of NTS as compared to a typical shield station there is rather good evidence that the difference is only 0.2 m_b, probably because of favorable effects of local focussing due to upper mantle structure. I know of no significant counter-examples or even general geophysical ideas which are inconsistant with this picture, and in each case that I have had a hand in examining the data personally the results were not in conflict with the above remarks. In what follows I shall try to present various data and summaries of others' work all of which seem to me to support this point of view.

Supporting Regional Geophysical Studies

The Basin and Range (BR) and the rest of the Western United States (WUS) are clearly different from the Eastern United States and Canadian Shield (termed EUS hereafter) in that the elevation is much higher and it is more seismically active. It is generally known of course that heat flow is higher and that electrical and magnetic measurements reveal a layer of high conductivity at shallow depths in the WUS. At the 1979 Fall AGU M. McNutt showed that, isostatically speaking the EUS responds as an elastic plate whereas the WUS responds a series of disconnected blocks. Pn, Pg, and Lg propagate in completely different ways in the WUS and EUS e. g. amplitudes decay much more rapidly in the WUS. Studies which show differences in Pn velocity and in travel time residuals from teleseismic events, with slower velocities in the WUS are well known. (Recently a study by Gibowicz (1970) of the 1964 Alaskan aftershocks has been cited as a counter-example since these events have early arrivals in the WUS. This is clearly not a convincing counter-example because of plate effects and of the absence of averaging over azimuths).

Perhaps the most convincing study from an overall geophysical point of view is that of Molnar and Oliver (1969). They used the propagation efficiency of the phase Sn, which propagates in the upper mantle, to elucidate the tectonic regions of the earth. Sn does not propagate across spreading zones where new plate is being created, nor behind island arcs where volcanos exist, see Figure 1. This suggests that partial melt was absorbing these high frequency waves. The phase Sn also does not propagate in the Basin and Range, see Figure 2 for a detailed picture and discussion of the United States and Canadian Shield. Note the exception in the WUS of good propagation just along the coast.

Direct Studies of P-wave Amplitude

Evernden and Clark (1970) and Booth, Marshall and Young (1974) computed

magnitude residuals using earthquakes measured for LRSM bulletins at LRSM stations. A map of the results of Booth et. al. is given as Figure 3 where we see a general picture of low amplitudes in the WUS and high amplitudes in the EUS. A notable exception is station KNUT with a zero residual in the WUS. Exceptions of this sort due to focussing and defocussing are to be expected. North (1977) performed a similar analysis using ISC readings as measured at WWSSN stations, in Figure 4 we see his results. The overall picture is the same as for Booth et. al.; low amplitudes in the WUS, higher in the EUS. A notable exception is the station BKS on the California Coast which has a high amplitude; this correlates with the good propagation from Baja to BKS as seen by Molnar and Oliver (1969). See Figure 2.

Butler (1979) has also analyzed the WWSSN stations in the United States using only explosions from the USSR, and earthquakes selected to be simple from the Northeast and Southeast. His results are very similar to those of North (1977) in general; and for the stations they have in common, BKS, GOL, ALQ, TUC, FLO, and ATL the relative amplitudes are very similar; the only exception being ATL. In the northeast Butler finds a confused picture suggesting moderate absorption. This is in agreement with numerous geophysical studies and in slight disagreement with the study of Booth et. al. (1974). However, this disagreement probably simply reflects a different sample out of an intermediate population.

Taking the overall picture, stations in the Southwestern United States have low amplitudes for P waves as compared to shields and the EUS in general.

Der (1977) investigated the effect that plane-parallel crustal structure might have on these amplitudes. Analyzing the data of Booth et. al. (1974) he found that crustal effects were very significant, and that removing the effects revealed more clearly the difference between the WUS and EUS; much of the scatter which seems to exist within each region on a map would be reduced were crustal corrections applied. Figure 6 shows the log of the expected crustal amplification plotted versus the Booth et. al. magnitude residual. We see that the separation between regions is enhanced by these corrections. For example, when its large crustal effect is taken into account, station KNUT is not so anomalous.

Also in Figure 6 we see a narrow ellipse denoted by NTS. The location of this region is a result of analysis of SDCS data, and work by Der and others places stations on granite at Climax stock and Gold Meadows stock, and stations at Yucca and Pahute Mesa along this ellipse. There is general agreement between workers at Geotech and at Sierra Geophysics that the average separation between OB2NV and RKON is as pictured on this Figure, approximately 0.2 magnitude units; and that, if corrected for crustal response all the stations at NTS would have the same amplitude.

Thus we arrive at the conclusion that NTS is not so anomalous with respect to shield stations as it might have turned out to be had it been more typical of the WUS. It's anomaly is only 0.2 magnitude units, and by a reciprocity theorem an explosion detonated at OB2NV would be expected to have a magnitude only 0.2 units less than that of a shot detonated at RKON. After crustal correction all other stations occupied outside of NTS in the WUS as part of the SDCS project exhibit anomalies more typical of the WUS, e. g. stations at the site of Faultless, Gasbuggy, and Rulison. Analysis of a station at the Shoal site yielded a similar conclusion; typical of the WUS. Only NTS is the exception.

Variations of 0.2 magnitude units and greater have been observed over a space of 50 km at the LASA and NORSAR arrays, see for example Chang and von Seggern (1979), (that paper discusses an appropriate reciprocity theorem) and Blandford (1974). These variations have been traced by most workers in the field to focussing and defocussing due to variations in the velocity structure below the Moho.

Supporting Studies of Amplitude and Spectra, Including S Waves

Figure 7 from data in Der et. al. (1978) shows the only explosion recorded both at OB2NV and RKON for which OB2NV is not in or near the shadow zone. We see that the OB2NV signal is much lower in frequency and has an overall trace amplitude about 0.5 magnitude units less than that at RKON. Due to the correcting effect of the A/T processing this results in a magnitude difference of only 0.36 m_b. The effect is even more dramatic in the spectra; at low frequencies the ratio seems to be trending to a 1:1 ratio, whereas at 4 Hz the ratio is nearly 1:100. The difference in t* implied by the average slope is 0.26 units; but there is some evidence for a decrease in the difference in t* from perhaps 0.5 below 2 Hz to 0.1 above 2 Hz. At HNME for this event the spectral ratio was flat ($\Delta t^* = 0$) and the amplitude twice that at OB2NV.

Figure 8 from von Seggern (1977) shows the signals at a common single instrument at EKA of signals from Milrow and Boxcar, two events very closely matched in medium, yield, and depth. Even in the time domain we see longer periods and less high frequency for the event from NTS. EKA is one of the few digital stations at teleseismic distance from both events. For these two events von Seggern determined a difference in m_b of 0.3 m_b averaged over a common network of 23 well distributed teleseismic stations.

Figure 9, taken from plots in the report by Marshall (1972) attempts to compare signals from Piledriver, a shot in granite, to shots in granite from Semipalatinsk. To find shots of equal yield, in light of the foregoing one would like to have an m_b 5.5 shot at Semipalatinsk to compare to the 5.23 Piledriver event. One can see from the difference between the Semipalatinsk signals that careful yield matching is necessary to avoid the frequency effects of cube-root-scaling. Even so, it is apparent, that the two smaller Semipalatinsk events have much higher frequency than Piledriver and that the larger Semipalatinsk event shows slightly higher frequency in the main pulse and much higher frequency in the coda. Some years ago I compared Piledriver to well-selected Semipalatinsk events at NPNT and observed dramatic differences of the appropriate sort in the spectra. Unfortunately I have lost these results. Studies of this type comparing Piledriver to Semipalatinsk should be begun immediately at such stations as EKA, NPNT, and possibly other stations if workers can be confident that data from the same single instrument is always used.

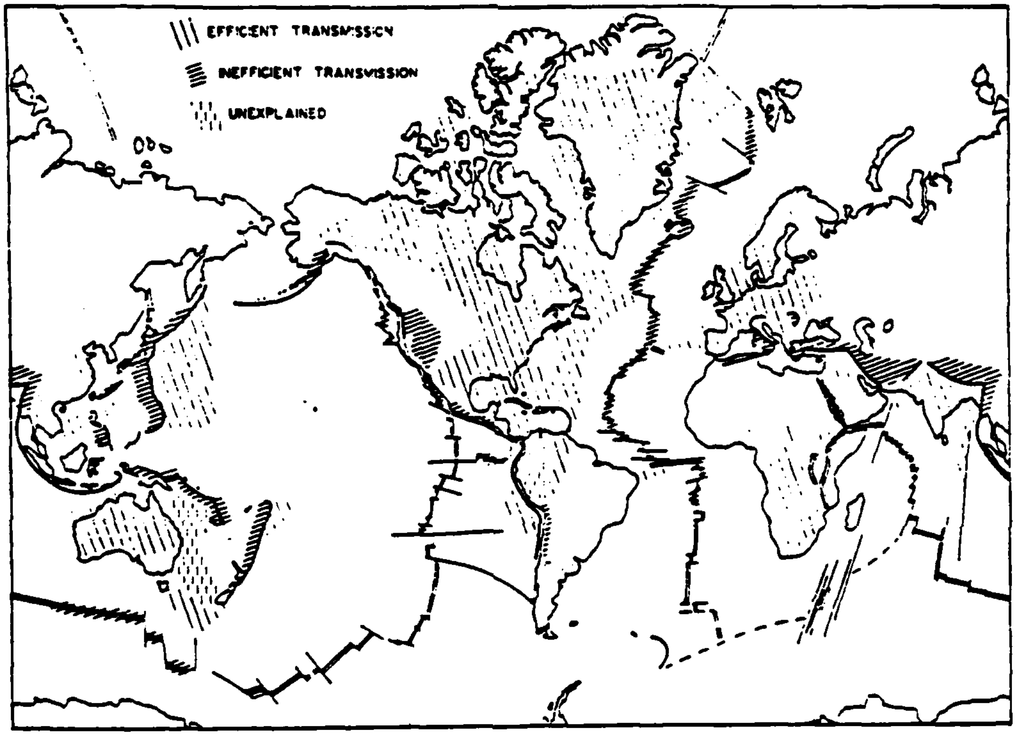
Figure 10 from Noponen (1975) shows the spectral ratio at NORSAR of four NTS events with mean $m_b = 5.2$, to six E. Kazakh events with mean $m_b = 5.5$; just the right ratio to be of approximately equal yield. We see again that the ratio falls off dramatically with frequency. This result could, however, be criticized on the basis that the NTS shots were not in granite.

Figure 11 shows the only short-period shear wave so far discovered to be recorded at both RKON and OB2NV. We show the radial component, the transverse is quite similar. The spectra are quite different and the implied difference in t* is about 4 times the t* difference for P, as is appropriate.

Figure 12 shows short-period shear wave transverse data from an earlier earthquake when the full LRSM network was installed, note the high frequency and amplitudes (note gains) in the EUS, and low frequencies and amplitudes in the West, especially in the Southwest.

This event and four others were corrected for effects of a double couple radiation pattern which was in accordance with published focal mechanisms for each event, and the corrected amplitudes are plotted in Figure 13; again we see the low amplitudes in the WUS and higher amplitudes in the EUS.

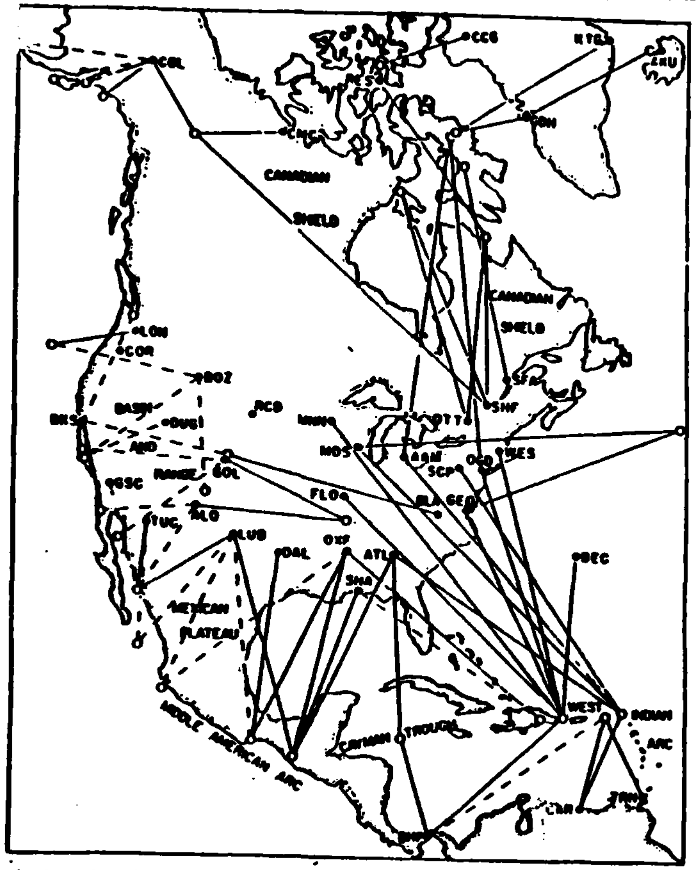
All of the above results represent an attempt on my part to present an unbiased cross-section of available data. I have never seen personally any data which seemed to me to point to conclusions opposite to those in the introductory summary. I have been told that plots of network m_b versus yield exist which point to the opposite conclusion, however I am aware of many subtle difficulties in determining network magnitudes, and could not endorse such conclusions unless I could read a carefully written manuscript with its supporting data.



Summary of regions where S_n propagates efficiently. Island-arc structures are represented by bold dark lines; crests of mid-ocean ridge, by double lines; and fracture zones, by single lines (adapted from *Isacks et al.* [1968]). A summary of the data and its interpretation are presented. Regions with efficient transmission are largely stable regions (shields or deep ocean basins) and cover areas traversed by paths that transmit strong S_n phases. In general, paths crossing the concave side of island arcs or crests of the mid-ocean ridge do not transmit S_n. In some regions the data limit attenuation to a relatively narrow zone, but in others the data merely conform with the pattern of inefficient transmission across these features. Some regions for which the data are ambiguous or inconsistent are discussed elsewhere in this paper.

Molnar and Oliver (1969)

(1)



North America. Efficient transmission is indicated by solid lines; inefficient transmission, by dashed lines. Parts of this region are illustrated in Figures 9, 10, and 11.

