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**BROADBAND SEISMOLOGY AND THE DETECTION AND VERIFICATION
OF UNDERGROUND NUCLEAR EXPLOSIONS**

by

Mark Andrew Tinker

A Dissertation Submitted to the Faculty of the

DEPARTMENT OF GEOSCIENCE

In Partial Fulfillment of the Requirements
For the Degree of

DOCTOR OF PHILOSOPHY

In the Graduate College

THE UNIVERSITY OF ARIZONA

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OF UNDERGROUND NUCLEAR EXPLOSIONS

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Final approval and acceptance of this dissertation is contingent upon
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Graduate College.

I hereby certify that I have read this dissertation prepared under my
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SIGNED: Maria Chela Jank

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My deepest gratitude to my advisor Dr. Terry Wallace, . . .

Nearly all men can stand adversity, but if you want to
test a man's character, give him power.

ABRAHAM LINCOLN

If there be any truer measure of a man than by what he
does, it must be by what he gives.

ROBERT SOUTH

Character is formed, not by laws, commands, and
decrees, but by quiet influence, unconscious suggestion,
and personal guidance.

MARION L. BURTON

a mentor, scholar, leader, teacher, scientist, provider of opportunities, and friend.

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One hundred years from now it will not matter what kind of car I drove
What kind of house I lived in
How much I had in my bank
Not what my clothes looked like.

One hundred years from now it will not matter what kind of school I attended
What kind of typewriter I used
How large or small my church
But the world may be a little better
Because I was important in the life of a child

MARGARET F. POWERS

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Dedicated to:

Kristen Male Tinker
My wife and closest friend

- and -

Olive Fast Wheeler
My grandmother, who fostered in me a love of Arizona and
the American Southwest, which in turn, had a profound initial
influence on my attending the University of Arizona in 1991.

December 12, 1902 - January 5, 1992
Grover Hill, Ohio - Tucson, Arizona

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ABSTRACT

On September 24, 1996, President Clinton signed the Comprehensive Test Ban Treaty (CTBT), which bans the testing of all nuclear weapons thereby limiting their future development. Seismology is the primary tool used for the detection and identification of underground explosions and thus, will play a key role in monitoring a CTBT.

The detection and identification of low yield explosions requires seismic stations at regional distances (<1500 km). However, because the regional wavefield propagates within the extremely heterogeneous crustal waveguide, the seismic waveforms are also very complicated. Therefore, it is necessary to have a solid understanding of how the phases used in regional discriminants develop within different tectonic regimes. Thus, the development of the seismic phases Pn and Lg, which compose the seismic discriminant Pn/Lg, within the western U.S. from the Non-Proliferation Experiment are evaluated.

The most fundamental discriminant is event location as 90% of all seismic sources occur too deep within the earth to be unnatural. France resumed its nuclear testing program after a four year moratorium and conducted six tests during a five month period starting in September of 1995. Using teleseismic data, a joint hypocenter determination algorithm was used to determine the hypocenters of these six explosions.

One of the most important problems in monitoring a CTBT is the detection and location of small seismic events. Although seismic arrays have become the central tool for event detection, in the context of a global monitoring treaty, there will be some dependence on sparse regional networks of three-component broadband seismic stations to detect low yield explosions. However, the full power of the data has not been utilized, namely using phases other than P and S. Therefore, the information in the surface wavetrain is used to improve the locations of small seismic events recorded on a sparse network in Bolivia. Finally, as a discrimination example in a complex region, P to S ratios are used to determine source parameters of the M_w 8.3 deep Bolivia earthquake.

CHAPTER 1

INTRODUCTION

HISTORICAL DEVELOPMENT OF A COMPREHENSIVE TEST BAN TREATY

The first nuclear explosion, codenamed TRINITY, was detonated on July 16, 1945, in a desert valley called Jornada del Muerto (walk of the dead) in central New Mexico. The Manhattan Project had come to fruition. Four years later, on August 29, 1949, the Soviets detonated "Joe-1," their first nuclear test. It was a copy of the plutonium implosion weapon dropped on Nagasaki built from stolen plans. The design was never put into production and was detonated solely as a demonstration of nuclear capability. Humans had discovered and unleashed the power of the atom.

Although the nuclear bomb was largely responsible for ending Japanese aggression in World War II, the arms race between the United States and the Soviet Union had begun, with both nations endorsing the philosophy that the best defense was an unstoppable offense. Nuclear testing, primarily atmospheric, climaxed in 1962 with 139 tests conducted by four countries (Fig. 1.1). This was largely in response to two factors: 1) the end of the 1958-1961 U.S. - Soviet moratorium, and 2) the upcoming Limited Nuclear Test Ban Treaty (LTBT). In the mid 1950s, concern was growing about stopping weapon development as well as the environmental impacts of radioactive fallout from atmospheric tests. This led to the development and signing of the LTBT in 1963. It was a trilateral agreement between the United States, Soviet Union, and Great Britain that prohibited nuclear testing in space, in the atmosphere, and under water. In fact, the key leaders of these three nations were in favor of a comprehensive ban. However, such a Comprehensive Test Ban Treaty (CTBT) would have to wait because seismology was perceived as inadequate for monitoring underground nuclear explosions.

Figure1.1 A graph showing the total number of nuclear explosions detonated per year by country. The right side of the box represents the total number of nuclear explosions detonated each year irrespective of country. The left side of the box represents the total number of nuclear explosions detonated by a single country (times a factor of 10) separated according to whether or not the explosions was detonated in the atmosphere (before the LTBT) or underground.

At the time of negotiations, only one underground test had been conducted (RAINIER, at the Nevada Test Site in September, 1957). Furthermore, there were fundamental questions about seismology: 1) seismology as a science was not wide spread; only limited global data existed and then, only in analog form making rapid assessment impossible; and 2) seismic stations were few and far between; furthermore, they had a limited range of frequencies and low dynamic range. Thus, seismology could not support a major arms control agreement. This realization launched new efforts in instrumentation and seismological research. Project VELA began in 1959 under the Advanced Research Projects Agency of the United States Department of Defense. A portion of VELA included a plan by the United States Coast and Geodetic Survey to build a World-Wide Standard Seismograph Network (WWSSN). The network quickly grew to approximately 125 analog stations in 31 countries, and the science of seismology blossomed. Furthermore, a separate entity of seismology focusing on the detection and discrimination of underground nuclear explosions came into being. This was prompted by the need for such a discipline and was spurred on by the increasing amounts of data from the continuance of underground tests.

The Threshold Test Ban Treaty (TTBT) was signed on July 3, 1974. It restricts the yield of nuclear tests conducted by the United States and the Soviet Union to be less than 150 kt. This yield was based on the level of shaking caused in Las Vegas by testing at the Nevada Test Site. The TTBT posed a challenge to seismology that was based not on detection but yield determination. Seismically determined yields are a function of the statistical methods used to estimate the "magnitude" and are dependent on the material in which the test was conducted. Thus, during this time period, improvements in monitoring focussed on reducing the uncertainty in yield estimation. Nonetheless, a TTBT was the next logical step following a LTBT in terms of both global disarmament and the seismological state-of-the-art.

Towards the end of the twentieth century the Soviet Union collapsed which virtually ended the cold war between the super-powers. In subsequent SALT (Strategic Arms Limitation Treaty) agreements, the world's two largest nuclear stockpiles began to diminish. Furthermore, both the United States and the new countries from the defunct Soviet Union also strived to limit testing and the development of new generations of nuclear weapons (vertical proliferation). The logical conclusion of these efforts would finally have been a CTBT. This treaty would encompass as many nations (nuclear or non-nuclear) as possible. However, it was simultaneously discovered that Iraq was aggressively pursuing nuclear weapons development. This changed the focus of a potential CTBT and presented a new challenge to seismology for it must now address the spread and/or development of nuclear weapons within non-nuclear states (horizontal proliferation). This task is far more difficult than focusing on one nation, and it is equally if not more important given the diverse nationalities in pursuit of nuclear arms. Geographically, these areas include the Middle East, Pakistan, and North Korea, and it is very likely in the future there will be new areas of concern. Countries may also attempt to test outside their own geographical borders. Further, in the mid 1990s, both declared nuclear nations China and France actively pursued nuclear testing. However, on September 24, 1996, The United States and all other nuclear nations of the world except for India signed the CTBT, thereby limiting the future development of nuclear weapons. According to President Clinton, this was, "the longest sought, hardest fought prize in arms control history." The 40 year battle to stop the testing of nuclear weapons was finally won.

ENFORCING A CTBT: A SEISMOLOGICAL PERSPECTIVE

To enforce the CTBT, seismology must now monitor the entire planet. Towards this end, one of the key questions is, what is the smallest event that can be both detected and identified?

Modern seismic instrumentation is limited only by the noise levels in the earth. It has a large dynamic range, records the full spectrum of usable frequencies, and these high quality, broad band stations number in the hundreds. Furthermore, the digital age allows all seismic signals to be recorded digitally and sent anywhere in the world in a matter of seconds. Thus, the data is of high quality and accessible in near real-time. The science of seismology in combination with the location and total number of stations is now the limiting factor .

As a monitoring tool, a seismic network must be able to both detect and identify the source of seismic signals. Unlike enforcing the TTBT, where a yield bordering at 150 kt is recorded by many stations due to its size (in other words, detection is not a problem), a CTBT must address the limits, or detection and identification thresholds, of a seismic network. Of course, the closer a station is to the source, the stronger and higher quality the signal. This is because the signal is strong enough to be well recorded above the background noise also recorded by seismometers. Furthermore, detection must occur at more than one station so that a location may be accurately estimated. Once a seismic event has been detected, it is then necessary to identify the source of the signal. Is it an earthquake, a legitimate chemical explosion, a rock burst, or a nuclear explosion? The smaller the event, the more sources of seismicity from which it must be distinguished. For example, a fully tamped 10 kt explosion at NTS has a seismic magnitude of approximately 5.1. At Novaya Zemlya, Russia's nuclear test facility, and Lop Nor, China's nuclear test facility, this magnitude is slightly higher (≈ 5.2) due to the material properties in which the explosion is contained.

There are an average of 7000 earthquakes a year that range in magnitude between 4.0 and 5.0. Further, there are tens of thousands more earthquakes per year with a magnitude less than 4.0 (Table 1.1). All of these events must be scrutinized in order to separate the

potential suspect nuclear explosion from the legitimate sources. This is not a trivial task. As a good rule of thumb, a larger source is recorded at more stations. Therefore, there is a higher confidence in its location (latitude, longitude, depth, and origin time) as well as a higher confidence in its identification. It follows that detection is easier than identification. Thus, the smallest event that can be detected with reasonable confidence, known as a detection threshold, is usually lower than an identification threshold. These thresholds can vary from region to region as a function of, for example, the number of recording stations, propagation characteristics, and source characteristics.

TABLE 1.1 Worldwide earthquake frequency by magnitude.

Seismic Magnitude	Number of Earthquakes per	
	Year	Day
4.5	2,300	6
4.0	7,100	19
3.5	19,000	52
3.0	68,000	186
2.5	209,000	572

Seismic Discriminants

Identifying a suspect event requires invoking what are called seismic discriminants. Seismic discriminants seek to exploit fundamental differences between earthquakes and explosions. For example, the most fundamental and useful seismic discriminant is the source depth. All underground nuclear explosions will be detonated within the uppermost crust; the deepest to date being 2.5 km. This is for two reasons. One, a depth greater than a few kilometers exceeds the ability of emplacement technology. Secondly, an underground explosion is detonated at a specific depth according to its predicted yield. This rubblelizes and collapses the overlying material so that all radioactive gasses are efficiently trapped. If there were to be a radioactive leak, not only is this a health risk, but it compromises any

clandestine intentions since the radioactivity is readily detectable. Although determination of source depth can be difficult and have a large uncertainty, 90% of all seismic events occur too deep to be unnatural, immediately reducing the number of events to be evaluated by other means. Further discrimination of the remaining 10% appeals to source phenomenology.

Explosion sources in isotropic, homogeneous media produce compressive energy that radiates symmetrically outward in all directions about a point. Therefore, intuitively, an explosion should produce no shear energy. On the other hand, earthquake sources are shear dislocations along a plane. Thus, the P and S energy from an explosion should be different than that from an earthquake. Furthermore, single hole explosion source-time histories are impulsive, resulting in a higher corner frequency, whereas earthquake source durations are considerably longer yielding a lower corner frequency. In other words, seismic waves from explosions have more energy at higher frequencies for a given "size." For example, Figure 1.2 is a spectral comparison between the Non-Proliferation Experiment (NPE, a ≈ 1.1 kt chemical explosion) and an earthquake at approximately the same magnitude and distance. The corner frequency of the earthquake (0.8 Hz) is considerably lower than that of NPE. These two concepts of body wave energy and spectral content form the foundation for most discriminants.

With the need to monitor small events, discrimination has recently focused on regional (<1500 km) techniques. However, smaller earthquakes have shorter (more explosion-like) source durations. Thus, regional discriminants usually focus on P and S energy at higher frequencies by evaluating different frequency or spectral bands. Spectral discriminants are based on the differences in the spectral content of regional phases between explosions and earthquakes. Seismologists utilize this phenomenon to compare not only the spectral content of certain phases but also to compare the spectral energy of different frequency bands of the same phase. Spectral ratios compare the difference in energy at different

Figure 1.2 A comparison of the spectral content of the Non-Proliferation Experiment (NPE) and a similar magnitude earthquake at approximately the same distance. NPE has a higher corner frequency (2.5 Hz) implying more high frequency energy while the earthquake has a lower corner frequency (0.8 Hz) and more long period energy (<0.1 Hz).

frequency windows of the same phase. For example, the ratio of the Pn spectrum at 1-2 Hz to that at 2-4 Hz is denoted as $P_n(1-2)/(2-4)$. Phase ratios compare the spectral energy between two phases of the same frequency window (P_n/L_g at 6-8 Hz), and cross spectral ratios compare different phases at different frequency windows ($P_n(4-6)/L_g(2-4)$). Regardless of the discriminant used, the ratio value is plotted compared to a field of other sources. Successful discriminants will separate explosion ratio values from earthquake ratio values.

International Data Center

To test the logistics of global monitoring, the Group of Scientific Experts Third Technical Test (GSETT-3) began full scale operations in January of 1995 at the prototype International Data Center (IDC) at the Advanced Research Projects Agency's (ARPA) Center for Monitoring Research in Arlington, Virginia. The purpose of this global seismological experiment was not only to develop rapid seismic monitoring for a CTBT as well as provide practical technical information for CTBT negotiators, but also to establish an international system that could evolve and adapt to support negotiated monitoring requirements. Thus, it would become an international data center geared towards global monitoring. The IDC incorporates four disciplines towards nuclear monitoring. After seismic, there is radionuclide, the detection of radioactive elements, hydroacoustic, and infrasound, which monitor the world's oceans and atmosphere respectively for the propagation of energy within each medium that may be produced by a nuclear blast.

BROADBAND SEISMOLOGY

Broadband seismology plays a key role in the enforcement of a CTBT, aspects of which will be investigated in the following chapters. First, the effect of a complex region consisting of many geologic terrains on the seismic phases Pn and Lg will be evaluated. This is accomplished using 50 broadband seismic stations in the western United States that

recorded the Non-Proliferation Experiment, a large (1 kt) chemical explosion. Subsequently, teleseismic locations are used to investigate France's nuclear testing facility in the Pacific by using a relative location algorithm to relocate 6 recent explosions. A regional location algorithm for broad band data that incorporate surface waves is then explored. Thus, the information from the full waveform is used while the dependence of phase travel times on a velocity structure is avoided. Finally, P to S ratios, analogous to most discriminants, are used to determine source parameters for aftershocks of a great, deep earthquake.

CHAPTER 2

REGIONAL PHASE DEVELOPMENT OF THE NON- PROLIFERATION EXPERIMENT WITHIN THE WESTERN UNITED STATES

INTRODUCTION

In September of 1996, a Comprehensive Test Ban Treaty (CTBT) was approved by the United Nations. This treaty will prohibit the testing of all nuclear weapons. Seismology is the primary tool used for the detection of underground explosions and thus will play a key role in enforcing compliance of a CTBT. However, to effectively monitor a CTBT, seismologists must be able to confidently detect and identify low yield explosions. This requires seismic stations at regional distances (<1500 km). The regional wavefield propagates within the extremely heterogeneous crustal and upper mantle waveguide, and thus, the seismic waveforms are very complicated. Therefore, it is necessary to have a solid understanding of how the phases used in regional discriminants (Pn and Lg) develop within different tectonic regimes and change as a function of distance. Using the large number of explosions and earthquakes at the Nevada Test Site, Walter et al. (1995) performed a thorough analysis of various discriminants and source effects. They demonstrated the effectiveness of certain regional discriminants by showing a good separation between earthquake and explosion populations. However, only two stations were in operation long enough to record the sufficient number of events necessary for their analyses. Furthermore, since all of their events were located within NTS, it was not necessary to correct for path effects. Thus, they were constrained to only two station-source paths. This allowed them to determine the effect of the source in evaluating the effectiveness of each discriminant. Regional phases such as Pn and Pg are very dependent

on regionally varying lithospheric structure. Therefore, the behavior of a regional discriminant may be very different for another area of the world. The identification of an unknown event recorded on a monitoring seismic array, if no prior explosion data exists for that array, would require comparing that event to other populations, such as the explosion and earthquake databases from the western United States. Without an understanding of path effects, such a comparison would be uncalibrated and non-definitive.

An experiment designed specifically to study the difference between chemical and nuclear explosions was detonated at the Nevada Test Site on September 22, 1993 (Denny, 1994; Hannon, 1994). This explosion, known as the Non-Proliferation Experiment (NPE), was a multi-fold experiment, including measurements of the hydrodynamic shock waves and free field particle velocities, accelerations, and stresses to measurements of electromagnetic pulses, gas samples, and atmospheric waves (Denny and Zucca, 1994). A large portion of the experiment focused on local and regional seismic surface measurements. The purpose of these include, 1) verifying possible differences between the seismic signature from a nuclear and chemical explosion, 2) investigating the evolution of regional seismic phases as they propagate across major tectonic provinces, and 3) determining the effects a change in geologic environments has on seismic discriminants.

The Southern Arizona Seismic Observatory at the University of Arizona participated in NPE by recording the explosion on an east-west, broad band seismic profile (Wallace et al., 1994). Ten broad band stations were installed on a line trending east from the Nevada Test Site to the Lawrence Livermore National Laboratory permanent seismic station KNB at Kanab, Utah. The purpose of this profile was to investigate the development of regional seismic phases and to explore the effects a change in geologic environments has on seismic discriminants. At the time of NPE, more than 30 permanent broad band stations were in operation in the western U.S. With the two portable broad band profiles (UA profile and

the profile operated by Lawrence Livermore National Lab (LLNL) and Cambridge University (McCormack et al., 1995)), NPE was the most seismically (broadband, 3-component) instrumented explosion in history. More than fifty regional broad band stations recorded the event, allowing not only for investigation of phase development at greater distances, but also as a function of azimuth (Fig. 2.1).

THE NON-PROLIFERATION EXPERIMENT

The United States Department of Energy conducted NPE in a large cylindrical chamber (15.2 meters in diameter by 5.5 meters in height) 390 meters under Rainier Mesa at NTS. The cylinder was filled with approximately 1.29 million kilograms of a 30/70 emulsion to ANFO (ammonium nitrate and fuel oil) blend (Denny and Zucca, 1994). This entire volume was detonated successfully at one minute past midnight on September 22, 1993, and recorded at over 50 broadband seismic stations.