Molnar and Oliver (1969)

(2)

Discussion of Figure 8. Best examples of high-frequency S_n phases studied propagate from earthquakes in West Indies to North America. This phenomenon is well documented [*Shurbet*, 1962; *Linehan*, 1940], and S_n phases with frequencies higher than 5 cps have been observed for similar paths [*Isacks and Oliver*, 1964]. Propagation of S_n to stations as far as 35° is very efficient, and no major effect of continental margin or root of Appalachian Mountains on transmission of S_n was observed. One exception exists; S_n is rarely recorded at SHA for West Indies events, though well recorded there for many events in Mexico. The best example of S_n at SHA was for the event closest to SHA. Hence, probably most of these weak S_n phases result from high attenuation near the West Indies themselves. Also, excellent examples of P_n phases were observed for paths to North America [*Shurbet*, 1962]. P_n phases probably are often confused with pP and give rise to erroneous depths of focus. Moreover, for very small events P_n probably is confused with P, because of higher attenuation of P than of P_n.

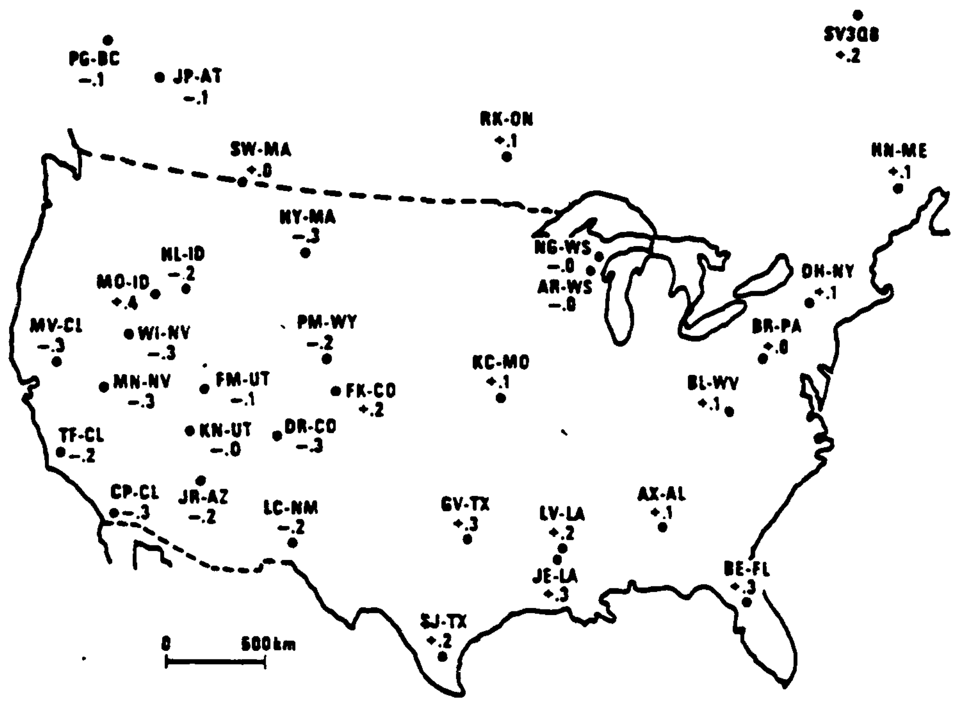
Propagation of S_n is efficient under Canadian shield. Owing to lack of appropriate earthquakes and WWSSN stations, data were supplemented by data from Canadian seismograph network, which is of WWSSN character. Many paths studied are very long; in fact, one between Baffin Island and ATL is 39° in length, the longest path for which S_n was observed in this study. *Brune and Dorman* [1963] noted several S_n phases at distances as great as 4500 km. At such distances S_n is not strong enough to be given a quality factor of 2 or 3, but it still attests to efficient propagation.

Paths parallel to and east of Alaskan Peninsula transmit S_n efficiently to COL, but paths crossing concave side of this continental extension of Aleutian arc do not.

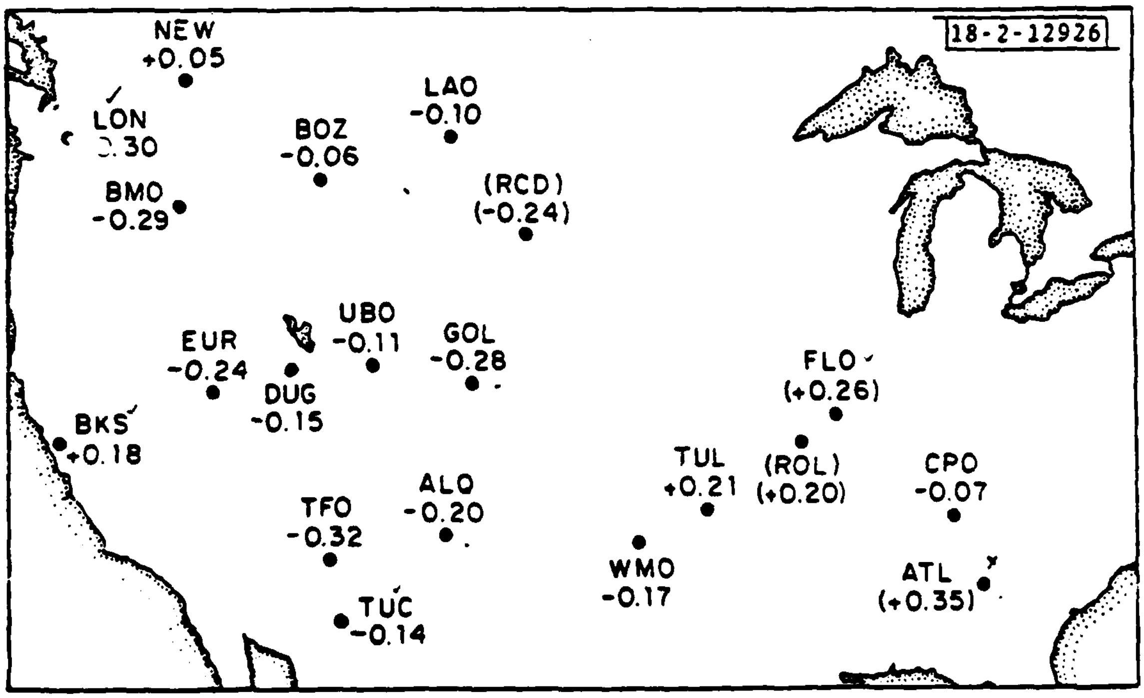
Some authors suggest that the East Pacific rise extends into North America beneath Colorado plateau and Basin and Range provinces [*Cook*, 1966; *Hamilton and Myers*, 1966]. Vine [1966] suggested that the East Pacific rise once underlay this region but was subsequently displaced westward. S_n is not observed for paths that cross this region. Data are inadequate to resolve pattern of attenuation more clearly because overlapping paths do not define a narrow zone of high attenuation and none are short.

Region	Quality Factor 3	2	1	0
Aleutians to COL (College, Alaska)	8	2	2	4
Canadian shield and the Arctic	4	18	4*	11*
Basin and Range	0	0	2	22
North Atlantic basin and eastern United States	42	34	6	3
Middle American arc to North America	4	19	8	43
Caribbean	17	2	2	8
Concave side of Lesser Antilles	6	1	1	3
Across the Cascades	0	0	0	9

* Most of these paths are very long (longer than 30°).



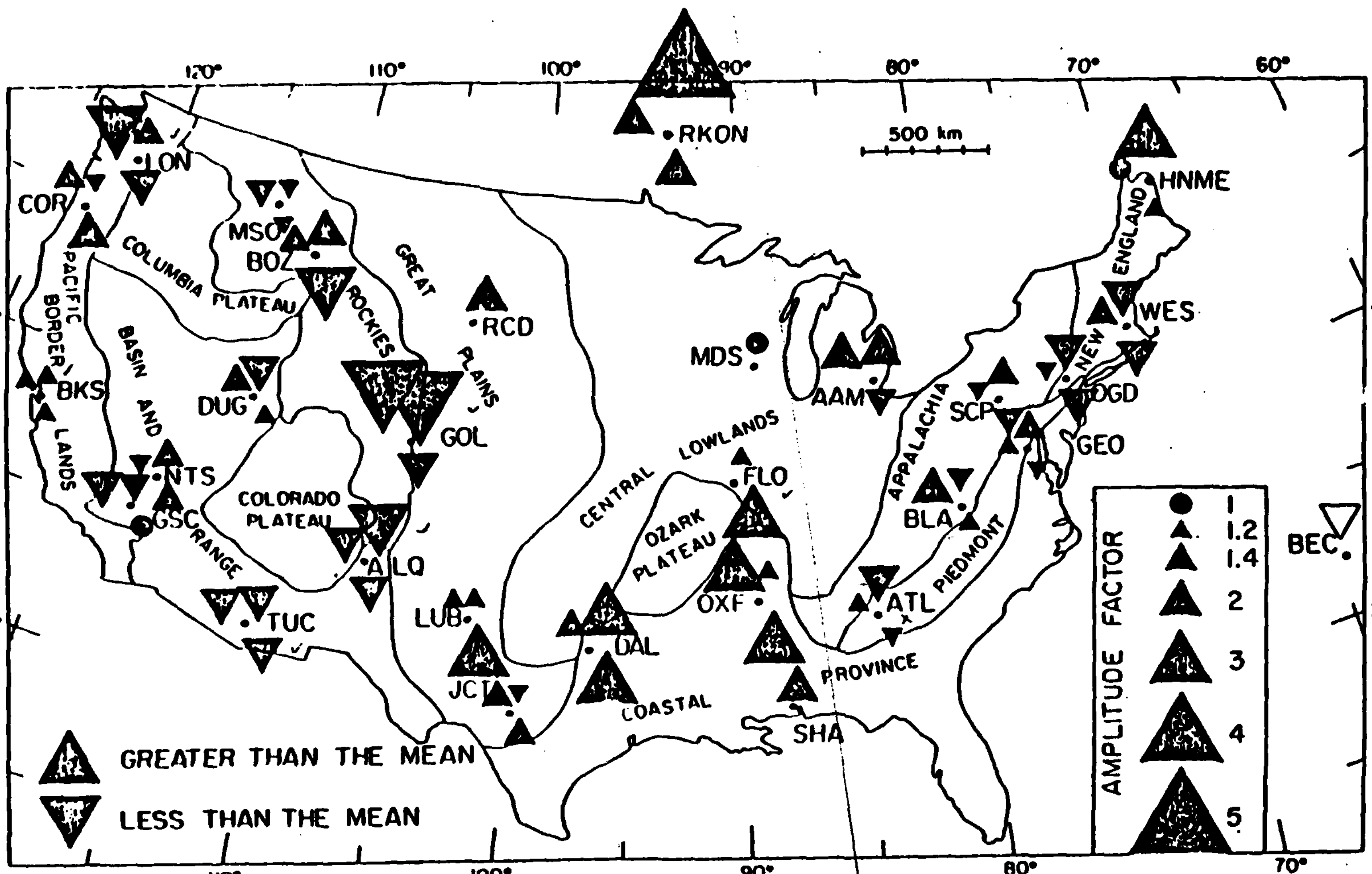
Short-period magnitude residuals based on the LRSM bulletin in the United States compiled by Booth *et al.* (1974).



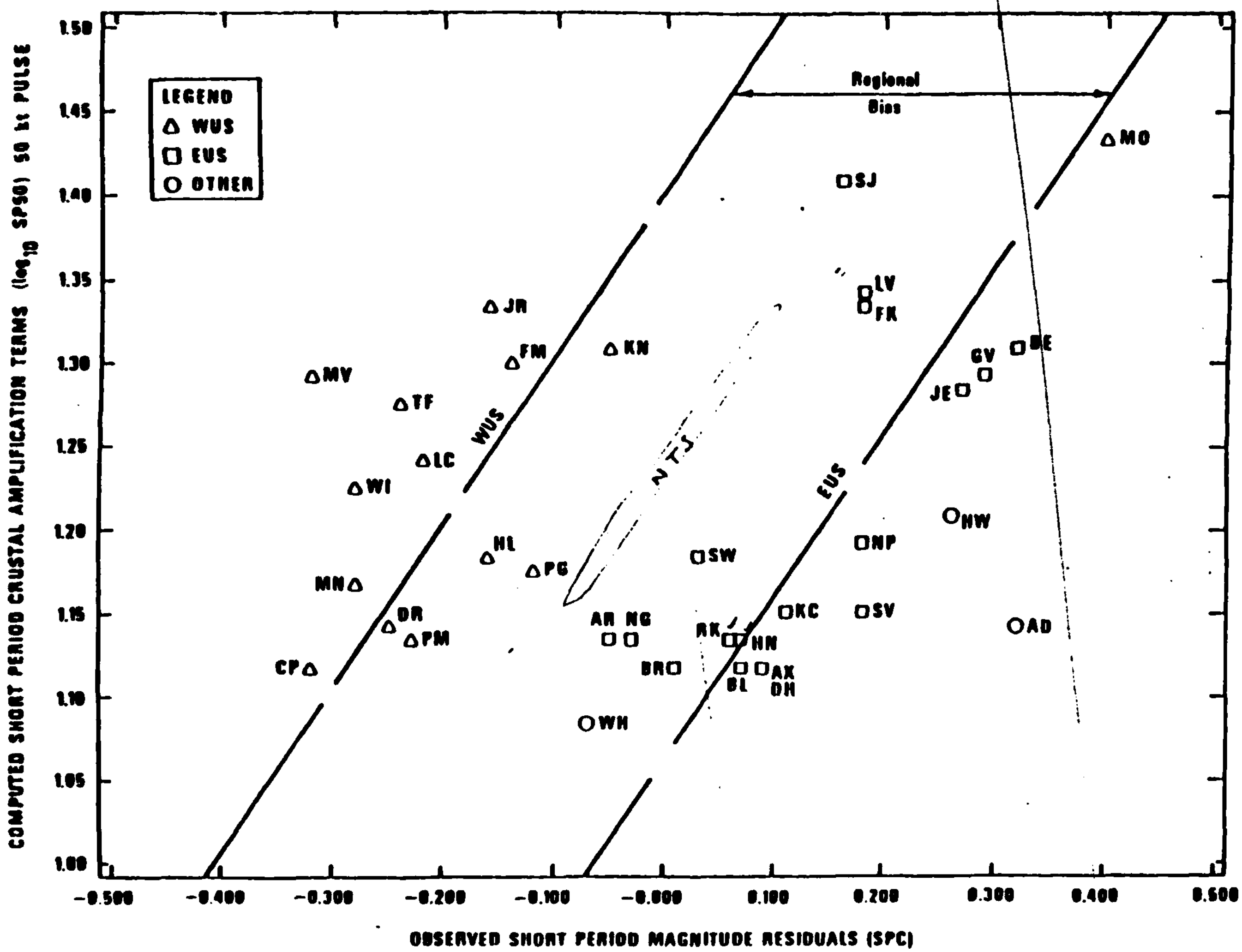
Mean biases for stations in the continental USA. Values in parentheses are from Table IV; all others are from Table II.