The University of Arizona seismic profile consisted of 10 broad band STS-2 seismometers (frequency band 40 - 0.05 Hz), Reftek digitizers, and GPS clocks. Nine of the ten stations were operational during the explosion. The closest station to the shot point is 73 km; the remaining stations were located at approximately 20 km intervals beyond successive stations. Station KNB, the eastern most station in the profile, is 302 km from the shot point. The profile covers the transition between the Basin and Range and the Colorado Plateau (Station 1 is the furthest and Station 10 is the closest to the shot point, Figure 2.1). The remaining stations that recorded NPE are permanent stations operated by Incorporated Research Institutions in Seismology (IRIS)/Global Seismograph Network (GSN), Lawrence Livermore National Lab (LLNL), Sandia National Lab, TERRAScope, the Berkeley Digital Seismic Network (BDSN), and another portable broad band profile (trending 320° from NTS, 250km long) deployed by LLNL and Cambridge University, for a total of 50 stations. These stations range in distance from 73 km to 1029 km from NTS.

NPE PHASE DEVELOPMENT

Geologic setting

The southwestern U.S. is a tectonically complicated region, containing diverse and distinct geologic provinces. A large portion of the southwestern U.S. is occupied by the Basin and Range province, usually divided into northern and southern portions (Fig. 2.1). The northern Basin and Range has an average elevation approximately 1000 meters higher than the southern Basin and Range (Saltus and Thompson, 1995). The Basin and Range has a relatively thin crust and slower than average Pn velocities; the north has an average crustal thickness of approximately 30 km and an average Pn velocity of ≈ 7.8 km/sec, while the south has a more variable crustal thickness of 26 to 36 km and a slightly higher average Pn velocity of 7.8 - 8.0 km/sec (Hearn et al., 1991; for a complete review see Parsons, 1995). East of the Basin and Range is the Colorado Plateau, an area that has remained stable and intact during Tertiary extension and maintains an average elevation of ≈ 2 km. The plateau has a slightly variable ($\approx 15\%$) crustal thickness between 37 and 48 km (Parsons et al., 1996; Zandt et al., 1995), and Pn velocities range from 7.8 to 8.1 km/sec (Hearn et al., 1991; Beghoul et al., 1993). To the west of the Basin and Range is the Sierra Nevada. There has been considerable work done on the crustal structure of the Sierra Nevada, and the existence of a "crustal root" is still a subject of controversy. A recent experiment (Flidner et al., 1996) modeled the Sierra Nevada with a small root (crustal thickness ≤ 43 km) and extremely low Pn velocities (7.6 to 7.8 km/sec) (Hearn et al., 1991). The thickest portion of the Sierra Nevada crust (≈ 43 km) is just south of the raypath between NTS and station CMB as shown in Figure 2.1 (Flidner et al., 1996).

The Nevada Test Site (NTS) is located in the central Basin and Range of southern Nevada. Of the fifty stations that recorded NPE, approximately 26 are located within the Basin and Range (Fig. 2.1). The UA profile samples the Basin and Range (stations 10

Stations That Recorded NPE

Figure 2.1 A map of the western U.S. showing the location of the permanent broadband stations that recorded NPE (star) as well as the location of the two portable profiles (inverted triangles). The symbol of the stations correspond to the geologic terrain or category in which they are located. Triangle (regular or inverted) = Basin and Range Terrain; Circle = Peninsular Ranges of southern California; Diamond = Sierra Nevada Terrain; Square = Pacific coast area on the side of the Great Valley from NTS; Hexagon = stations located greater than 750 km from NTS.

through 4) and the westernmost Colorado Plateau (stations 3, 1, LDS, KNB). Along the UA line, the 50-75 km wide transitional zone observed between the Basin and Range and the Colorado Plateau to the north and south, appears to be absent (Zandt et al., 1995). Although these four Colorado Plateau stations (stations 3, 1, LDS, KNB) are shown in Figure 2.1 as part of the Basin and Range, the following analysis does not ignore it is part of a different geologic province. The remaining stations are located in the Sierra Nevada terrain or three other categories that have been grouped by ray path according to the provinces traversed or distance (Fig. 2.1). Seven stations form the Peninsular Range group located in southern California (circles in Fig. 2.1). The Peninsular Ranges is a complex transition region separating North America from the Pacific. Stations located along the west coast of California compose the Pacific Coast group. Energy arriving at these stations has traversed the Basin and Range, the Sierra Nevada, and the Great Valley of California (squares in Fig. 2.1). The final group represents stations located at a distance greater than 750 km from NTS.

Pn and Pg Velocities and Structure Associated With the Two Portable Profiles

The University of Arizona seismic profile (UA) recorded NPE at 9 stations. Station 2 was not operational. However, a fortunate happenstance was that at the same distance and nearly the same azimuth was permanent station LDS, operated by Sandia National Laboratory. The vertical component waveforms from the entire UA profile are shown in Figure 2.2. The Pg velocity determined from a least squares fit is 6.04 km/sec. The apparent Pn velocity is a slow 7.6 km/sec, expected on the downward-dipping structure associated with the crust thickening to the east entering the Colorado Plateau. The shaded region in the record section outlines the Lg velocity window between 3.55 and 3.0 km/sec. Lg does not clearly emerge until Station 6 at 166 km. The same Lg behavior is observed for the Northwest profile (Figure 2.3). The apparent Pn velocity for the Northwest profile is

Figure 2.2 A travel time curve of NPE with a reduced velocity of 6 km/sec, recorded by the University of Arizona portable seismic profile. The apparent Pg velocity (6.04 km/sec) and the apparent Pn velocity (7.58 km/sec) were determined by a least squares fit to the data. The grey area between 3.55 and 3.0 km/sec represents the Lg velocity window.

Figure 2.3 A travel time curve of NPE with a reduced velocity of 6 km/sec, recorded on the Northwest line (McCormack et al., 1994). The apparent Pg velocity (6.06 km/sec) and the apparent Pn velocity (7.85 km/sec) were determined by a least squares fit to the data. The grey area between 3.55 and 3.0 km/sec represents the Lg velocity window.

7.85 km/sec, and the Pg velocity of 6.06 km/sec (crustal average) is slower than the Pg velocity of the UA line.

Using a ray inversion code (RIC), we inverted the Pn and Pg times from both profiles and found the Moho dips downward to the east (UA line) and remains virtually horizontal to the northwest. We constrained the Moho to be flat and used a Pn velocity of 7.85 km/sec for the Northwest profile and 7.9 increasing to 8.0 km/sec for the UA profile. However, a flat Moho, or Moho represented by only one line segment, is not consistent with the seismic data or topography. Thus, to make the problem more realistic as well as to examine the non-uniqueness of the problem, we also inverted for a Moho structure consisting of 3 to 5 straight segments over the 400 km length of each model. This resulted in a crustal thickness under NTS of approximately 30 km. From NTS, out to 300 km on the UA line (the furthest distance sampled), the crust stepped downward to a thickness of 40 km. This is consistent with thicknesses determined from a receiver function analysis using the UA profile by Zandt et al. (1995). The Northwest line (nearly 370 km long) has a crust which is a reasonably constant thickness of approximately 30 km.

Lg and Pn Amplitudes and Attenuation

Lg is a complex crustal wavefield recorded on all three components of ground motion. It is usually interpreted as the superposition of a large number of higher mode Love and Rayleigh waves propagating in a crustal wave guide (Pomeroy, et al., 1982). Therefore, it is sensitive to both the vertical and lateral velocity variations within the earth's crust. To evaluate the phase Lg in the western U.S., we first attempted to remove the effects of attenuation and distance. Following the methodology of Chavez and Priestley (1986) we measured the distance decay of the spectral amplitudes of the Lg wave train recorded on the UA profile. We assumed that the Lg amplitude followed the relationship:

$$\log A(f,R) + \log R^{0.5} = \log S(f) - |(1.364f) / Q(f)| t \quad (1)$$

where A is the spectral amplitude observed at a distance R , f is frequency, t is the travel time, S is the source term, and Q is the quality factor. This is the equation for a straight line in $\log(S)$ and $1/Q(f)$. We know R and t for each station. We then determined the amplitude (A) at a fixed frequency (f) for all the stations and solved for S and $Q(f)$ using least-squares. In the above equation, the source term is the intercept and the Q term controls the slope. By measuring the spectral amplitudes over all frequencies, we obtained the source and Q spectra. Once Q is determined at each frequency, we then inverted for apparent Q as a function of frequency. This model does not consider attenuation due to scattering, thus, it provides only a measure of apparent Q , not intrinsic Q (Chavez and Priestley, 1986). For the UA line, we determined $Q_{Lg}(\text{vertical}) = 388 f^{0.55}$ as compared to $Q_{Lg}(\text{vertical}) = 217 f^{0.71}$ determined for the Northwest line by McCormack et al. (1994). This is reasonable given the greater structural integrity of the Colorado Plateau compared to that of the Basin and Range. This Q value was determined using stations east of and including Station 6, since L_g is not fully developed on stations 7 through 10. For the tangential component, we determined $Q_{Lg}(\text{tangential}) = 362 f^{1.08}$ and McCormack et al. (1994) determined $Q_{Lg}(\text{tangential}) = 180 f^{0.65}$ for the Northwest line. Finally, we determined Q for all the stations that recorded NPE in which we believe we could effectively convert to displacement. For these 43 stations in the multi-province southwestern U.S. we determined $Q_{Lg}(\text{vertical}) = 238 f^{1.28}$ and $Q_{Lg}(\text{tangential}) = 199 f^{1.16}$. These Q values are substantially lower than the 6 stations used from the UA line, because these Q values are averaged across the many geological provinces (Fig. 2.1) sampled by the travel path from NTS to the 43 stations.