R. North (1977)

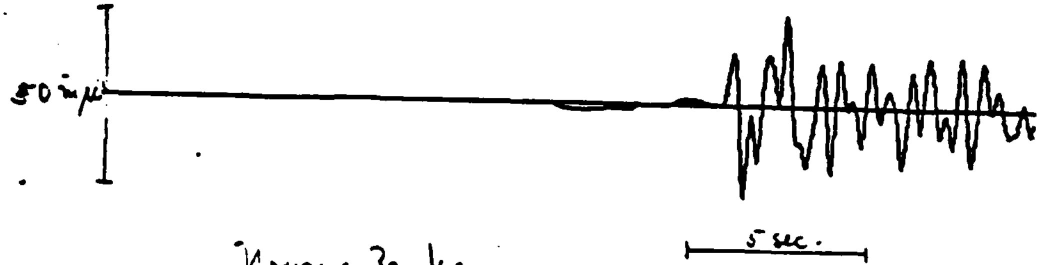
SHORT PERIOD P WAVE AMPLITUDE ANOMALIES



R. Butler (1979)



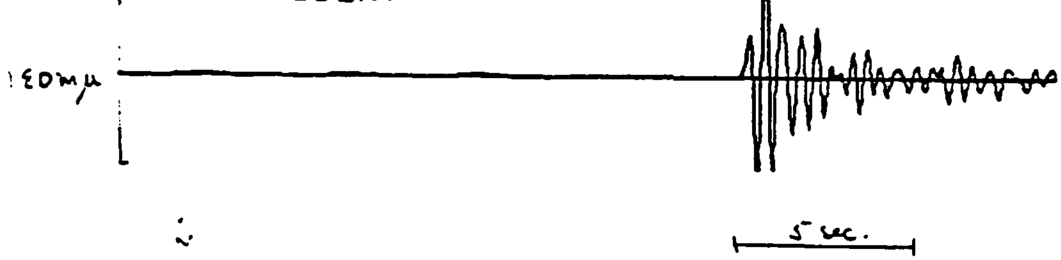
Short period crustal amplification term for a pulse for a 50 kt nuclear explosion plotted against short period magnitude residuals of Booth et al (1974). Regression lines for the EUS and WUS populations are drawn separately. Regional magnitude bias is shown on the top of the figure.



Novaya Zemlya

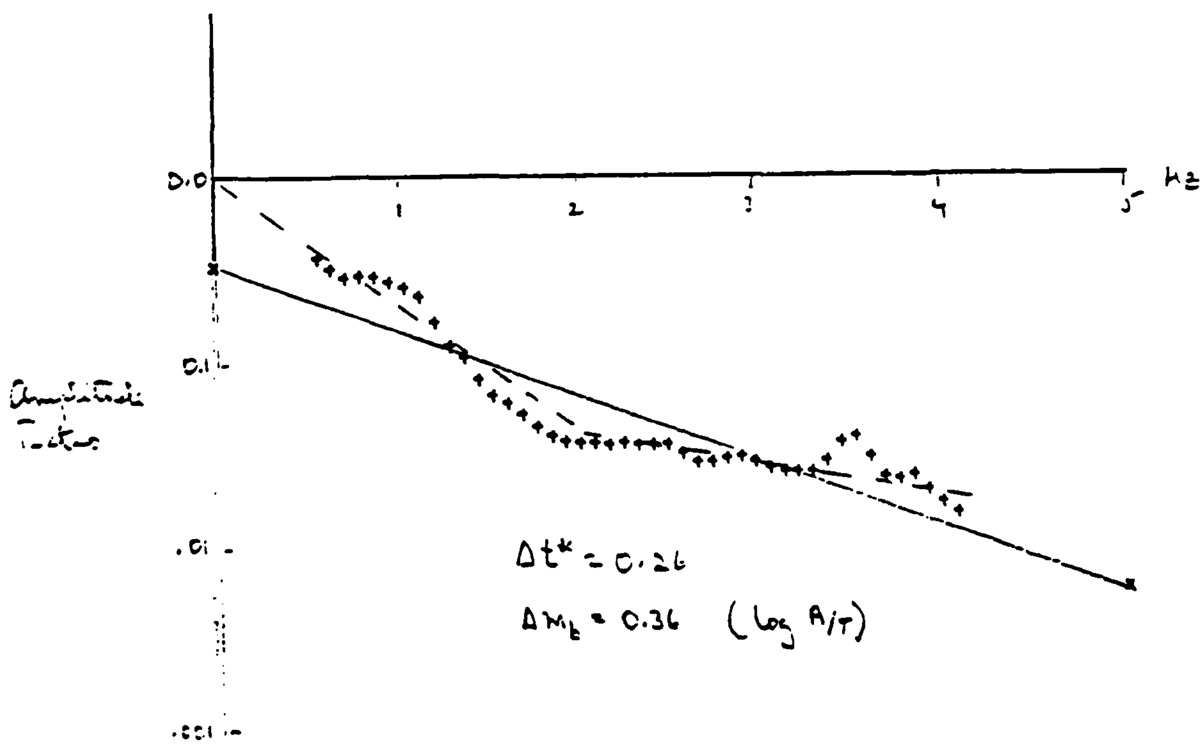
29 SEP 76

OB2NV



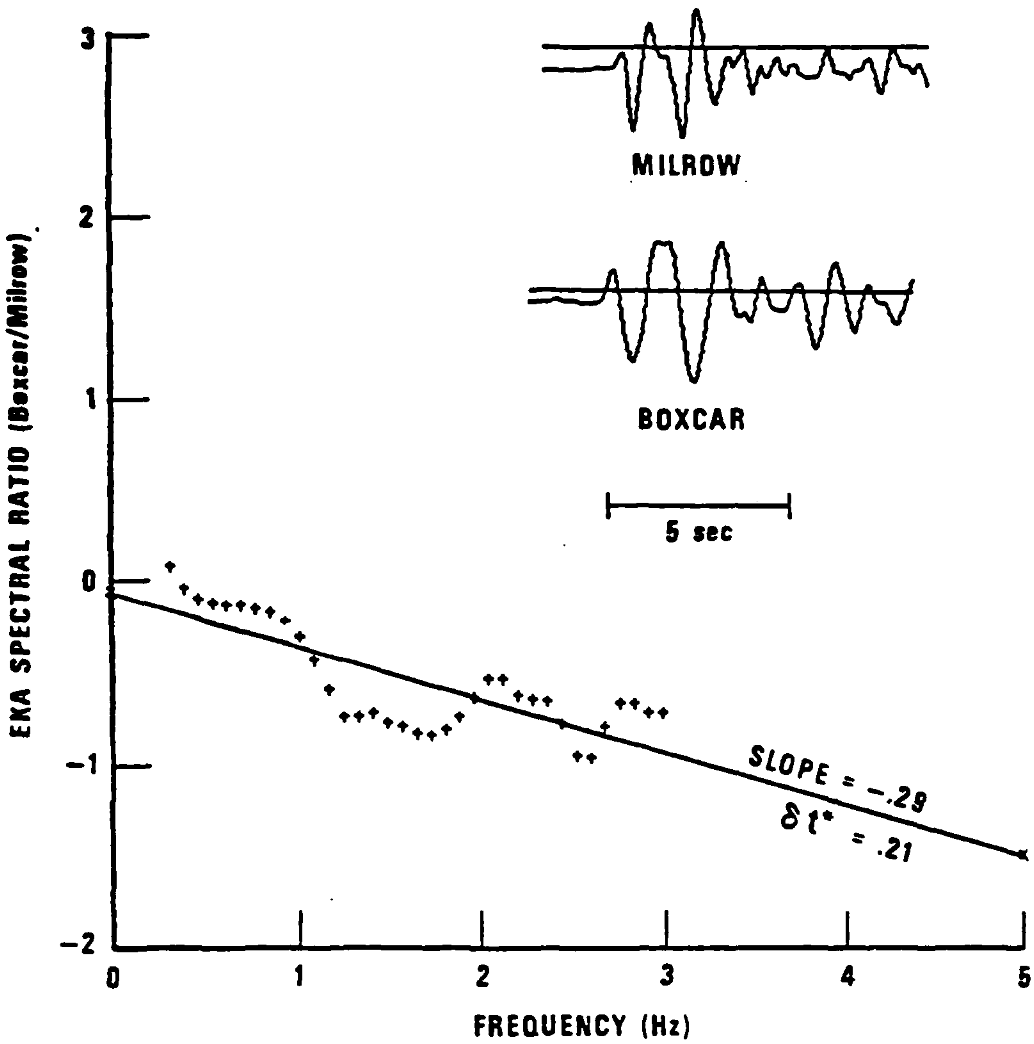
29 SEP 76

RK-ON

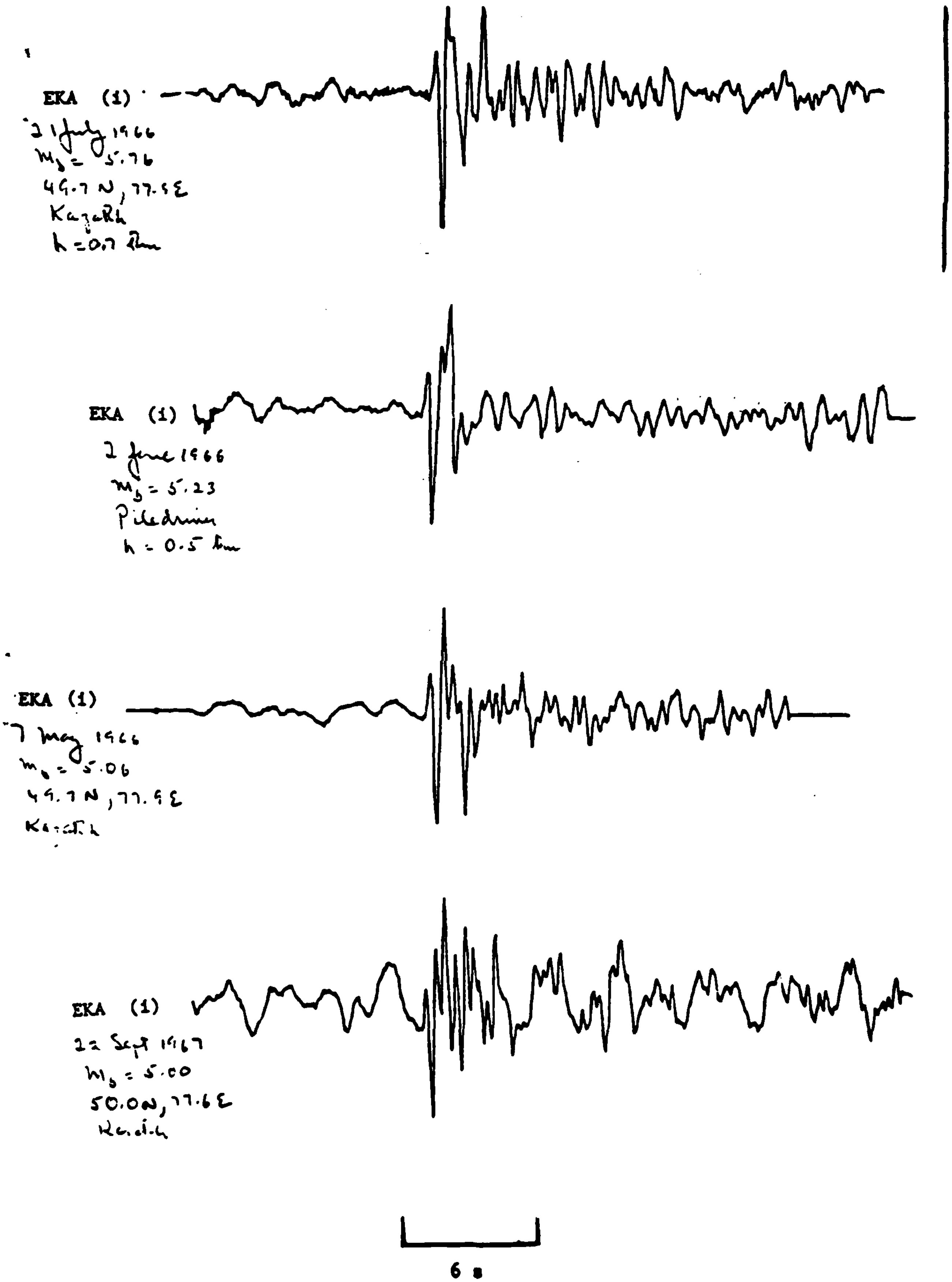


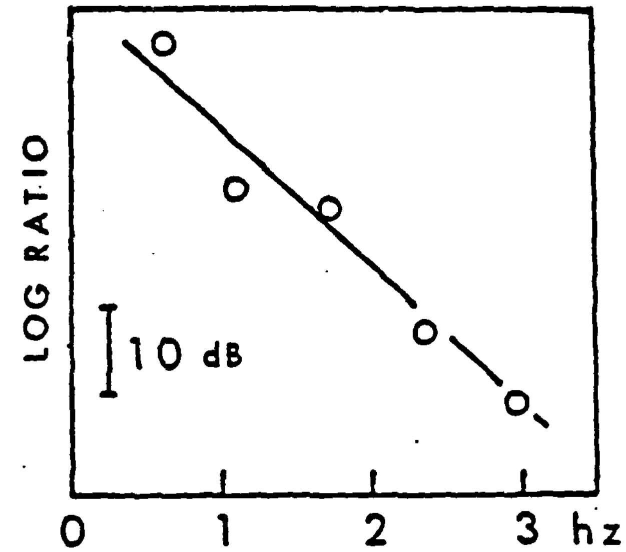
OB2NV

RK-ON



Spectra for MILROW and BOXCAR P-waves recorded at EKA.