There are large local variations in the amplitude of P_n across both profiles (Figs. 2.2 and 2.3). For example, there is a pronounced amplification of P_n energy at station 5 on the UA line (Fig. 2.2) and at station BRH, on the Northwest line (Fig. 2.3). A possible

explanation of this observation is the focusing and defocusing of Pn energy by undulations of the Moho (Zandt et al., 1995; Schultz et al. 1995). Starting with a simplified version of the crustal structure of the Northwest line (McCormack et al., 1994), we added three "undulations" to the structure of the Moho (Fig. 2.4). All three undulations (two negative, one positive) have an amplitude of only one kilometer. Using RIC, we traced on the Pn energy (Fig. 2.4). Even these small variations in the Moho can focus and defocus Pn energy over only tens of kilometers at the surface. Alternatively, Schultz et al. (1995) numerically model using reflectivity the amplified Pn at station BRH on the Northwest line. They show that when material contrasts are large enough, a local plane layered structure can result in significant Pn amplification. However, this site effect has a profound implication for lateral variations in structure since our stations have dense spacing (20 km); if the plane layered material contrasts were deep in the crust, then the Pn amplification could not be distinguished at one station over a seismic line with 20 km average spacing. Given what little is known of the geology at station BRH on the Northwest line, Schultz et al. (1995) produce amplified Pn arrivals using a low velocity volcanic layer either at the surface or a few hundred meters deep. Therefore, we prefer using topography on the Moho to explain amplified Pn arrivals that are not associated with local site effects.

Pn is not a simple geometric arrival, rather it is a combination of an effervescent wave and energy turning in the upper mantle. If it were a simple geometric ray, then the rays would turn in the upper mantle and be sensitive to the upper mantle velocity gradient. Therefore, both frequency dependent attenuation and geometrical spreading are affecting the Pn waveform with distance. We use the power spectral method of Chun et al. (1989) to determine a decay constant of the form $(\Delta^{-n(f)})$ (Δ = distance, f = frequency, n = decay constant). The displacement power spectrum of Pn ($P(f,\Delta)$) is related to a term containing

Figure 2.4 A plot of Pn ray paths through a simplified crustal structure (from McCormack et al., 1994) of the Northwest line. Three one kilometer undulations were added in random locations. Notice the focusing and defocusing of Pn energy at the surface occurs over distances of tens of kilometers.

all the source and receiver function information ($D(f)$) and decay constant in the linear equation

$$\log P(f, \Delta) = D(f) - 2n(f) \log \Delta. \quad (2)$$

In a similar fashion to solving for Q , the slope of $\log P$ versus $\log \Delta$ at a given frequency is $-2n$. Measurements over all the frequencies yields the decay constant "n" spectra. Fitting a line in a least-squares sense to the "n" spectrum yields the decay constant $\Delta^{-n(f)}$. McCormack et al. (1994) determined a decay constant of $\Delta^{-(1.84 + 0.48 f)}$ for the Northwest line. For the UA line we determined the Pn decay constant to be $\Delta^{-(1.34 + 0.19 f)}$ (1-10Hz). For the 38 stations in the southwestern U.S. in which the Pn phase could be confidently picked, we determined a Pn decay constant of $\Delta^{-(1.29 + 0.05 f)}$ for the frequency window of 1 - 6 Hz. At higher frequency bands, the range in data values was too large to effectively measure the decay.

SPECTRAL DISCRIMINANTS

Spectral discriminants quantify the differences in the spectral content of regional phases of explosions and earthquakes. The full seismic waveform from an explosion may have more energy at higher frequencies (>0.5 Hz) than an earthquake for a given "moment." For example, Figure 2.5 is a spectral comparison between NPE and an earthquake of approximately the same magnitude and distance. The corner frequency of the earthquake (0.8 Hz) is considerably lower than that of NPE. However, many NTS explosions show less high frequency energy than nearby earthquakes as shown in comparisons of spectral ratios (Taylor et al., 1988, Walter et al., 1995).

Phase discriminants (for example, Pn/Lg (6-8 Hz)) and cross-spectral discriminants (for example, Pn (4-6 Hz)/Lg (2-4 Hz)) seek to exploit the differences in P and S energy in explosions and earthquakes. Theoretically, explosions in a isotropic, homogeneous media produce compressive energy that radiates symmetrically outward in all directions about a

Figure 2.5 A comparison of the spectral content of the Non-Proliferation Experiment (NPE) and a similar magnitude earthquake at approximately the same distance. NPE has a higher corner frequency (2.5 Hz) implying more high frequency energy while the earthquake has a lower corner frequency (0.8 Hz) and more long period energy (<0.1 Hz).

point. Therefore, intuitively, an explosion should produce no S waves while having a strong P-wave first arrival. Earthquakes, on the other hand, are shear dislocations along a plane and produce strong S-wave energy. Although S-wave energy is commonly recorded from explosions, the P and S energy from an explosion is different than that from an earthquake. Thus, P/S ratios can discriminate between earthquakes and explosions, especially at higher frequencies (Walter et al., 1994). Although NPE was a chemical explosion, as discussed later, it is seismically indistinguishable from nuclear explosions. Thus, it is valid to use the expansive NPE data set to evaluate the effect of path on seismic discriminants.

With the advent of broadband seismology, discrimination techniques are able to incorporate wide spectral bands. Furthermore, using regional, high frequency phases (0.5-15 Hz) allows for the detection and possible discrimination to smaller magnitudes. For example, Taylor et al. (1989) and Walter et al. (1995) explored the effectiveness of various spectral ratio discriminants for numerous earthquakes and explosions located at the NTS and recorded on the Lawrence Livermore National Laboratory (LLNL) seismic network (4 stations). However, any technique that incorporates regional phases such as Lg, Pg, and Pn also requires an understanding of path effects (propagation through different geologic terrains) and distance dependence. Walter et al. (1995) explored the effect of source medium properties on spectral ratio performance since the source-station paths from the multiple sources at NTS were virtually the same for each of the two LLNL stations they used (MNV, KNB). They found that the source medium properties of gas porosity, strength, and depth significantly affected the Pg/Lg and spectral ratio discriminants.

To effectively monitor a Comprehensive Test Ban Treaty, a thorough understanding of both source and path effects on the applied discriminant is paramount if the discriminant is to be transportable, since most other areas of the world will not have the number of events

or stations found in the western U.S. with which to calibrate discriminants. With the large number of stations that recorded NPE, we can examine the effect of different terrains on spectral ratios. We first explore the spectral ratios of Lg/Lg and Pn/Pn to ascertain the contribution of each towards the phase ratio Pn/Lg .

Lg/Lg

As a discriminant, Lg is usually measured in a frequency window between 0.5 to 10 Hz. After removing the effects of distance and attenuation, we discovered that many of the Lg/Lg ratios as a function of distance exhibit scatter. For example, in Figure 2.6 there is a plot of the spectral ratio $Lg(4-6)/(2-4)$ as a function of distance. Although this ratio is not commonly used in discrimination literature, it effectively demonstrates the following; the observed scatter is a function of both the geologic terrain or region in which the station is located in some sort of combination with the path traveled. The scatter is limited to about one order of magnitude. Thus, using Lg in a discriminant would influence the separation observed between earthquakes and explosions based on what station was used. The horizontal lines in the ratio value vs. distance plot in Figure 2.6 represent the average and one standard deviation of the mean of the ratio values of all stations less than 600 km away from NTS. Shown on the map in Figure 2.6 are the stations corresponding to the ratio values greater than one standard deviation (white triangle); greater than the average (white circles); less than the average (black circles); and less than one standard deviation (black triangles). Clearly, there are some geologic factors contributing to the scatter. To explore this further, we separated the 50 stations into five terrains or categories (ratio value vs. terrain plot in Figure 2.6). The first category or terrain (triangles) represent stations within the Basin and Range terrain. The circles, diamonds, and squares represent the Peninsular Ranges, the Sierra Nevada terrain, and the Pacific coast, respectively (see also Fig. 2.1). Stations further than 750 km are plotted as hexagons. We observe a separation of the ratio

Figure 2.6 A map of the western U.S. showing station locations that correspond to specific spectral ratio values of the ratio $L_g(4-6)/(2-4)$. In the upper right are two plots of the ratio values. One plot shows the log of the ratio values versus distance. The symbols equate to the terrain of the corresponding station as shown in Figure 1 but now with the addition of shades of grey: dark triangles = Basin and Range; light circles = Peninsular Ranges of southern California; light grey diamonds = Sierra Nevada; medium grey squares = Pacific Coastal region on the other side of the Great Valley from NTS; white hexagons = stations located greater than 750 km from NTS. The horizontal lines represent the mean and one ($\pm$) standard deviation of the mean. The length of these lines span a distance range that represents the stations used to determine the average. Stations outside this distance range are plotted for reference but were not used to calculate the average. On the map, values greater than one standard deviation are shown as white triangles. Values greater than the mean but less than one standard deviation are shown as white circles. Values less than the mean but greater than one standard deviation (-) are shown as light black circles. Values less than one standard deviation (-) are shown as black triangles. The rings approximate the distance from NPE in kilometers. The geologic terrains are outlined in white. The second plot in the upper right is a terrain plot showing the ratio values relative to terrain so that the effect of terrain may be more easily evaluated. Notice that the ratio values of the stations in the Basin and Range are distinctly lower than the other stations

values based on terrain. Ratio values for stations within the Basin and Range cluster at a lower value than for stations outside of the Basin and Range.