EKA (1)
21 July 1966
m_b = 5.76
49.7 N, 77.9 E
Kazakh
h = 0.7 km
EKA (1)
2 June 1966
m_b = 5.23
Piledriver
h = 0.5 km
EKA (1)
7 May 1966
m_b = 5.06
49.7 N, 77.9 E
Kazakh
EKA (1)
22 Sept 1967
m_b = 5.00
50.0 N, 77.6 E
Kazakh
6 s

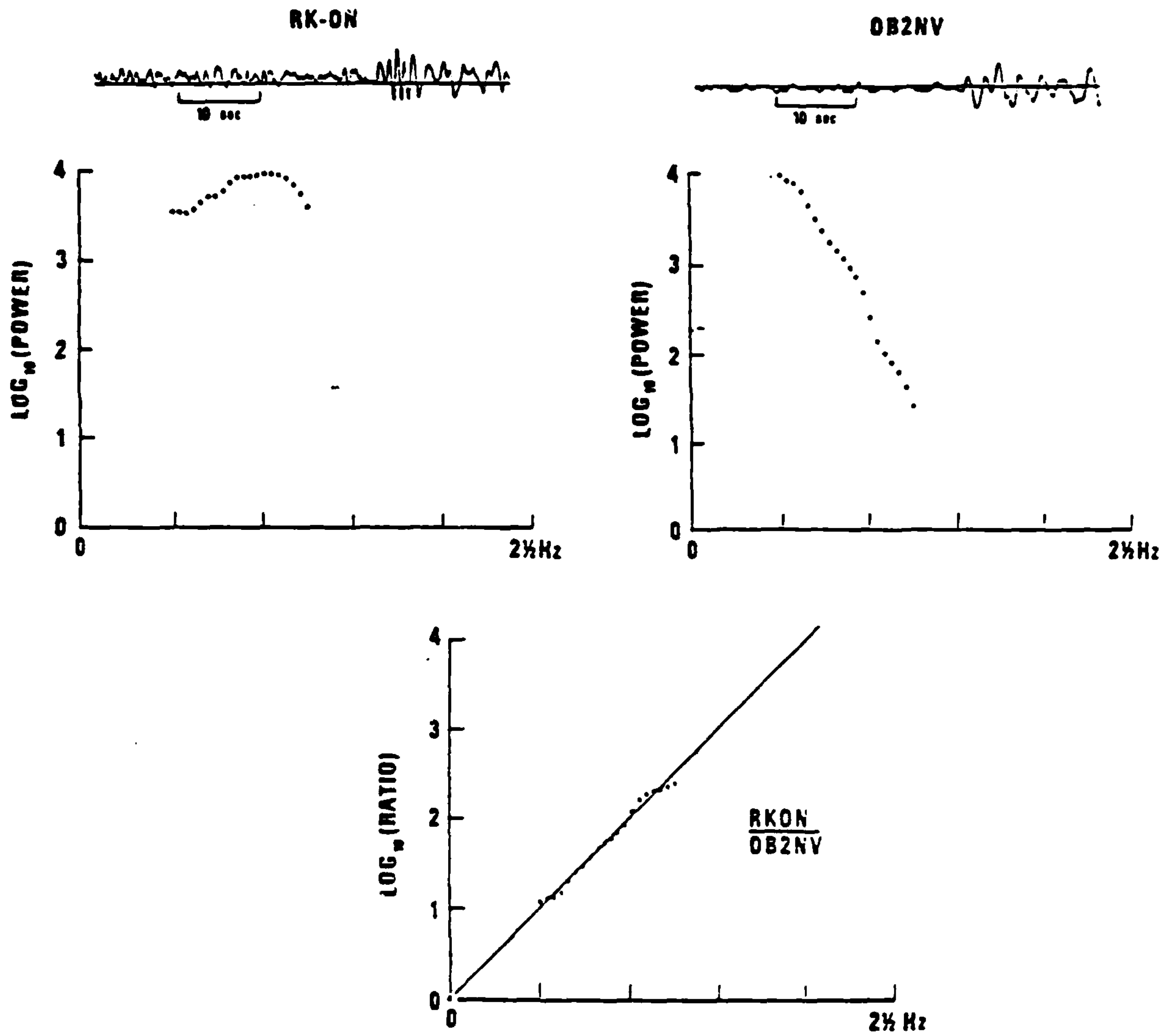


Spectral ratio between NTS and E. Kazakh explosions in the magnitude range (m_b) 5.0-6.0. Assumption of similar average source spectrum gives for attenuation difference between paths NTS-NORSAR and E. Kazakh-NORSAR the value t^*=0.28, from the visually fitted line in the figure. This figure is in agreement with the attenuation difference between southwestern North America - NORSAR and Central Asia - NORSAR paths, determined from the ratio of earthquake spectra[12] as 0.279 ± 0.004.

-2-results of an inversion experiment.

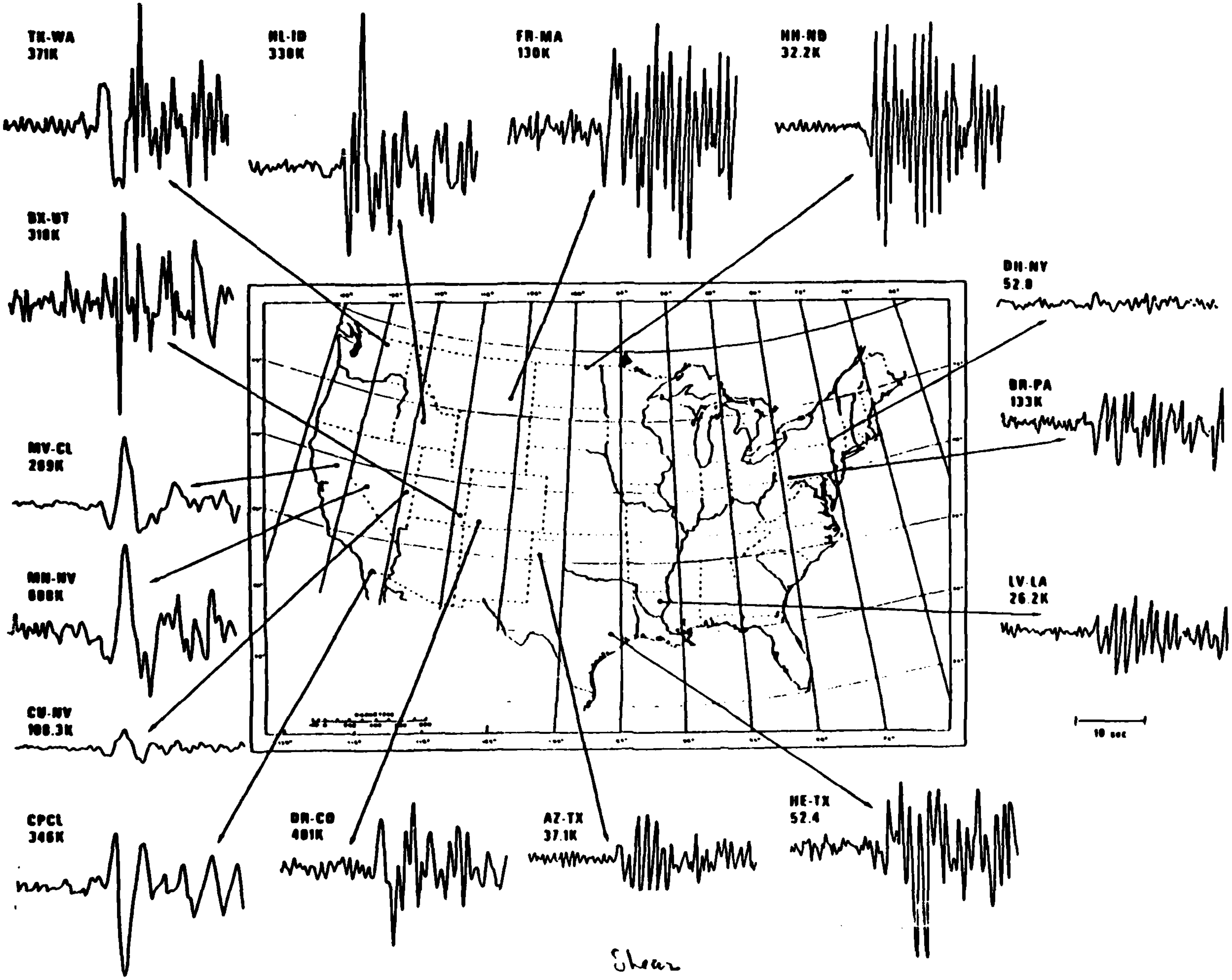
t^* (NTS-NORSAR)	0.403 ± 0.076
t^* (E. Kazakh-NORSAR)	0.150 ± 0.027
The common f_o	0.61 ± 0.25 hz
RMS	2.5 dB

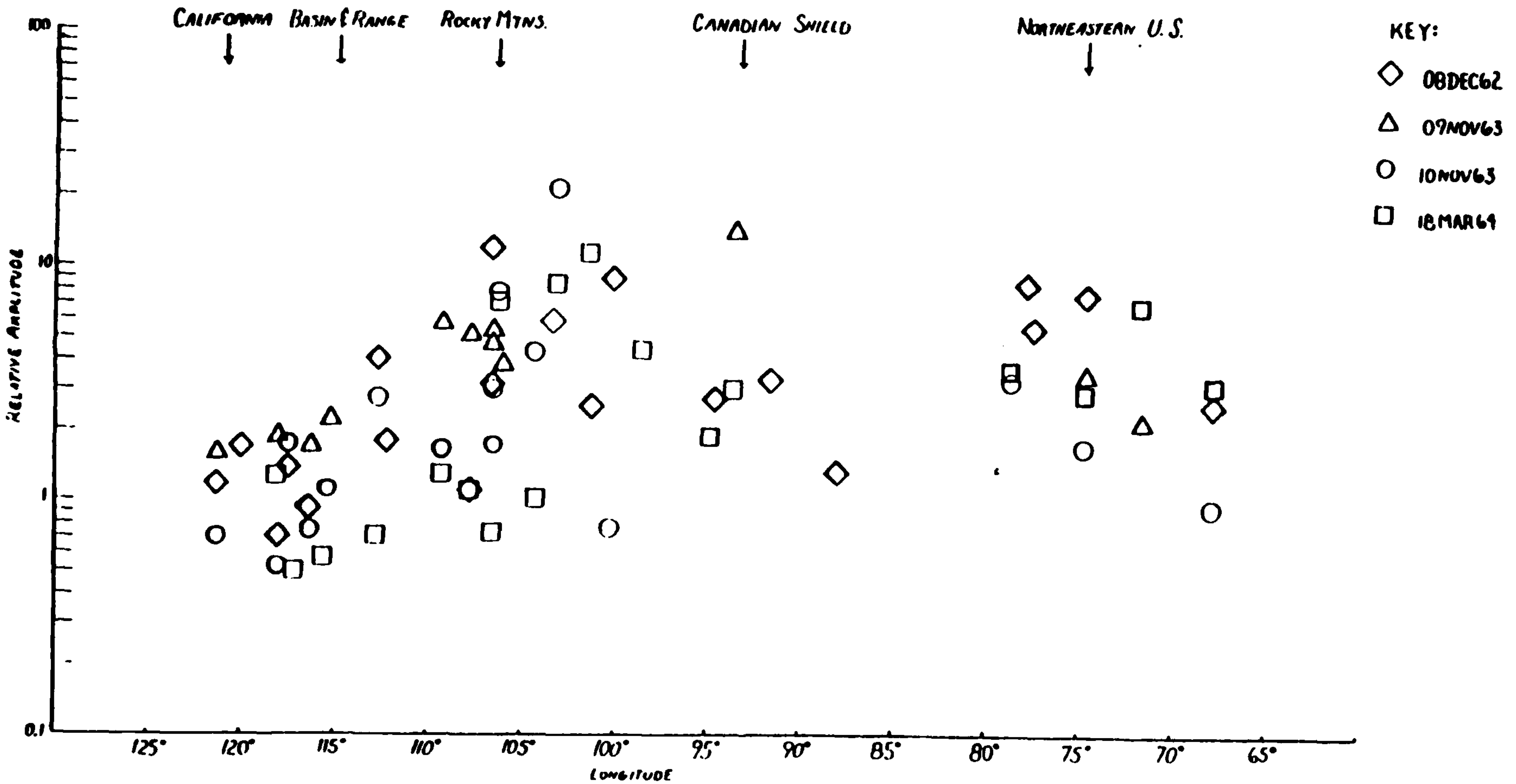
RADIAL
4 SEP 77
23:20



Shear

PERU-BRAZIL BORDER 09.25 71.5°W.
10 NOV 63 OT = 01:00:38.8




CALIFORNIA
BASIN & RANGE
ROCKY MTNS.
CANADIAN SHIELD
NORTHEASTERN U.S.
KEY:
08DEC62
09NOV63
10NOV63
18MAR64
RELATIVE AMPLITUDE
100
10
1
0.1
125°
120°
115°
110°
105°
100°
95°
90°
85°
80°
75°
70°
65°
LONGITUDE

Shear

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10 March 1980

Col. George Bulin
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Arlington, Virginia 22209

Dear George:

Enclosed is a little addendum to my comments on State-of-the-Art: Regional Attenuation Effects with Special References to the Nevada Test Site. In that section, in my discussion of Figure 9, I mentioned some NPNT data. I have found the data, and the attached spectral ratios are the results. The two USSR shots are the only ones at NPNT from the granite test site that were close to PILEDRIVER in magnitude, were already digitized, and were not clipped. All three events have m_b = 5.6 (ISC), and Dahlman and Israelson give an estimated yield for 07 March 69 of 46 kt; and for 16 Nov 64, 49 kt. This may be compared to 56 kt for PILEDRIVER.

These ratios are the most directly relevant seismological data available. They bypass questions such as "Is NTS typical of the WUS? Is HNME a good analog for Semipalatinsk? What is the absolute value of t*?" The weakest point is, of course, that focusing and defocusing beneath either NTS and/or Semipalatinsk can force the relative magnitudes to depart from what would be true on the basis of relative absorption at 1 Hz. Another weakness, not too great in my opinion, is that perhaps the USSR explosions are not actually in granite, although they are from Degelan; or that the granite is different enough from that in the Climax stock that the spectral shape is different.

I would appreciate it if you could include this as part of my "original" contribution.

Sincerely,

Bob

Robert R. Blandford

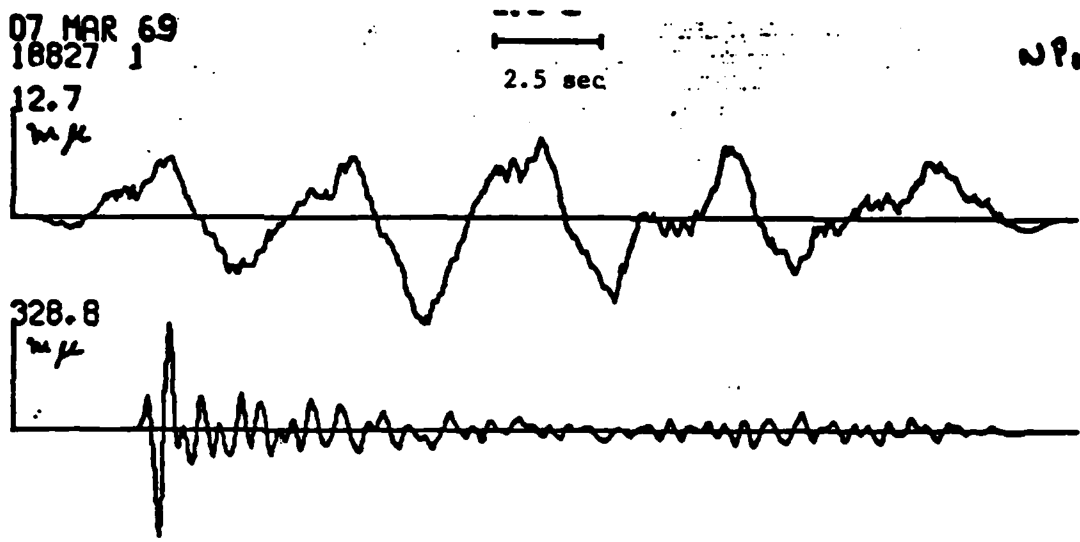
RRB/lms

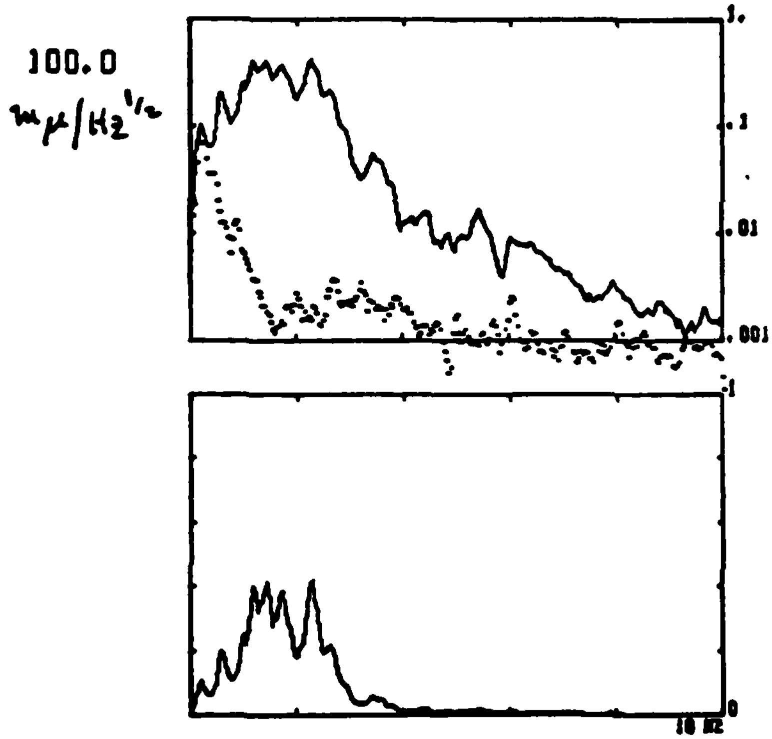
Enclosure

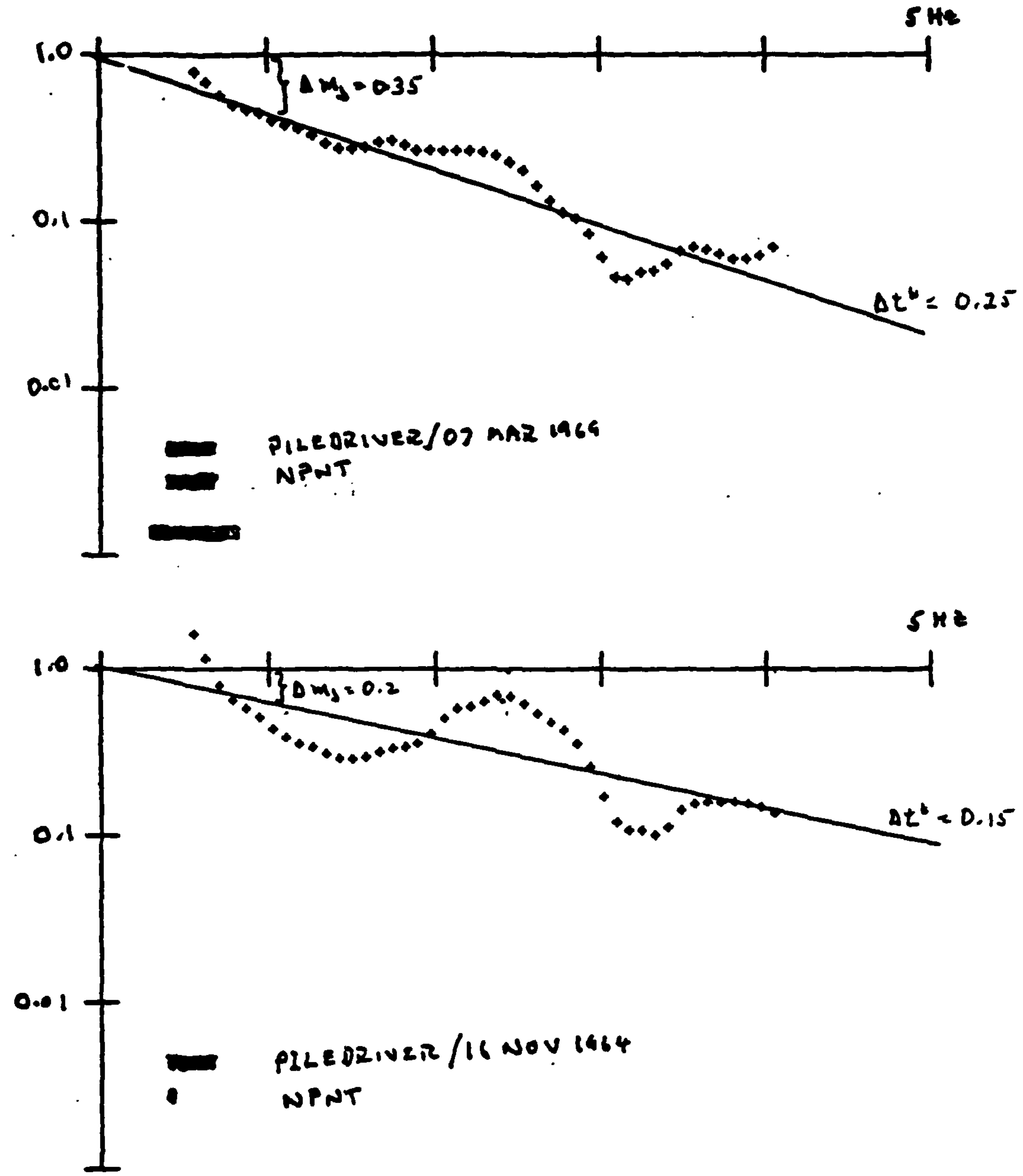
cc: M. Shore, T. Bache,
D. Harkrider, A. Ryall
W. Best, Z. Der, E. Herrin

07 MAR 69
18827 1

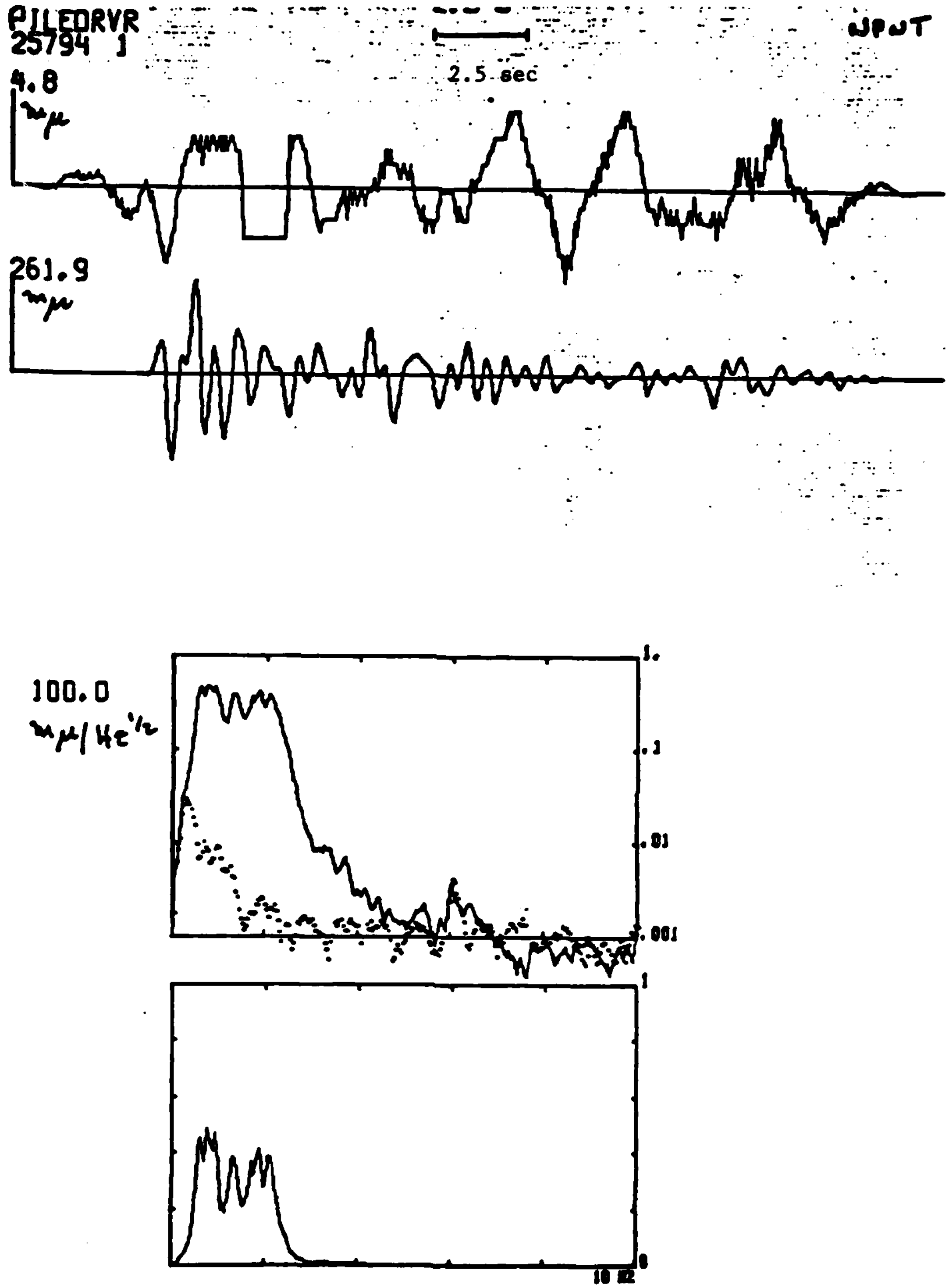
NPNT







Spectral ratio of granite explosion from NTS and granite-test-site explosions in the USSR as seen at NPNT. Ratios are computed for those frequencies for which S/N > 3 for both shots. The results indicate an average $\Delta t^* = 0.2$ leading to an expected average magnitude difference due to absorption at 1 Hz of 0.28. The max--ima in the ratios near 2.5 Hz may be due to a pP null for the USSR events. The null for PILEDRIVER has been shown to be weak and located near 6 Hz by Shumway and Blandford(1977). The large amplitude due to the 16 Nov 1964 pP null may be leading to an underestimate of Δt^* and Δ mb for this event. If so then the true average values may be closer to those indicated for the 07 March 69 event. Note that the difference in spectral magnitude at 1 Hz as compared to that determined by any plausible extrapolation to low frequency is > 0.4mb. This is completely independent of any theoretical concepts of Q, t*, absorption band corners, etc., and if the difference is due to absorption in the earth then we must conclude that there is a magnitude difference due to absorption irrespective of the numerical values of Q, t*, or absorption band corner frequency.



PILEDRVR
25794 1
WPWT
2.5 sec
4.8
mμ
261.9
mμ
100.0
mμ/Hz^1/2
1.
.1
.01
.001
1
0
10 HZ

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Robert G. North

REGIONAL ATTENUATION EFFECTS ON P-WAVES

(Writer discusses these effects only at frequencies of teleseismic interest since he has no experience with higher frequency data)

The primary means of estimating yield by seismic means continues to be the use of magnitude-yield relationships, particularly those involving body-wave magnitude m_b. The attenuation of the short-period (~1Hz) P-waves from which m_b is measured is dominated by crustal and upper mantle effects near source and receiver, and since substantial differences in velocity structure exist in these depth ranges between different regions of the world it is quite certain that similar large variations in Q exist. The existence of large lateral variations in the attenuation of all types of seismic waves has long been recognized. Evidence from the propagation of crustal body-wave phases (Romney et al, 1962), (Molnar and Oliver, 1969), the transmission of long-period P- and S-waves (Solomon and Toksoz, 1970), and surface wave amplitudes (Solomon, 1970) has demonstrated the presence of large differences in Q in the upper mantle. These results all indicate that attenuation is highest in the regions of mid-ocean ridges, subduction zones, and 'rift' structures such as the western US, and lowest in stable regions such as shields and deep ocean basins. High attenuation also appears to be correlated with high heat flow and negative velocity anomalies (Romney et al., 1962), (Evernden and Clark, 1970)

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As a result of these large variations in Q, substantial differences in the amplitudes of short-period body-waves, and thus in m_b, can be expected. Substantial station biases in m_b have been noticed by Evernden and Clark

(1970), Basham (1969), and Bune et al. (1970). North (1977) used magnitudes reported in the ISC bulletin to determine station magnitude corrections for over 100 stations. These corrections, which were well correlated with structural divisions such as shield and rift, ranged from +0.37 to -0.32 m_b units, corresponding to amplitude differences approaching one order of magnitude. These corrections correspond to receiver (station) differences in Q only, and similarly large differences between source region Q must also be considered. In the worst possible case one may consider a source in a rift region (e.g. Baikal rift) situated such that the only recording stations are also in zones of high attenuation such as a subduction region (e.g. behind Pacific island arcs). It is clear that the resulting m_b in such a case could well be up to 1 magnitude unit less than that for a source of similar size located and recorded in regions of low attenuation (e.g. E. Kazakh observed in Scandinavia).