P_n/P_n

P_n is also frequently used as a discriminant within some portion of the 1 - 10 Hz frequency band. As mentioned earlier, P_n amplitude can be markedly affected by Moho topography causing focusing and defocusing of energy. To quantify this for all stations, we compared the amplitude of P_n (1- 6 Hz) to the decay constant of $\Delta^{-(1.29 + 0.05 f)}$. The plot in Figure 2.7 shows the proportionality constant relating the observed P_n amplitude and expected P_n amplitude according to the decay constant as a function of distance. The horizontal lines in the plot represent the average and one standard deviation of the mean of the values. Shown on the map (Fig. 2.7) are the stations and their corresponding proportionality values relative to the mean. Therefore, a value greater than the mean represents amplified P_n energy and values less than the mean represent diminished P_n energy. For example, the amplified P_n energy at station 5 on the UA line (Fig. 2.2) and BRH5 on the northwest line (Fig. 2.3) can be seen (white triangle and white circle, Fig. 2.7). Clearly, this factor alone will contribute to complexities in using P_n as a discriminant. Thus, we compared in Figure 2.8 the spectral ratio P_n (2-4)/(4-6) as a function of azimuth. The ratio values span slightly more than one order of magnitude. Along the portable profiles, P_n is clearly affected by azimuth; the UA line has higher than average ratio values (white in Fig. 2.8) and the Northwest line has lower than average values (black in Fig. 2.8). Since these seismic profiles had variable P_n amplifications (Fig. 2.7), the P_n path must have an effect. In other words, P_n is influenced by Moho structure. Furthermore, for all the stations, there is no separation in the ratio values based on terrain; all the terrain values plot within the scatter of the Basin and Range values. Thus, the

Figure 2.7 A plot of the proportionality constant between the Pn energy observed and that expected from the determined decay constant vs. distance. The horizontal lines represent the mean and one ($\pm$) standard deviation of the mean. On the map, values greater than one standard deviation are shown as white triangles. Values greater than the mean but less than one standard deviation are shown as white circles. Values less than the mean but greater than one standard deviation (-) are shown as black circles. Values less than one standard deviation (-) are shown as black triangles. Thus, the high values (white triangles) represent amplified Pn energy and the low values (black triangles) represent diminished Pn energy. The rings approximate the distance from NPE in kilometers.

Pn Energy off Moho

Figure 2.8 The same plot as Fig. 2.6 except for the spectral ratio of $P_n(2-4)/(4-6)$ is plotted as a function of azimuth.

contribution of Pn energy is a function of Moho structure, which correlates with azimuth, and though not observed in the ratio values, is associated with geologic terrain.

Pn/Lg

As seen in Figure 2.9, the cross-spectral ratio Pn (1-2) / Lg (2-4) values are plotted as a function of distance. There is a reasonable separation in the ratio values between those stations inside the Basin and Range compared to other terrains (Fig. 2.9). However, there is also more scatter in these values compared to the first two examples (Figs. 2.6 and 2.8) with this particular phase ratio spanning two orders of magnitude. Thus, when comparing the ratio value to a preexisting earthquake and explosion population, the station used would markedly influence the amount of separation observed. On average, stations located outside the Basin and Range tend to have higher ratio values than those stations located within the Basin and Range.

Hunter's Trophy Phase Comparison

The source spectrum of NPE is approximately twice that for a nuclear explosion of similar yield and detonated in similar material (Goldstein and Jarpe, 1994; Denny et al., 1996). This occurs due to the fact that chemical explosions are much more efficient with coupling energy into the ground (Denny et al., 1996). However, in a discrimination setting, yields will not be reported and must be determined seismically or via other instrumentation. Thus, the scaled source spectra of NPE to a similar nuclear explosion were analyzed and found to be virtually identical (Denny et al., 1996) especially at higher frequencies (Goldstein and Jarpe, 1994). At local and regional distances, the waveform data between NPE and similar nuclear explosions is remarkably similar (Smith, 1994; Walter et al., 1994).

On September 18, 1992, almost one year to the day earlier than NPE, a low yield ($m_b \approx 4.4$) nuclear device named HUNTER'S TROPHY was detonated in N-tunnel at NTS. The

Figure 2.9 The same plot as Fig. 2.6 except for the cross-spectral ratio of $P_n(1-2)/L_g(2-4)$.

location was only 500 meters away from the NPE detonation. Since the locations and magnitudes of HUNTER'S TROPHY and NPE are virtually the same, these two events are ideal for comparison between nuclear explosion and chemical explosion waveform signatures. Fourteen permanent stations recorded these two explosions; Pn at nine stations and Pg at five stations. The Pn (Fig. 2.10) and Pg (Fig. 2.11) broadband waveforms not only show remarkable similarity in terms of shape but amplitude as well. The ratio value indicated in each box is the ratio of the first swing of P from HUNTER'S TROPHY (HT) to NPE. The Pn mean ratio for the 9 stations is 0.99, indicating the amplitude for the two events is virtually the same. Similarly, the Pg mean ratio value is 1.45. There appears to be no correlation of ratio value with distance or azimuth. Furthermore, Walter et al. (1995) performed a thorough analysis of short-period phase and spectral ratio discriminants for 130 nuclear explosions, NPE, and 50 earthquakes that occurred at NTS and were recorded on two LLNL stations. Regardless of the discriminant, NPE always plotted within the explosion populations. Thus, because the seismic difference between a single-hole chemical explosion and similar yield nuclear explosion is small, chemical explosions could be used to seismically calibrate other regions of the world.

DISCUSSION

With the need to monitor small events, discrimination has recently focused on regional techniques. However, smaller earthquakes have shorter (more explosion-like) source durations. Thus, regional discriminants usually focus on P and S energy at higher frequencies where fortunately, they are the most effective. Regional spectral discrimination has had a long history (Bakun and Johnson, 1970; Bennett and Murphy, 1986; Taylor et al., 1989). Taylor et al. (1988) explored the spectral ratios $((1-2)/(6-8))$ of Pg, Pn, and Lg for explosions and western U.S. earthquakes. Walter et al. (1995) studied the same spectral ratios, as well as the phase ratio Pn/Lg. They observed a good separation of the

Pn Comparison of HT and NPE

Figure 2.10 A map showing the locations of the broadband stations that recorded Pn from both HUNTER'S TROPHY (HT) and NPE. The white boxes show the Pn waveform generated by each explosion. The ratio value represents the ratio between the amplitude of the first swing of Pn from HT to the amplitude of the first swing of Pn from NPE.

Pg Comparison of HT and NPE

Figure 2.11 A map showing the locations of the broadband stations that recorded Pg from both HUNTER'S TROPHY (HT) and NPE. The white boxes show the Pg waveform generated by each explosion. The ratio value represents the ratio between the amplitude of the first swing of Pg from HT to the amplitude of the first swing of Pg from NPE.

explosion and earthquake populations located in and near NTS. However, the power of these techniques lies within the path calibration. In these cases, the only stations used to evaluate the discriminants were the four LLNL stations located around NTS (approximately 220 km to 400 km distance). As stated earlier, this is because these are the only stations in operation long enough to provide a sufficient catalog. Thus, the major difference in the waveforms of Walter et al. (1995) is the result of source and not the result of path and hence, the effectiveness of each discriminant can be evaluated. Taylor et al. (1988) applied a simple first order distance correction to their data set. However, to monitor a CTBT, regional discriminants will be applied to data recorded at regional distances. Thus, path calibration must be derived from earthquakes if the area is seismically active. If it is also to be compared to a population of events recorded elsewhere, such as NTS, it will be necessary to remove path effects based on the geologic terrain traversed.

A promising regional discriminant, P_n/L_g , incorporates the higher frequencies of P and S energy. However, we observe variations in the P and S energy as a function of azimuth and distance. As seen in Figure 2.6, the lower ratio values of $L_g(4-6)/(2-4)$ (triangles on the ratio value plot) correlate with stations located in the Basin and Range. For stations located outside the Basin and Range, the ratio values are considerably higher (circles, diamonds, squares, and hexagons). Is this truly a terrain effect? Because NTS is located within the Basin and Range, stations located in other terrains are farther away. Thus, it is possible we are not observing a terrain effect but a distance effect, implying that we may not have effectively removed geometrical spreading. However, as seen in Figure 2.6, there is clear separation between stations within the same distance range (≤ 400 km) based on terrain. Furthermore, station TUC in Tucson, Arizona, is located ≈ 730 km from NTS. This distance is nearly double that of the next farthest station from NTS yet still located in the

Basin and Range. The ratio value for TUC plots in the center of the range of Basin and Range values. Thus, it appears L_g is affected by terrain, not azimuth or distance.

The phase P_n is markedly affected by azimuth. The two portable profiles, both of which are primarily located within the Basin and Range (stations 1-3 from the UA profile are in the Colorado Plateau), show consistently different values for the ratio $P_n(2-4)/(4-6)$; the UA profile has higher ratio values than the Northwest profile (Fig. 2.8). The UA profile in Figure 2.12 was simplified from Zandt et al. (1995), and the Northwest profile was created from the simplified structure discussed earlier. P_n energy samples deeper into the mantle for the UA profile than for the Northwest profile. Thus, it is possible that the higher frequencies are attenuated more under the Colorado Plateau resulting in the higher ratio values shown in Figure 2.8. Furthermore, with the exception of the UA profile, all stations with a ratio value above the average (white triangles and white circles) appear to be randomly distributed. In other words, the high ratio values have no correlation with distance, azimuth, or terrain. However, without exception, these stations all correspond to a diminished P_n arrival shown as black circles or triangles in Figure 2.7. In other words, a station with an above average $P_n(2-4)/(4-6)$ value has a below average P_n proportionality constant, implying that at these stations, $P_n(4-6)$ decreased more than $P_n(2-4)$ relative to other stations. Thus, where P_n is diminished, the longer periods are attenuated less.

In summary, P_n is a difficult phase. Its energy is not directly affected by regional crustal velocities associated with geologic terrain as seen in Figure 2.8, where the ratio values all plot within the spread of the Basin and Range. However, it can be affected by geologic terrain in terms of Moho depth and structure. We note that the focusing and defocusing of P_n energy shown in Figure 2.7 is only applicable for sources at NTS. In other words, these same stations may not experience amplification at different back

Figure 2.12 Pn ray path from NTS along the UA line from Zandt et al. (1995) (top). Pn ray paths from NTS along the Northwest profile using a simplified structure from McCormack et al. (1994) (bottom). Notice the Pn energy samples deeper into the mantle along the UA line.

azimuths since the Moho path is different. Therefore, a single station cannot be tagged solely as amplified or diminished; it is also a function of back azimuth.

The different frequency band measurements of the phase ratio P_n/L_g do not readily produce azimuthal or terrain station patterns. However, this is to be expected since the ratio P_n/L_g is a product of both a phase influenced by geologic terrain (L_g) and a phase influenced by Moho structure and azimuth (P_n). Thus, these factors must be accounted for when interpreting the ratio value of each station in, for example, Figure 2.9. There is a reasonable separation of P_n (1-2) / L_g (2-4) values based on terrain as seen in the upper right of Figure 2.9. Stations located within the Basin and Range have a smaller ratio value than stations located within the other terrains. This phenomena is the effect of the L_g contribution. The P_n contribution is more difficult to analyze. Depending on which frequency window of P_n is used, the P_n contribution tends to add scatter to the ratio values. In other words, the ratio values of the terrains tend to be less clustered and exhibit more overlap (Fig. 2.9).

What are the implications of using regional discriminants? If NPE was a suspect event that was recorded by a few stations, then the ratio value at those stations would be compared either to other known explosions/earthquakes recorded at those same stations and preferably along the same path (in other words, calibrated paths) or to an earthquake/explosion dataset generated with other stations from some other region of the world. This comparison would be used to determine if the event was an explosion or an earthquake. Preferably, and realistically, the earthquake catalog for the seismic stations that recorded the suspect event would contain earthquakes from approximately the same region with which to make direct comparisons and path evaluations. However, as demonstrated in this paper, the uncertainty would be large without an extremely thorough data set. Nonetheless, the earthquake/explosion dataset from NTS is the best in the world and

should be utilized. Thus, cross-comparisons between a known source-station (NTS and station KNB, for example) and the suspect source and station located in a different region of the world is inevitable. To avoid potential mis-identifications, path effects, site effects, and source media from both datasets (NTS and suspect) should all be considered so that comparisons truly evaluate source differences.

CONCLUSIONS

Regional discriminants are necessary to monitor low yield explosions. However, regional phases can be very complicated due to path effects, thus affecting transportability of discriminant techniques from area to area. We find values of the discriminant P_n/L_g to behave according to both the geologic terrain and the large (gross orientation) and small (topography) scale Moho structure of the path traversed. Specifically, $P_n(1-2) / L_g(2-4)$ has values nearly one order of magnitude less for stations located within the Basin and Range than stations located outside the Basin and Range. The spread of the L_g/L_g and P_n/P_n values is within one order of magnitude while the $P_n(1-2) / L_g(2-4)$ values cover 1.8 orders of magnitude. The increased scatter in the $P_n(1-2) / L_g(2-4)$ values within terrains, as well as the increased overlap between terrains, is due to the P_n contribution. Nonetheless, the P_n/L_g values still separate according to terrain due to the L_g contribution. If the luxury of a calibrated path does not exist, then these path phenomena must be accounted for when evaluating a suspect event.

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CHAPTER 3

THE SIX RECENT FRENCH NUCLEAR EXPLOSIONS: A RELATIVE DETERMINATION OF LOCATION, MAGNITUDE, AND YIELD USING OPEN BROADBAND STATIONS

INTRODUCTION

On September 24, 1996, President Clinton signed the Comprehensive Test Ban Treaty (CTBT) which will prohibit the testing of nuclear weapons. Enforcement of a CTBT requires monitoring the entire planet for potential evaders; however, this is not a trivial task. To address this, the Group of Scientific Experts Third Technical Test (GSETT-3) began full scale operations in January of 1995 at the prototype International Data Center (IDC) at the Advanced Research Projects Agency's (ARPA) Center for Monitoring Research in Arlington, Virginia. The purpose of this global seismological experiment was to develop an international system that could evolve into the International Seismic Monitoring System for the CTBT. Thus, it would become an international data center geared towards global monitoring.

Over 90% of the stations reporting to the GSETT-3 are part of the Incorporated Research Institutions for Seismology (IRIS) Global Seismographic Network (GSN) or are part of other "open" networks which are accessible to the academic community in near real time. A complimentary experiment to GSETT-3 is the "routine" processing of seismic waveforms to determine, for example, location and magnitude (see <http://www.cdcdc.org/> on the world wide web). The majority of these waveforms are available in near real time through the IRIS Data Management Center. Presently, groups like the Harvard Centroid Moment Tensor project utilize these data to provide rapid earthquake source parameters.

Similar projects can be designated to "monitor" specific regions of the world or seismic phenomena other than earthquakes.

After a four year moratorium, France resumed their nuclear testing program in the Pacific by conducting six tests during a five month period that began in September of 1995. Newly elected president Chirac stated that France needed to test the reliability of their nuclear stockpile in a one-time-only series of tests. Furthermore, he also stated they were in full support of a CTBT. Other than China's active testing program, there had been no nuclear tests since September of 1992 when the United States conducted its last test codenamed DIVIDER ($m_b = 4.4$, September 23, 1992). Thus, the French explosions provided a unique opportunity to evaluate the waveform products from open stations and automated retrieval mechanisms such as the IRIS Spyder™ system.

DATA

We obtained broadband data from the affiliated networks of IRIS (GSN, TERRAScope, GTSN) using the IRIS Spyder™ and X-retrieve interfaces. We also retrieved data for seven stations from the Berkeley Digital Seismic Network (BDSN) world wide web homepage. A total of 17 teleseismic stations were used to determine magnitudes (m_b), and a total of 24 teleseismic stations were used to determine hypocenters (Fig. 3.1). We employed a simple technique to determine the relative magnitudes (m_b) and used the Joint Hypocenter Determination (JHD) software of Dewey (1971, 1983) to determine relative hypocenters. All waveform data were from open stations and can be retrieved in near real time by anyone with access to the internet. Furthermore, this open data can easily be used by anyone to determine locations and magnitudes.

Figure 3.1 A map showing the great circle paths from the French testing site (star) to all the stations used in this study.

MAGNITUDE, YIELD, AND LOCATION

Both the determination of magnitudes and locations are relative to a reference or "master" event. We chose the explosion on October 1, 1995, as the master event, because it was recorded by the most stations, which implied it would have the best "location." Further, it most likely had the largest magnitude and thus, yield. The seismically determined coordinates reported by the IDC located this explosion directly on Fangataufa, a Pacific seamount in French Polynesia that had been used in other French nuclear tests. A comparison of waveforms (1 Hz) recorded at station CMB from the six explosions is shown in Figure 3.2. The y-axis scale is the same for all six records; the variation in amplitude is a function of magnitude (yield). Although there is remarkable similarity among the P-wave arrivals, there are slight differences that we attribute to variations in source depth.

Body wave magnitude is based on the actual ground motion amplitude (A) of the first few cycles of the P wave corresponding to a specific period (T). The amplitude and period are logarithmically related to m_b by

$$m_b = (\log A/T) + Q(h, \Delta) \quad (1)$$

where $Q(h, \Delta)$ is a correction for distance and depth. However, we are interested in the change in magnitude (ΔM) from a master event. Therefore, if we assume the same hypocentral location, at any particular station

$$\Delta M = \log (A_m/T_m) - \log (A/T) \quad (2)$$

where the subscript "m" signifies the master event. We limited the frequency content of the signal to 1 Hz. Thus, $T_m = T$, and equation (2) reduces to

$$\Delta M = \log (A_m / A). \quad (3)$$

We used 17 stations that all recorded the master event on October 1, 1995, and at least three of the other explosions. The November explosion was recorded by the fewest number of

Figure 3.2 A comparison of P-waveforms from the six explosions recorded at station CMB in California. All traces have been high passed filtered at 1 Hz and are plotted at the same y-axis scale. Thus, variation in amplitude is a function of magnitude (yield). of the first swing of Pn from HT to the amplitude of the first swing of Pn from NPE.

stations, because it was one of the smallest in magnitude. As reported by IRIS, the October 1, 1995, master event had a magnitude of 5.5. We determined the mean and one standard deviation of the change in magnitude determined at all stations for each event (Table 3.1).

TABLE 3.1. The explosion epicenter, origin time, magnitude, and estimated yield

Date	Julian Day	Origin Time	Latitude (deg)	Longitude (deg)	Magnitude (m_b)	Yield (kt)
Sep 5	(248)	21:29:59.61	21.894 S	138.760 W	4.78 $\pm$.2	$\approx$ 3
Oct 1*	(274)	23:29:59.40	22.25 S	138.76 W	5.5	$\approx$ 25
Oct 27	(300)	21:59:59.72	21.883 S	138.994 W	5.41 $\pm$.1	$\approx$ 19
Nov 21	(325)	21:29:59.80	21.917 S	138.952 W	4.82 $\pm$.2	$\approx$ 3
Dec 27	(361)	21:29:59.91	21.840 S	138.890 W	5.00 $\pm$.2	$\approx$ 5
Jan 27	(027)	21:29:59.37	22.270 S	138.750 W	5.37 $\pm$.1	$\approx$ 17

*October 1 is the master event

Seismically determined yields can be very difficult to estimate. Although the determination of magnitude/yield relationships is straightforward given the magnitude of explosions of known yield, the number of reported yields is small. Furthermore, most of these reported yields are for explosions conducted at the Nevada Test Site. It has been established that yields determined seismically are a function of the source medium and the travel path between source and receiver. The NTS magnitude/yield relationships can be off by as much as a factor of 2 when applied to other source regions (Douglas and Marshall, 1996). Generally, the source media can be divided into two end-member groups. Good coupling, which is usually obtained in hard or water saturated rocks, and poor coupling, which usually occurs in dry or porous rocks. Assuming the islands to be water saturated basalts, we used the following relationship derived from media associated with good coupling

$$m_b = 4.45 + 0.75 \log (Y) \quad (4)$$

where m_b is the body wave magnitude and Y is the yield (Douglas and Marshall, 1996). From this, we determine the yield of the October 1, 1995, reference event ($m_b = 5.5$) to be 25 kt. The remaining yields are given in Table 3.1.

The yield relation in (4) should be considered a minimum estimate for two reasons. First, the coupling for this source region may be lower than the worldwide average. Secondly, the magnitude reported by IRIS for the October 1, 1995, master event may be low. If either of these circumstances are valid, the yields reported in Table 3.1 could increase by a factor of two.