Adequate calibration of attenuation is thus essential for the application of magnitude-yield relations, and in fact different such relations are used for different test sites - e.g. Western US compared to E. Kazakh. Marshall et al. (1979) have hypothesised a relationship between P-wave velocity and upper mantle Q, and in particular that P_n, the upper mantle surficial P-wave velocity, is a good measure of Q. This hypothesis is fairly well supported by the data they used, and also (though less convincingly) by the results of Booth et al. (1974) and North (1977). At present such a relation is one of the few quantitative means of estimating upper mantle Q and its worldwide variation from existing data. It has the advantage that P_n may be measured for many paths whereas individual station

bias may only be determined at single points (the stations). An invaluable (though extremely unlikely) experiment would be the calibration of both sites and receivers by explosions of known yield.

Direct determinations of Q using explosive sources have been made for the Western US by Passechnik (1970) and Veith and Clawson (1972) and for the USSR by Berzon et al. (1974). A comprehensive study of the Soviet literature would no doubt yield much more information on Q in the USSR. It should be noted here that Q in the short period band may vary substantially with frequency, and that results based on high-frequency data from explosions at regional distances may not be directly applicable at teleseismic distances. A curious result has been obtained by Butler (1979) in a study of P waveforms recorded at WWSSN stations in the US from Soviet explosions. He claims to find no appreciable difference in amplitudes between Western and Eastern US. This claim is, in my opinion, not entirely supported by the data he presents and is directly counter to the result of all other studies.

In conclusion, the most comprehensive study of receiver effects (station magnitude corrections) is that of North (1977); the most valuable discussion of both source and receiver effects of regional variations in attenuation is that given by Marshall et al. (1979). Calibration of Q from a worldwide survey of P_n velocities has been attempted by Marshall et al. and could easily be extended to cover most of the Soviet bloc in some detail since continuing deep seismic sounding (DSS) surveys as reported in the Soviet literature provide increasingly detailed contouring of P_n. A valuable collection of translations of this literature has recently been

published by Piwinskii (1979). The validity of the P_n-Q relations of Marshall et al. should, however, be further investigated; it may well be seriously in error in very complicated regions such as subduction zones.

A detailed and up-to-date bibliography is provided by Marshall et al., and this is appended together with a few additional references.

R.G.North

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From Marshall et al. (1979)

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Frequency Dependence and Regional Variations of Q_{ScS}

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Report submitted to Defense Advanced Research
Projects Agency

February, 1980

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I. Frequence Dependence

In the study of seismic-wave propagation the intrinsic quality factor Q is usually assumed to be independent of frequency. A constant-Q model of seismic attenuation is both theoretically convenient and empirically consistent with most available data on wave amplitude decay (Knopoff, 1964; Anderson and Archambeau, 1964; Kanamori and Anderson, 1977; Jordan and Sipkin, 1977). Nevertheless, the possibility that Q varies with frequency has intrigued seismologists for many years. Gutenberg (1958) was evidently the first to advocate that the apparent Q of teleseismic P waves (Q_P) increases with frequency, and his hypothesis has received additional support from later studies (Kurita, 1968; Archambeau et al., 1969; Solomon and Toksöz, 1970; Solomon, 1972; Der and McElfresh, 1977; Lundquist, 1977). The variation of Q_P is difficult to measure, however, because Q_P is generally large (~1,000 at 1 Hz) and even substantial variations in its value have only small effects on P-wave amplitudes and wave forms. These are easily obscured by the uncertainties in source excitation and propagation effects other than anelastic attenuation. Consequently, the dependence of Q_P on frequency has not been precisely quantified.

SH-polarized shear waves are more severely attenuated than compressional waves, and their structural interactions are simpler; hence, they are often more suitable for the study of anelastic structure. In a previous report (Jordan and Sipkin, 1977), we recovered the attenuation operator for multiple ScS waves propagating in the western Pacific by applying a spectral stacking technique to digitally recorded data from High-Gain Long-Period (HGLP) stations in Hawaii and Japan. The spectral modulus of this attenuation operator yields the apparent Q of SH-polarized ScS waves (Q_{ScS}) as a function

of frequency. We observed no significant frequency dependence of Q_{ScS}; the data were consistent with the estimate $Q_{ScS} = 156 \pm 13$ throughout the band 0.006 to 0.06 Hz.

In a more recent study (Sipkin and Jordan, 1979), data from HGLP instruments at KIP and MAT and WWSS LP and SP instruments at KIP and GUA have been used to study the amplitude characteristics of ScS and multiple ScS waves from deep-focus earthquakes. The data at low frequencies (0.006 to 0.06 Hz) are consistent with our previously published estimate, $Q_{ScS} = 156 \pm 13$ (Jordan and Sipkin, 1977). However, at high frequencies (>0.1 Hz), Q_{ScS} appears to increase rapidly with frequency. Lower bounds on Q_{ScS} are obtained by assuming a flat source spectrum and ignoring any energy losses due to scattering; we find that Q_{ScS} must be greater than 400 at frequencies between 1 and 2.5 Hz. Correcting for a source spectrum with a corner at 0.16 Hz and an asymptotic roll-off of ω^{-2}, considered appropriate for these events, raises this estimate to about 750. The increase in Q_{ScS} at frequencies about 0.1 Hz is consistent with a spectrum of strain retardation times which has a high-frequency cutoff in the range 0.2 to 1.0 sec. At very low frequencies Q_{ScS} can be estimated from normal mode data; the best available models yield values of about 230. Comparison of these estimates with our data suggests that Q_{ScS} decreases with frequency in the vicinity of 0.01 Hz. Because the scattering coefficient increases rapidly with frequency, the fact that significant ScS amplitudes are observed at high frequencies implies that any bias in Q_{ScS} measurements due to scattering at low frequencies is probably small. We show that, although our data provide only integral constraints on the variation of Q_μ with depth, the regions in which Q_μ is frequency dependent occupy a substantial portion of the mantle, probably including at least part of the mantle below 600 km depth.

II. Lateral Variations

The quality factor for ScS waves, Q_{ScS}, is a parameter diagnostic of terrestrial anelasticity, averaging the anelastic properties of the entire mantle. Numerous estimates of this quantity have been derived from the spectral ratios of multiple ScS phase pairs (Kovach and Anderson, 1964; Sato and Espinoza, 1967; Yoshida and Tsujiura, 1975). In a previous paper (Jordan and Sipkin, 1977) the problem of ScS attenuation has been formally posed in the frequency domain as an inverse problem for a linear, complex-valued ScS attenuation operator, and its solution has been derived by standard least-squares techniques. The algorithm based on this analysis has a number of advantages over the classical spectral-ratio method. The spectral products and cross-products for various ScS_n phase pairs from different sources are phase-equalized and summed (stacked) prior to taking ratios. Stacking increases the signal-to-noise ratio (SNR), stabilizes the estimates, and helps to average out the effects of local heterogeneities and source variability. Moreover, measures of the SNR for individual phases are used to weight the signals in the stacks and to estimate noise-induced uncertainties in the model parameters.

In our 1977 paper this algorithm was applied to a set of 17 multiple ScS phase pairs digitally recorded by High Gain Long Period (HGLP) stations at Kipapa, Hawaii (KIP), and Matsushiro, Japan (MAT) from deep-focus events in the western Pacific. Stable estimates of the amplitude and phase response of the ScS attenuation operator were derived in the frequency interval 6-60 mHz. Within this band the apparent Q of ScS_n waves multiply reflected beneath the western Pacific was estimated to be 156 ± 13, and no significant frequency dependence of Q_{ScS} was observed. Our estimate of Q_{ScS} for the

western Pacific was considerably less than the values published for other regions. Kovach and Anderson (1964), for example, obtained $Q_{ScS} = 600$ for South America, and Yoshida and Tsujiura (1975) obtained $Q_{ScS} = 290$ for the Sea of Japan. Taken at face value, these observations require very large geographical differences in the attenuation structure of the mantle.

In a more recent study (Sipkin and Jordan, 1980), the techniques of Jordan and Sipkin (1977) have been employed in the assessment of the lateral variations of Q_{ScS}. Substantial regional differences in Q_{ScS} do exist, but these do not appear to be as extreme as the discrepancies among the published estimates imply.

The ScS_n phase-equalization and stacking algorithm of Jordan and Sipkin (1977) has been applied to an extensive set of HGLP and ASRO data to obtain regionalized estimates of Q_{ScS}. Tests of the algorithm using synthetic data reveal no significant sources of bias. The low value of Q_{ScS} previously obtained for the western Pacific (156 ± 13) is corroborated by additional data, and Q_{ScS} observations in other regions correlate with variations in crustal age and tectonic type. A representative value for the ocean basins sampled by our data is 150, with the best estimates being somewhat lower (135-142) for younger oceanic regions and somewhat higher (155-184) for older regions. The two subduction zones sampled, Kuril-Japan and western South America, are characterized by larger Q_{ScS} estimates than the ocean basins (197 ± 31 and 266 ± 57, respectively), and the difference between them is qualitatively consistent with the contrasts in upper mantle attenuation structure proposed by Sacks and Okada (1974). Continental regions are poorly sampled in this study, because the signal-generated noise in the vicinity of the ScS_n phases is generally larger for continental paths,

but a representative value is inferred to be $Q_{ScS} = 225$. For paths crossing China Q_{ScS} is observed to be lower (~180), providing additional evidence for a high-temperature upper mantle previously inferred from surface-wave and travel time measurements. Our best estimate for the average Earth is $Q_{ScS} = 170$ (±20%), which appears to be significantly lower than that predicted by normal mode data, suggesting some frequency dependence. Q^{-1}_{ScS} correlates with $ScS_n - ScS_{n-1}$ travel time along a line given by $Q_{ScS}{}^{-1} = (4.4 \times 10^{-4})\ \Delta T_{ScS} + 4.88 \times 10^{-3}$, where ΔT_{ScS} is the JB residual in seconds; this correlation favors a thermal control on the ΔT_{ScS} variations. It is inferred from the tectonic correlations that much, if not most, of the heterogeneity expressed in the Q_{ScS} and ΔT_{ScS} variations is confined to the upper mantle. Substantial differences in the attenuation structures underlying continents and oceans are implied. In fact, the average quality factor for the upper mantle beneath stable cratons may not be much less than that for the lower mantle.

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January 31, 1980

Col. George Bulin, USAF
Nuclear Monitoring Research Office
Defense Advanced Research Projects Agency
1400 Wilson Boulevard
Arlington, VA 22209

Subj: State-of-the-art Assessment

Ref: Ltr Romney to Rodean dtd 18 December 1979

Dear George:

I am sorry for this late response to your request for the subject assessment, but I have been on official travel during two of the four weeks since I received the referenced letter.

The following is based, for the most part, on work that I did with Peter Marshall and Don Springer during 1974-1977 and on some of my work during the past year. I also reviewed some of the appropriate literature while preparing this letter.

Regional Attenuation Effects on P-Waves

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There is substantial evidence for regional variations of seismic attenuation. A summary of the evidence published through 1976 is given in Section 2 of Marshall, Springer and Rodean (1979). Work has continued since then, with emphasis on collecting and analyzing seismic data, such as that obtained at NTS under the SDCS Project. Recent results were reported during one session of the VSC Research Review on 26-27 September 1979.

In my opinion, other kinds of work (not on the VSC Research Review agenda) should be done to relate regional variations in seismic attenuation to regional variations in other geophysical parameters including seismic velocity, heat flow, electrical conductivity, and gravity. Marshall, Springer and Rodean (1979) developed and applied an empirical relation between Q in the upper mantle and Pn velocity. More work is needed to test and prove this empirical relation and to determine whether there is a physical basis for such a relation. The value of a proven relation between seismic attenuation and one or more other geophysical parameters is that such a relation would provide a basis for estimating attenuation in the absence of direct measurements of attenuation. Some theoretical work by Chung (1977, 1979) indicates that there may be a relation among seismic attenuation, seismic velocity, and partial melt fraction in the low-velocity zone of the upper mantle. Walker, Stolper and Hays (1978) concluded that there are upper limits to the melt in the low-velocity zone because of stability considerations. Goetze (1977a, 1977b) and Shaw (1978) suggest that partial melting is not necessarily the cause of attenuation in the low velocity zone. Current experimental measurements by Brian Bonner at his Laboratory indicate that seismic attenuation in rocks may increase significantly with increasing temperature at temperatures substantially below the "dry" solidus

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Site Specific Propagation Effects

In the preceding, the subject is regional (large-scale) variations of attenuation. Marshall, Springer and Rodean (1979) applied data obtained over large regions to specific explosion sites. For example, they made body-wave magnitude corrections for attenuation in the upper mantle for explosions at NTS and at the Soviet test site in eastern Kazakhstan on the basis of Pn velocities in those respective regions. Some of my recent work (Rodean; 1979b, 1979c) is consistent with the indications by the Pn velocity data for these sites that seismic velocities beneath the Soviet site are significantly higher than beneath NTS. Assuming detonations on the minute, the Soviet explosions have an average ISC origin time of 2.3 seconds before the minute while NTS explosions have an average ISC orgin time of only 0.4 second before the minute, a difference of almost 2 seconds.

We have known for some years that the seismic signals from explosions in Rainier Mesa are generally stronger than those explosions of comparable yield below the water table in Yucca Flat and Pahute Mesa. This is illustrated by the Swedish yield estimates for explosions in these three areas that are analyzed in Rodean (1979a). Dahlman and Israelson (1977) based their estimates on announced yields for explosions in Yucca Flat and Pahute Mesa; they did not have any "calibration explosions" for explosions in Rainier Mesa. The similarities and differences among the seismic observations and the measured shot-point rock properties (Ramspott and Howard, 1975) for these three areas suggest to me that propagation effects, not seismic coupling, may be responsible for the stronger signals from Rainier Mesa. For example,

explosions below the water table in Yucca Flat and Pahute Mesa have a common m_b:yield relation but their average shot-point rock properties are different. On the other hand, the shot-point rock properties of Rainier Mesa are about the same as for Yucca Flat below the water table but are different from those for Pahute Mesa below the water table.

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Source Theory and Observation for Surface Waves

The following are some miscellaneous thoughts on this subject. I have not done as much work on this topic as I have on the other two.