We used the Joint Hypocenter Determination (JHD) method to determine explosion locations (Dewey, 1971, 1983). This method simultaneously calculates all the hypocenters relative to the October 1, 1995, reference event. Thus, the independent effects of travel time anomalies introduced by the different travel paths to the recording stations will be statistically minimized, resulting in more accurate locations relative to the October 1, 1995, reference event. We only used P-wave arrival times, because the onset time for S waves from explosions are less clear. A total of 24 stations located at teleseismic distances ($<100^\circ$) were used. Sixteen of the stations recorded at least 4 of the explosions, while the remaining 8 stations recorded an average of two explosions. The epicentral coordinates and origin times are given in Table 3.1. The epicenter and error ellipses for the explosions are shown in Figure 3.3. All French nuclear tests in the Pacific must occur on seamounts if they are to be underground and contained. Thus, the locations should correlate with these islands.

DISCUSSION

France has had an active nuclear testing program; only the United States and the former Soviet Union have conducted more tests. France conducted the last of approximately 15 nuclear tests in Algeria in early 1966. By late 1966, they had conducted their first test on

The Recent French Nuclear Explosion Locations

Figure 3.3 A map of the French testing site in the Pacific. The contour lines represent the -2000 m, -3000 m, and -4000 m bathymetric contours. The location of the master event (October 1, 1995) is shown by a black star. The relative locations of the remaining 5 explosions are indicated by their error ellipses and numbered chronologically. Recall that Mururoa and Fangataufa are atolls. Therefore, the only "dry land" is a ring of coral as indicated in black. The interior portion of the "island" is living coral under shallow water.

the Pacific island of Mururoa in the Tuamotu Archipelago, a portion of French Polynesia (Fig. 3.4). The French have conducted at least 110 tests on Mururoa, with magnitudes ranging from 4.0 to 5.9 (Lawson, 1994). They have also conducted approximately 8 tests on the neighboring island of Fangataufa (Lawson, 1994) including the first underground explosion at their Pacific test site on June 5, 1975 (Dahlman and Israelson, 1977). Up until that point, many of the early French Polynesia tests (≈ 35) were above ground. Mururoa and Fangataufa are atolls, seamounts ringed in coral that constantly grow upwards, striving for the sunlight and warmer water necessary for their survival, as the extinct volcano beneath them slowly sinks with the aging seafloor.

An island in the middle of the ocean is a "worse case scenario" for seismic monitoring. Almost without exception there are no seismic stations within regional (< 1500 km) distances. Stations that are less than 1500 km away are usually located on an island themselves and are "noisy." High seismic noise levels are created by oceanic waves striking the island. The high noise levels reduce the detection capability. Thus, waveforms from mid-oceanic explosions are usually only well recorded at teleseismic distances.

Using teleseismic data from open seismic stations we determined that four of the explosions occurred on the island of Mururoa and the remaining two explosions occurred on Fangataufa. The global array of open seismic stations allows monitoring down to small magnitudes; thus, relative magnitude and location determination is not limited to explosions recorded at selective stations. Anyone with access to the internet and moderate computing facilities can determine relative magnitudes and locations for earthquakes as well. Via the internet we were able to obtain a photograph (Fig. 3.4) of Mururoa taken by the six person crew of the space shuttle Endeavor just 23 days after the last French nuclear test (December 27) on that same seamount (<http://images.jsc.nasa.gov/html/earth.htm>).

Figure 3.4 A photo of Mururoa taken by the crew of the space shuttle Endeavor during the 74th shuttle mission (STS-72). The photo was originally taken in color at an altitude of 164 nautical miles on January 19, 1996, 23 days after the last French test on that same seamount.

The rapidly growing internet provides a phenomenal amount of information that is easily and quickly obtainable. In terms of seismic monitoring, the information era fosters the concept of a "neighborhood watch," where seismic data is exchanged freely and efficiently. Open access combined with the routine global monitoring of the GSETT-3 experiment provides the backbone of a seismic system which is difficult to evade.

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CHAPTER 4

EPICENTER DETERMINATION WITH SPARSE BROADBAND ARRAYS AT REGIONAL DISTANCES: THE INCORPORATION OF SURFACE WAVES

INTRODUCTION

One of the most important problems in monitoring a Comprehensive Test Ban Treaty (CTBT) is the detection and location of small seismic events. However, determining location becomes more difficult at smaller magnitudes primarily due to a decrease in the signal to noise. Furthermore, at smaller magnitudes, there are significantly more events than at larger magnitudes, compounding the difficulty of the location problem. Nonetheless, low magnitudes detection and location determination is paramount if a global CTBT is to be truly viable and deter clandestine and evasive nuclear tests. Thus, it is an active area of research. For example, seismic arrays have become the central tool for event detection in the context of monitoring a nuclear weapons treaty (Mykkeltveit et al., 1990) and much work has been done developing methodologies for large, dense, short-period arrays (Jurkevics, 1988; Suteau-Henson, 1990). For a low magnitude event, a few broadband stations from a global network may also be used in a single station (Vsevolozhsky and Wallace, 1997) or multi-station location. Somewhere between an array location and a single station detection lies sparse regional networks. Since the seismic coverage of the globe is non-uniform, there will be cases where small (<4.0) events are recorded on several broadband stations. Optimization of this information is essential to high performance of any monitoring effort.

There has been some work on event location with sparse broadband networks. For example, Bratt and Bache (1988), using two short period arrays, and Thurber et al. (1989),

using a three station network, incorporated P-wave backazimuth information into a travel time location algorithm. Bratt and Bache (1988) used a frequency-wavenumber analysis to determine backazimuth, and Thurber et al. (1989) estimated the polarization direction using a two-axis principal component method. Other backazimuth estimation techniques have been evaluated by Walck and Chael (1991). However, these methods all utilize P and S waves, hence the full power of the waveform, in particular the surface wave, has not been utilized. Furthermore, there is still a strong reliance on travel times, which are a function of the assumed velocity structure and the ability to correctly identify seismic phases, to determine distance. Therefore, we use information contained in the surface wavetrains to improve on body wave backazimuth schemes and altogether avoid the dependence on a distance determination by "focusing" the network on the best epicenter. The benefit of this technique is that it is completely transportable and can be automated. The data used in this study consists of small seismic events recorded on a sparse broadband network in Bolivia containing at peak operation up to 23 stations.

In developing processing techniques for the full waveform, we have worked on three methodologies. To determine backazimuth we use 1) a polarization filter for body waves, and 2) a correlation grid-search on mid-period Rayleigh waves (2 to 20 sec). The third methodology consists of combining the information from the body and surface wavetrains in a linear least squares inversion to determine location. However, it may be necessary to incorporate distance information if there is data from only a few stations (less than 3 or 4 stations, depending on the quality of the backazimuth determination). This potential fourth methodology requires using a three-component amplitude envelope time differential to estimate distance. Unfortunately, this also requires an *a priori* knowledge of the velocity structure to determine travel times which we seek to avoid.

DATA

In a cooperative project, the University of Arizona, the Carnegie Institute of Washington, and Lawrence Livermore National Laboratory deployed a total of 23 broadband stations (10 to .01 Hz) in Bolivia for over one year. The Arizona and Carnegie transect, known as BANJO, consisted of 16 stations forming an east-west transect. The BANJO line traversed extremes in elevation and geology. As shown in Figure 4.1, the stations span from Chile, over the Western Cordillera, Altiplano, Eastern Cordillera, and down into the sub-Andean zone and into the Chaco Plain. The SEDA array deployed by Lawrence Livermore National Laboratory consists of seven stations and strikes north-south along the eastern edge of the Altiplano (Fig 4.1).

The location of the sparse array in Bolivia resulted in events being recorded which traversed extremes in crustal geology; crustal thickness varies from 40 to 75 km. We test the location algorithm on one earthquake (April 3, 1994) that occurred during the early stage of the deployment. Thus, it was recorded with good surface waves at only 7 stations (Fig 4.2). This event has an epicenter located northeast of the array. In Figure 4.1 the circled star represents the Preliminary Determination of Epicenters catalog location (17.694S, 64.816W, depth = 38km, $m_b=5.3$, $M_s=4.4$), and the grey ellipse represents the location determined by a standard body wave travel-time algorithm using data from only the 7 stations in the BANJO array.

METHODOLOGY

Flinn (1965) showed that it is possible to use a three-axis principal component analysis to characterize the polarization of seismic waves. Subsequent techniques have been used by other investigators. Magotra et al. (1987) used only the horizontal components in a two-axis principal component method. Christoffersson et al. (1988) implemented a maximum-likelihood approach for backazimuth estimation, and Roberts et al. (1989) extended this to

Figure 4.1 A map of Bolivia showing the BANJO stations (white triangles), the SEDA array (inverted grey triangles), and GSN station LPAZ (*). The PDE location for the April 3, 1994, example event is shown by the circled star. It was also recorded at the seven circled BANJO stations from which a second location was determined (grey ellipse). The 12,000 foot contour line separates the darker from the lighter elevations.

Figure 4.2 A record section of the long period signal (5 - 20 seconds) for the event on April 3, 1994, recorded on the vertical components of the BANJO stations.

near real-time. Kvaerna and Doornbos (1986) developed a frequency-wavenumber algorithm to estimate backazimuth. Walck and Chael (1991) used a distributed network of three-component, short period stations in central and eastern North America to evaluate four of these different computational methods for determining backazimuth as well as the effects of receiver structure on backazimuth determination.

Unlike P-waves, surface waves are confined to travel along the surface of the earth; Rayleigh waves, tied to the P-SV system, have retrograde elliptical particle motion. Thus, coherent energy is defined in two dimensions by a plane that is associated with the backazimuth, instead of in one dimension by a ray, as is the case for the particle motion of P-waves. In the simplest sense, the "best plane" on the basis of particle motion gives the best backazimuth determined from Rayleigh waves.

We prefer the three-axis principal component analysis of Flinn (1965) to determine P-wave backazimuth. Jurkevics (1988) extended Flinn's method to arrays, however, this technique cannot be applied to a sparse network nor can it accommodate our technique of determining backazimuth from surface wave information. Therefore, we modify the method of Bratt and Bache (1988), who extended the travel time residual algorithm of Jordan and Sverdrup (1981) to include azimuthal data. The following sections describe the techniques used to estimate backazimuth from body waves and surface waves and how this information is used in an iterative least-squares inversion scheme to determine the epicenter.