It is generally assumed that the low-frequency or final value of the reduced displacement potential (RDP) is related to the surface wave amplitude. The problems of RDP measurement (e.g., accuracy of integrating over a sufficient period of time, effects of reflections from interfaces at later times) are such that the early part of the RDP (in the time domain) is more accurate than the latter part. Therefore the RDP may be a more accurate source for the short-period body waves than for the long-period surface waves.

Peter Marshall has an interesting hypothesis about changes in the Ms:m_b relation as the explosion depth is changed from above to below the water table (Marshall, 1978, p 50). He suggests that, when an explosion is at the level of the water table, Ms may reflect low-coupling in the upper medium while m_b reflects high-coupling in the lower medium. Interfaces may affect Ms in another way. Hudson and Douglas (1975) made calculations of Rayleigh waves in a system consisting of a source in an elastic layer over an elastic half-space. They found a connection among the group velocity of Rayleigh waves, the spectral amplitudes of surface waves generated by a source, and the resonance of vertically-travelling P waves in the surface layer. A minimum in a group velocity curve is reflected as a maximum in the spectral amplitudes. Also when a sharp impedance contrast exists between the surface layer and the half-space, the group velocity minimum in the fundamental mode occurs close to a period equal to four times the P-wave travel time from the surface to the

interface. Wheeler, Preston and Frerking (1976) found such a P-wave resonance in close-in ground motion data for seven of eight Yucca Flat tests studied. The waves were trapped between the free surface and the Paleozoic rocks. The wave period was equal to four times the transit time between the two surfaces, and it was independent of explosion yield and depth. A question: does this P-wave resonance phenomenon have significant effects on the surface waves?

References on Surface Waves

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I hope that the above is of use to you.

Sincerely,

Howard C. Rodean

HCR:dt

Regional Attenuation Effects on P Waves and Effects of Attenuation on Surface Waves. Charles Archambeau - University of Colorado

Introduction.

The effects of attenuation for both surface and body waves can be described, most appropriately, in terms of intrinsic dissipation functions $Q_\alpha(r, f)$ and $Q_\beta(r, f)$ for the earth. Here Q_α and Q_β are Q functions for compressional and shear wave losses respectively, and both are functions of radius and frequency[+]. They are also functions of the other spatial coordinates, but it is easier and actually most appropriate to define different Q models for different geologic provinces. It is quite clear from observational results for body waves (e.g., Archambeau, Flinn and Lambert, 1969) that much of the dissipation of P waves takes place in the low velocity zone. Therefore high attenuation is correlated with high heat flow and large P delays, these in turn characterizing geophysical-geological provinces. In particular, shield areas with low heat flow and negative P-delays show low attenuation, while active tectonic provinces with high heat flow and positive P-delays show high attenuation. All of these effects are clearly related to the depth span and intensity of the low velocity zone. These correlations are clearly shown form the studies of pP pulses from earthquakes in trench zones (e.g., Barazangi, Pennington and Isacks, 1975), as well as from teleseismic P wave observations from explosions (e.g.,

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[+] Q_α and Q_β can be related to each other under the assumption that dissipation in pure compression is very small relative to losses in shear. Then, for typical mantle elastic velocities, $Q_\alpha \approx 9/4\, Q_\beta$. For details see Anderson et al., 1965.

Der et al., 1975; Der and McElfresh, 1976). Thus, for body wave magnitudes, one expects variations in m_b which are directly related to the geologic provinces of the source and receiver. In particular, sources in tectonic provinces will show reduced m_b values at 1 Hz relative to the same sources in shield regions. Since tectonic provinces show, in general, highly variable low velocity zone thicknesses and correspondingly variable heat flow values and P delays, one can also expect variability in the m_b reduction, from quite large reductions to rather small reductions, depending on precisely where the source is located.

Similar statements can be made about surface wave attenuation relative to Geological-Geophysical provinces. That is, the strongest attenuation occurs within the low velocity zone and strong surface wave attentuation is correlated with regions of high heat flow, large P (and S) wave delays and tectonic activity. For shorter period surface waves not penetrating the low velocity zone however (i.e., for periods less than 30 seconds), the attenuation is not as large as for longer period surface waves (i.e., the observed Q is about 300 compared to observed Q values of around 100 for surface waves in the period range 30-200 seconds) and there is less regional dependence in attenuation (e.g., Solomon, 1972). However, Mitchell, 1975, has shown that for rather short period surface waves, near 5 seconds, the attenuation is quite strongly regionally dependent. Nevertheless, he finds that for the longer periods up to 30 seconds, there is little regional dependence. This very short period regional dependence in attenuation is probably more related to scattering than to anelastic effects, in that tectonically active provinces usually show larger near surface lateral variability in velocity structure than do the more stable provinces.

Thus M_s values based on 20 second Rayleigh waves do not show strong regional "Q-bias".

The frequency dependence of the anelastic dissipation has only recently been considered in any great detail. Originally Archambeau et al., 1969, showed that P_n phases in the Western U.S. were attenuated such that the high frequencies required high Q_α values than the lower frequencies - that is, the Q_α appeared to increase with frequency in the frequency range from .5 to 3 Hz. These direct observations were also in agreement with the observation that Q_α (and Q_β) models obtained from low frequency surface waves had lower Q_α values, essentially everywhere in the mantle, when compared to the Q_α model obtained from high frequency (1 to 3 Hz) body wave observations. The upper mantle Q_α models that have been obtained from low frequency surface wave and free oscillation data and from high frequency body wave data are shown in Figure 1. The model SL8 is from the analysis of free oscillation data by Anderson and Hart , 1978; the model MM8 is from surface wave data inversion by Anderson et al., 1965; and the model AFL is from body wave data inversion by Archambeau et al., 1969. Each model applies only to the frequency range covered by the data used to obtain it. The trend of these results is toward high Q values with increasing frequency of the data used in this inversion.

Solomon, 1972b, proposed a frequency dependent intrinsic Q for the mantle involving activated processes that satisfied the observed long period surface wave dispersion quite well. Liu and Archambeau, 1975 and 1976, showed that this model fit the total set of surface wave and free oscillation data quite well and that it predicted relatively large shifts in the dispersion (group velocity versus frequency) and free oscillation periods,

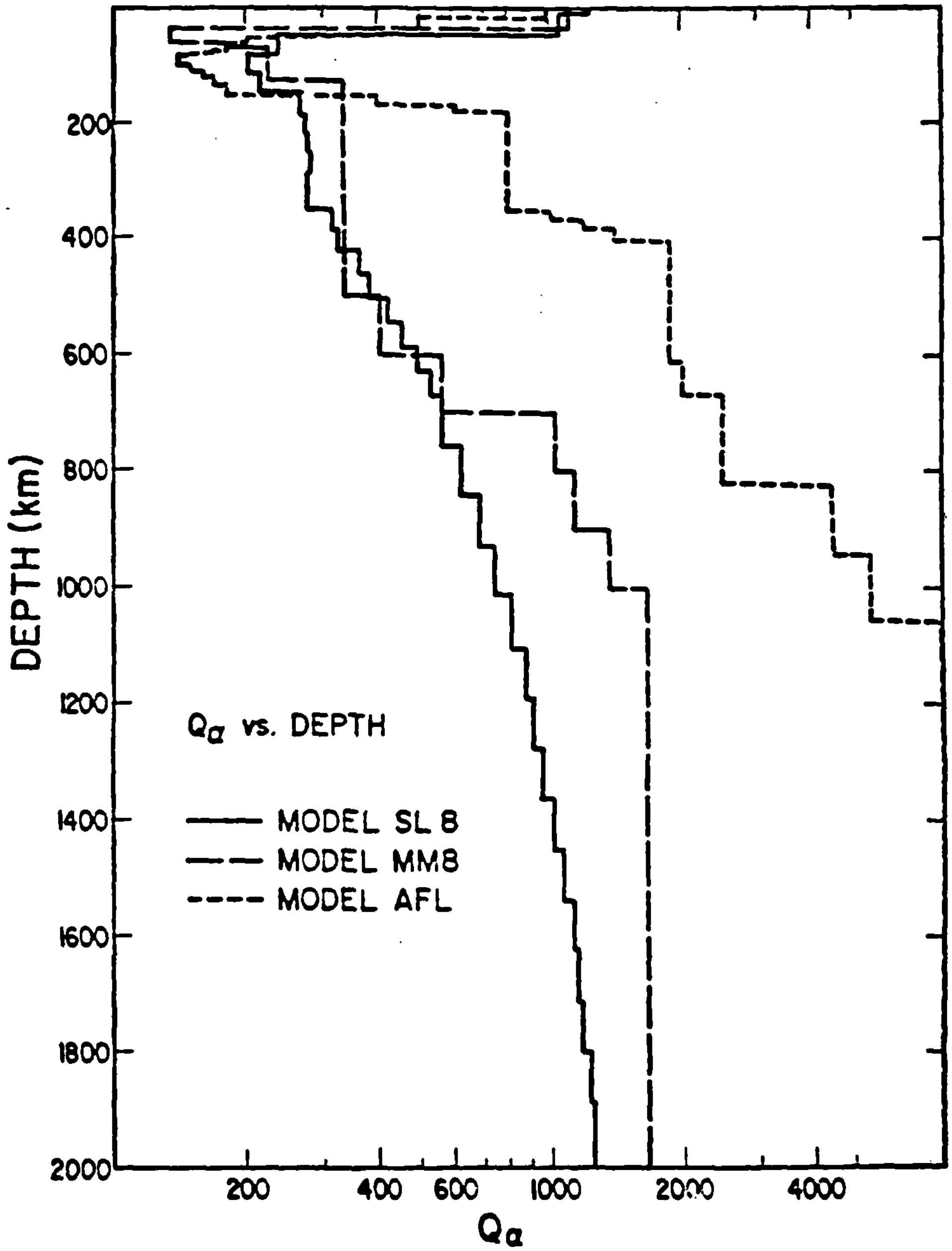

Figure 1. Q_α-vs-depth. Models SL8 and MM8 were derived from free oscillations and surface waves, respectively. Model AFL was derived from P body waves in the range 1-5 Hz. Note that Q tends to increase with the frequencies in the data set.

showing that the effective velocity structure sensed by low frequency waves is different than that for higher frequency waves. Liu et al., 1976, expanded upon these results and proposed an absorption band intrinsic Q model that consisted of a distribution of activated processes, each with a different characteristic relaxation time corresponding to a superposition of many absorption processes acting to dissipate energy. This model was also shown to be compatible with observations of surface wave and free oscillation dissipation. The absorption and Q model amounts to an extension of Solomon's model, wherein many activated processes are allowed rather than one or two, and is more realistic in terms of the known microphysics of crust-mantle materials.

Currently this kind of intrinsic Q model is being used to constrain the frequency dependence of the intrinsic Q in the earth, in order to invert for both the depth dependence and the shape of the absorption band at each depth (and hence the intrinsic frequency dependence of Q_α and Q_β). Figure 2, from Lundquist, 1980, shows the form of the absorption band models being used. Such an absorption band applies at each depth in the earth and varies with temperature pressure, material chemistry and phase state. The parameters τ_1 and τ_2 are low and high frequency "relaxation times" corresponding to the half amplitude points on the "Q-filter" in the frequency domain. These parameters are treated as unknowns and are obtained, as functions of depth, by inverting the observed attenuation data.

The Q models shown in Figute 1 have very poor resolution of Q variations in the crust, mainly because little or no very short period surface wave data was used for the inversion with the surface wave and free oscillation data

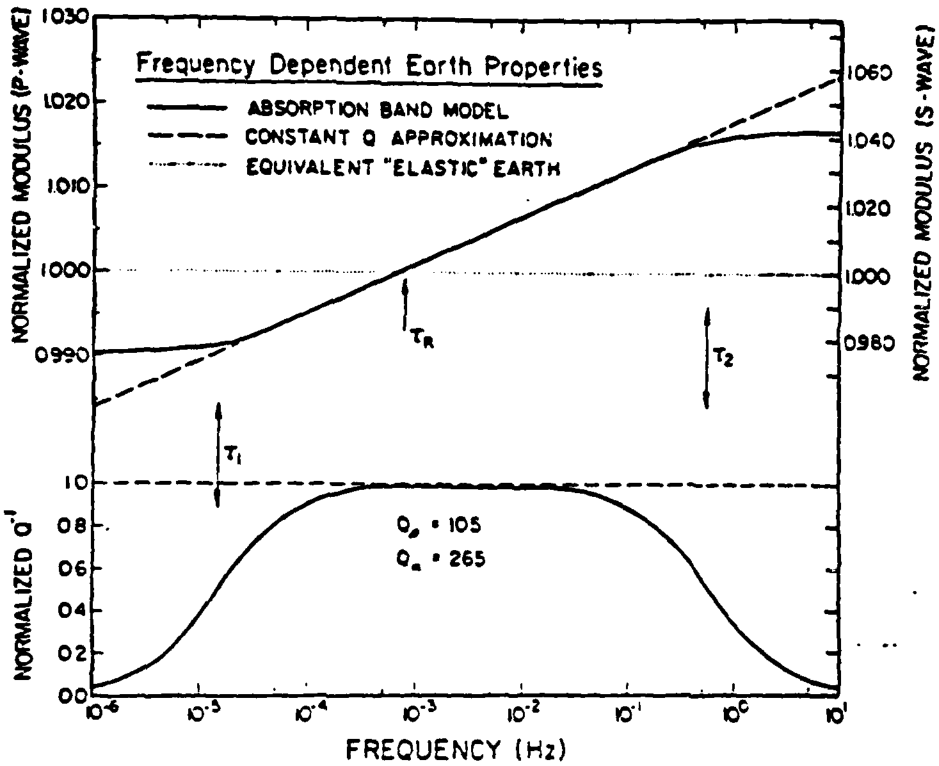

Figure 2. Frequency dependent Earth properties-vs-frequency. The normalized modulus is a direct measure of dispersion, and the absorption band is a direct measure of per-cycle attenuation. Note that the difference between dispersion for constant Q and absorption band Q is small but that the difference in implied attenuation is very large for $\omega > 1/\tau_2$.

and, in the case of the body wave derived model, the sampling of the crust using teleseismic P waves was minimal. Mitchell, 1980, has however studied relatively short period surface wave propagation in the Eastern U.S. (with periods from 1 to 40 seconds) and obtained crustal Q_β models in some detail. He has also shown that the intrinsic Q_β can be best described by a frequency dependent Q-function of the form:

$$Q_\beta(\omega, r) = Q_\beta(r)\omega^\zeta$$

with ζ between .3 and .5 for the period range 1 to 40 seconds. Here again it appears that the intrinsic Q increases with increasing frequency, however such a conclusion based on the fits given by Mitchell may be premature. In any case his models show a Q_β average of about 250 in the upper crust (0-15 km) and near 1000 for the lower crust (15-40 km). These values are significantly higher than the Q_β values in the low velocity zone of the upper mantle, where $Q_\beta \simeq 50\text{-}100$ is appropriate.

It is clear that the effects of attenuation on surface wave magnitudes, measured at 20 seconds, are not as extreme as are attenuation effects on body wave magnitudes. First, there is little observed regional variation in attenuation in this period range. Second, the attenuation is not very large, that is the Q of the crust, while of course variable in both frequency and with depth, is quite high. Thus, corrections in M_S for attenuation could be made and they would not be very large. It is of course important that M_S be measured at 20 seconds.

Frequency Dependent Q Models for Teleseismic P waves and Mantle Surface Waves

The best (i.e. only) first order frequency dependent Q model for the upper mantle has been obtained by Lundquist, 1980. The model uses

an absorption band intrinsic Q of the type shown in Figure 2. The model is obtained by first taking the previously determined low frequency Q models (the MM8 and SL8 models in Figure 1) and the high frequency model AFL in Figure 1 as appropriate Q variations in the mantle in the frequency ranges for which they are defined. That is, the frequency dependent model is constrained to give, to first order at least, the SL8 model at very low frequencies and the AFL model at high frequencies, near 3 Hz. The observed Q models in Figure 1 turn out to imply that there is one absorption band model for the mantle beneath the low velocity zone, having regular properties varying with depth in a manner consistent with the temperature-pressure variations in the earth in this depth range, and a separate, very different, kind of absorption band which appears to be confined to the low velocity zone. The absorption band for the low velocity zone appears to be narrow (i.e., τ_1 and τ_2 relatively close in value) while the lower mantle absorption band appears to be very broad (i.e., τ_1 very large and τ_2 near .1 sec). The second absorption band associated with the low velocity zone may be a consequence of a partial melt state within the zone. In any case it is confined to this zone and therefore varies with the extent and intensity of the low velocity zone.

Using such a rough double absorption band Q model as a starting point, the frequency dependent Q model can be refined by adjusting the various absorption band parameters (in particular the "high and low" frequency relaxation times τ_2 and τ_1 plus the maximum Q^{-1} level of the absorption band at each depth) to fit frequency and time domain observations. In particular, Lundquist adjusted the starting double absorption band model to be such that

when attenuation corrections are applied to observed earthquake and explosion P wave spectra, then corrected source spectra had high frequency asymptotic behavior of the form $1/\omega^2$ or $1/\omega^3$, as is expected from source theory considerations. Further, he used the resulting, somewhat refined, Q model to predict time domain synthetic P wave forms and further adjusted the model to achieve detailed fits to the first cycle of the P wave train from explosions. (Only the first cycle of the P wave train is reasonably well predicted by currect explosions models. Further, it is relatively free from uncertainties introduced by near source structure, tectonic release and spall phase production.)

The net result was that the initial double absorption band model, inferred from the low and high frequency Q models of Figure 1, fit the observations from NTS explosion events very well, with little adjustment necessary. Thus this model closely corresponds to the free oscillation model SL8 at low frequencies and the body wave model AFL at high frequencies and predicts the behavior of a mantle Q at other frequencies such that both spectral and time domain observations are well satisfied. Figure 3 shows the properties of this double absorption band model (solid line) as a function of depth and frequency in the earth. The upper inset indicates the apparent Q_α in the crust, which is poorly resolved but is high, as indicated. The next inset shows the typical form of the double absorption band in the low velocity zone. The dotted line shows the single absoprtion band that would exist if the low velocity zone were absent, so that the departure of the solid line from the single absorption band Q_α indicates the effect of the second absorption band associated with the low velocity zone. At greater depths the variation in the absorption band is such that the maximum level of Q_α for

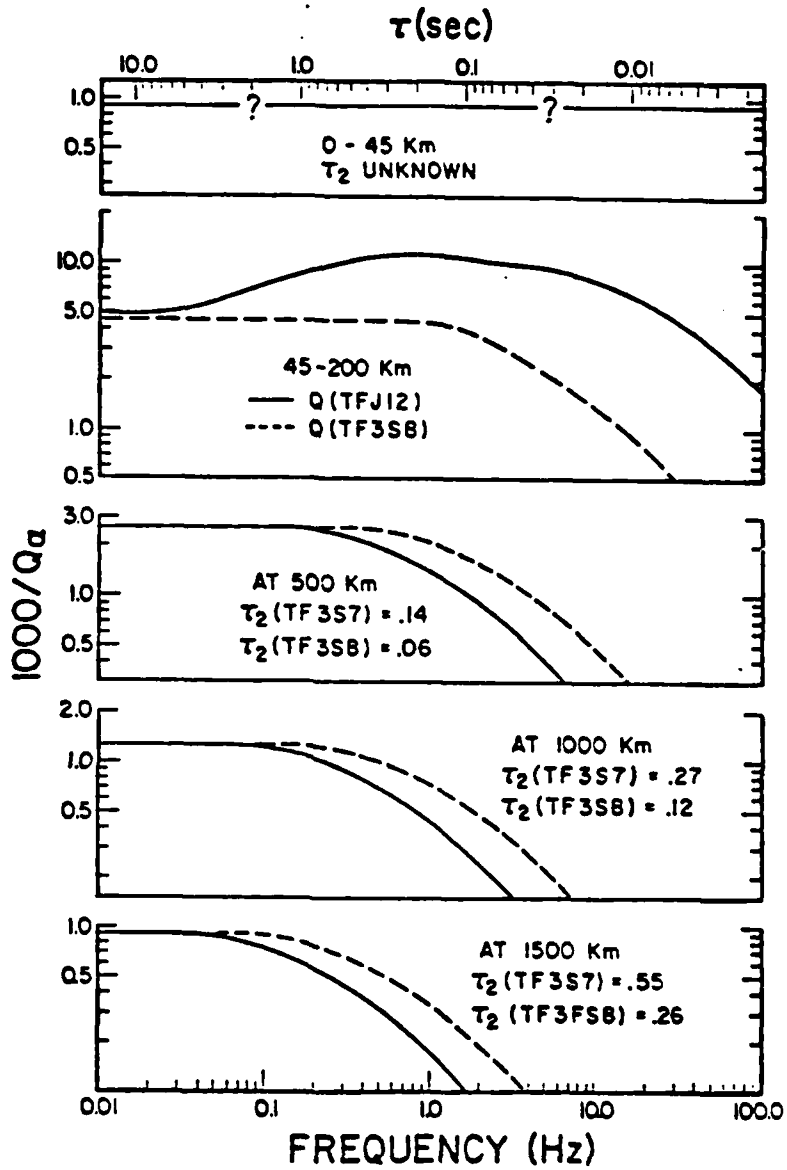

Figure 3. Absorption bands in the Earth-vs-frequency and depth. The dashed line represents the single absorption band model, TF3S8, and the solid line represents the double absorption band model, TFJ12. At low frequencies, both models converge to Q^{-1}(SL8), and model TFJ12 converges to Q^{-1}(AFL) at 3 Hz.

the band increases ($1/Q_\alpha$ decreases) and the relaxation time τ_2 increases; both uniformly in a manner controlled by the temperature-pressure increases with depth.

On the other hand, for non-tectonic regions, the Q_α variation with depth and frequency was found to be somewhat different than for the Basin and Range region. In particular, using explosion data from Novaya Zemlya, so that a stable platform region was sampled, the Q_α variation with depth was intermediate between the single absorption band variation shown by the dotted line in Figure 3 and the double absorption band model for the Basin and Range. (See Lundquist, 1980 for details.) This appears to be due to a less intense and thinner low velocity zone for the stable platform region and a correspondingly more depth confined and less intensive second absorption band in the 45-200 km depth range. This of course again implies regional variations in attenuation, but specifically that this variation is controlled by the presence or absence of the second absorption band. Further, because of the nature of this absorption band, in particular its frequency band width, the frequency dependence of the absorption can be quite different from region to region.

The consequences of this kind of Q model, in terms of t^* (total travel time divided by the effective Q over the path of the wave), are shown in Figure 4 for the double absorption band model. Clearly t^* is quite strongly frequency dependent.

These results have a number of important implications. First it seems evident that t^* should <u>not</u> be used in modeling work, but rather the Q_α or Q_β models should be used and modeling should be done in the frequency domain in order to properly account for both the depth and frequency dependence of the Q and for the different apparent "elastic" velocities sensed by waves of different frequency. Second, high frequency seismic energy is

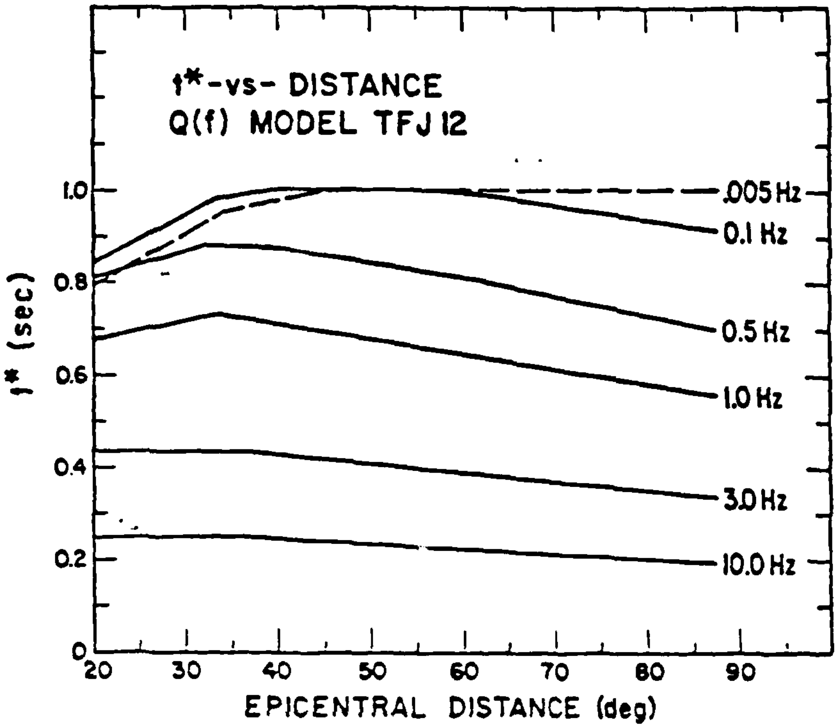

Figure 4. t*-vs-distance for selected frequency. The velocity model used is QM2, and the Q model is the double absorption band model, TFJ12. t*(TFJ12) converges to t*(AFL) at 3 Hz and converges to t*(SL8) as frequency decreases.

propagated with much less attenuation than was previously supposed (by some at least). Finally, the efficiency of the high frequency propagation may be highly variable from province to province and will be correlated with the extent and intensity of the low velocity zone.

Effects of Attenuation on Surface Waves

The Q models derived from body waves can be used to predict the attenuation of surface waves and vice-versa. Obviously combinations of surface and body wave data can be used to infer Q_α and Q_β as well, and either can be predicted from the results. Therefore the Q_α models discussed in the previous section can be used to infer Q_β (from a relation such as $Q_\alpha \approx 9/4\, Q_\beta$), and the resulting model can be used to predict, to first order at least, the expected surface wave attenuation. Lundquist's model is , in this regard, adequate for the prediction of the longer period surface wave attenuation (T > 40 sec) but is not well enough defined in the crust to give very accurate predictions for shorter period surface waves. Inasmuch as the 20 second period fundamental model Rayleigh wave, in particular, is of major interest in view of its use in M_s calculations, it is necessary to consider high resolution crustal Q models, such as are being obtained by Mitchell (1980). It seems sufficient here to only refer to Mitchell's work and to recall the general comments made earlier in the introduction. In particular, that a slight frequency dependence is inferred in Q_β, with the Q_β increasing with increasing frequency; that the mean Q_β in the crust is relatively high, and that regional dependence of the attenuation of surface waves in the range $5 \lesssim T \lesssim 40$ sec is small. All in all it does not appear that crustal surface wave attenuation is particularly difficult to deal with for purposes of M_s corrections, to obtain yield estimates and for discrimination.

More difficult, and much larger, are corrections in M_s for tectonic effects and probably for lateral variations in structure. Uncertainties in these latter effects completely overwhelm any correction uncertainties due to anelasticity effects.

Summary: State of Knowledge and Research Needs.

The essential conclusions of this report are:

(1) Frequency dependent Q models appear to be required in both the crust and upper mantle. Absorption band models, with the Q magnitude and frequency dependence varying with depth appear to be physically realistic and to satisfy the available data.

(2) A single absorption band appears to be appropriate for the entire mantle exclusive of the low velocity zone. Within the low velocity zone, when present, a second narrow absorption band appears to existand accounts for the increased attenuation and different frequency dependence of the attenuation in tectonic regions. This second absorption band is the likely mechanism for variablility of body wave absorption from region to region.

(3) High frequency seismic body waves propagate with relativly great efficiency from (and within) regions not having a well developed low velocity zone. Tectonic zones will typically absorb much more of the high frequency energy and this will generally result in lower m_b values. For this reason m_b should be measured "spectrally" (i.e., by narrow band filtering at 1 Hz with the first cycle of the P wave train selected) to avoid measuring m_b at different effective periods; and Q corrections should be made in order to account for differences in the regional Q structure.

(4) Surface waves in the 5-40 sec period range are not attenuated strongly by the crustal Q and there is no strong regional dependence in the attenuation in this period range. For longer periods there would,

however, be some fairly significant regional variations due to the variations in the low velocity zone. Because of the inferred high Q of the crust, especially for high frequencies, it is also implied that near regional range body waves (out 200 km or so from a source) will be weakly attenuated and high frequencies should be propagated efficiently in all cases. In regions with little or no low velocity zones (V_p), the range of efficient high frequency propagation could be much greater-perhaps out to 15° or greater.

Some research that could provide needed detail and better quantify the first order models so far obtained, includes:

(1) Simultaneous matching of explosion event body wave seismograms in the near, regional and teleseismic distance ranges with the objective of eliminating uncertainties in the source function, so that Q models could be obtain that were relatively free from trade-off problems with the source function.

(2) Use long period surface waves and high resolution analysis methods for station to station analysis of attenuation to obtain Q models that would be free from source trade-off problems. This approach would also give regional Q models.

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