Body Wave Backazimuth

The polarization of three-component P-waves is a function of the backazimuth. A data covariance matrix C is first developed

$$C = \begin{bmatrix} \sigma_{nn} & \sigma_{ne} & \sigma_{nz} \\ \sigma_{ne} & \sigma_{ee} & \sigma_{ez} \\ \sigma_{nz} & \sigma_{ez} & \sigma_{zz} \end{bmatrix} \quad (1)$$

where $\sigma(i,j)$ represents the zero-lag cross power

$$\sigma_{ij} = \frac{1}{N} \sum_{i=1}^N x_i y_j \quad (2)$$

In (2), x and y are equal to the north-south, east-west, or vertical components of the seismic data. The data covariance matrix C is rotated into its principal directions, which lie along the eigenvectors of C . The rotated covariance matrix C' has the form

$$C' = \begin{bmatrix} \lambda_1 & 0 & 0 \\ 0 & \lambda_2 & 0 \\ 0 & 0 & \lambda_3 \end{bmatrix} \quad (3)$$

where the diagonal elements are the real eigenvalues of C , with $\lambda_1 \geq \lambda_2 \geq \lambda_3$. The three eigenvectors associated with the eigenvalues define a polarization ellipsoid that has any arbitrary orientation with respect to the original north, east, and vertical system. P-waves propagate along, and are contained within, the true radial direction between the source and receiver. Thus, the dominant eigenvector (associated with λ_1) for P-waves is assumed to lie in the radial direction of propagation. The backazimuth is the projection of this eigenvector, or radial direction, onto the horizontal plane at the receiver.

Trade-offs between resolution and estimation variance are a function of both window length and bandwidth. Short windows and narrow bands distinguish between close arrivals and illuminate the frequency varying properties of polarization, whereas longer windows and wider frequency bands yield more stable and reliable estimates. On average, we use a

sliding time window of 2 seconds in length that steps at one half the window length for the total duration of the waveform record (Fig. 4.3). Thus, as is the case for most regional events, Pn is sampled followed by the remaining signal. We experienced no adverse effects from potential converted phases or off-azimuth scattered energy with longer window durations consistent with the analysis of Walck and Chael (1991). Stable backazimuths are determined for the time-sampled P energy.

Since we are interested in the most stable estimates, we want to utilize a wide range of frequencies. However, a single wide frequency band will result in the frequency component at the peak amplitude dominating the polarization estimate. Therefore, we incorporate a wide-band frequency estimate following Jurkevics (1988) that provides spectral balancing over a range of frequencies. To achieve high signal to noise in the P-waves, it was necessary to high pass filter the signal at least at 0.5 Hz.

Surface Wave Backazimuth

Regional distance surface waves can be very complex. They are highly dependent on the crust and upper mantle structure and extremely sensitive to source radiation patterns. Furthermore, at regional distances, Love and Rayleigh waves can arrive with little separation in time. Nonetheless, they can provide valuable backazimuth information. To extract this information, we use mid-period (1-20 seconds) Rayleigh waves in a grid search on azimuth (at 10° increments) to determine where the radial component best correlates with the vertical component. Theoretically, the retrograde elliptical motion implies that the radial and vertical components should correlate well with a 90° phase change. The magnitude of the correlation between the trial radial to vertical and trial tangential to vertical is determined and normalized by their respective standard deviations and statistical kurtosis (third moment of the population). The ratio between these two values (R) is determined for each 10° increment in azimuth (Fig. 4.4) and the process is repeated.

Figure 4.3 A map showing the P-wave backazimuth vectors determined by the 3-component polarization technique on data recorded at station CRIS. The vectors were determined using a 2 second time window sliding at 1 second increments and represents the horizontal projection of the eigenvector associated with the largest eigenvalue. The length of the vector is an indication of the "quality" of the determination. The longer the vector, the more robust the backazimuth. The color of the vector partitions what 2 second portion of the P-wave was used in the inversion. The two traces in the upper left show, as a function of time, the 3 component amplitude, which is merely the square root of the trace of the covariance matrix, and the backazimuth. Thus, after an initial 180° shift, the algorithm determines a consistent backazimuth throughout the P-wave arrival (shades of white).

Figure 4.4 A map showing the Rayleigh wave grid search results for each station. The disks display the large (white) and small (black) ratio values as a function of azimuth, with north at the top of the disks. The white represents where there is good correlation between the radial and vertical components, and the black represents where there is good correlation between the tangential and vertical components. Thus, the white estimates the Rayleigh wave backazimuth and its "thickness" or "width" represents the resolution of the estimate. The disks are plotted with respect to each station's location.

$$R = \frac{\text{correlation value (radial)} * \text{sdev (radial)} * \text{curtosis (radial)}}{\text{correlation value (tan.)} * \text{sdev (tan.)} * \text{curtosis (tan.)}} \quad (4)$$

The maximum correlation value (R) should occur at the backazimuth. However, in some cases the ratio was dominated by the Love wave arrival due to a source radiation effect. This would result in an indicated backazimuth tangential to the true backazimuth. This is accounted for in the relative timing of the two surface waves. If the maximum correlation occurs between the tangential and vertical components and sooner in time than the radial to vertical correlation, the surface wave is predominantly Love wave energy. Although this technique is redundant every 90° in azimuth, the backazimuth can be uniquely determined by examining both the timing and the particle motion. The true backazimuth occurs where the Rayleigh wave particle motion is retrograde elliptical instead of prograde elliptical.

Epicenter Determination

Both the body wave and surface wave backazimuth techniques are sensitive to both the regional structure as well as the receiver structure. Therefore, it is necessary to take advantage of the full benefit of utilizing array data by "focusing" the array on the best location, thereby averaging out any localized effects. We use a standard least squares inversion scheme to iterate upon the backazimuth information from each station to determine the best solution. This technique involves iteratively solving a perturbation vector $\Delta \underline{x}$ that minimizes the residuals $\underline{r}_w$ between the observed azimuths and those azimuths computed from a trial epicenter $\underline{x}_c$ (Aki and Richards, 1980),

$$\Delta \underline{x} = (\underline{A}_w^T \underline{A}_w)^{-1} \underline{A}_w^T \underline{r}_w \quad (5)$$

where $\underline{A}_w$ is the weighted system matrix. The new estimate of epicenter $\underline{x}$ is $\underline{x}_c + \Delta \underline{x}$.

The system matrix $\underline{A}$ has dimensions $2N \times M$. Since we are only solving for epicenter, we have only two model parameters (M). In terms of N , each station providing information adds two rows to the residual ($\underline{r}$) and system ($\underline{A}$) matrices. The $\underline{A}$ matrix contains the azimuthal information of partial derivatives of the data with respect to hypocentral Cartesian coordinates x (easterly) and y (northerly). Backazimuth α is measured clockwise from north. Following Bratt and Bache (1988),

$$\frac{\partial \alpha}{\partial x} = - \frac{\cos(\alpha_e)}{\sin(\Delta)} \frac{0.0090^\circ}{\text{km}} \quad (6)$$

where the station is initially Δ radians distant and α_e radians azimuth from the event. Similarly, the northerly partials are

$$\frac{\partial \alpha}{\partial y} = \frac{\sin(\alpha_e)}{\sin(\Delta)} \frac{0.0090^\circ}{\text{km}} \quad (7)$$

The partials for the P-wave azimuthal information fill one row, and the partials for the surface wave azimuthal information fill a second row for each station used.

A trial epicenter is chosen by inspecting the first motions or the relative arrival time of energy across the array. The trial azimuths are calculated and the data residual vector determined. Subsequently, the $\underline{A}$ matrix is filled and normalized by the appropriate standard deviations before computing the perturbation vector ($\Delta \underline{x}$).

Distance determination

When using only a few stations, it may be necessary to add distance information to the inversion, or at least use the distance information to provide a first order check to the epicentral solution. Normally, determining distance requires correct identification and accurate time picks of the onset of the seismic phases used, usually P and S. Subsequently, this timing information is inverted against an *a priori* knowledge of the velocity structure of the region to determine distance from each station to the source. In fact, using the collective

Figure 4.5 A record section of the 3-component amplitude for the seven stations in the BANJO network that recorded the April 3, 1994, event.

Figure 4.6 A map showing the source to station distance determined from the envelope analysis. Thus, the "best" location according to the P-wave and surface wave time picks is to the east of the PDE location. Distances determined for stations DOOR and CRUZ are significantly less than for the other stations. This is an effect of the variable and complex structure the collective network samples.

timing information is the most implemented and routine inversion procedure for determining seismic locations. The goal of this investigation is to include surface wave information into a location scheme, however, picking the onset of the Love or Rayleigh wave can be very difficult. Therefore, we use the 3-component amplitude (Jurkevics, 1988), or "envelope" analysis, to determine distance.

The 3-component amplitude is the square root of the trace of the covariance matrix in equation 1. Figure 4.5 shows the 3-component amplitude for the seven BANJO stations that recorded the example event. Even where P is very emergent (stations SALI and HIZO, Fig. 4.5), its onset can still be clearly picked using envelope analysis. The maximum amplitude, representing the surface waves, can also be easily distinguished. Although envelope analysis allows for easy identification and timing of P-waves and surface waves, an *a priori* knowledge of the velocity structure is still necessary to invert for distance. Based an Altiplano structure, the following equation can be used to determine distance (x , in kilometers) for each station.

$$time = 0.3948 + 0.1854x + 4.808E^{-5}x^2 \quad (8)$$

where the difference in time between the onset of the P-wave and maximum amplitude of the surface wave is *time*. Figure 4.6 shows the resultant station to source distances determined with equation (8). These circles represent the intersection of a lower projection hemisphere with the surface of the earth. Where these seven hemispheres intersect is the best location. However, only four of the seven distance determinations coalesce about a point that is located east of the assumed "best" epicenter. Of the remaining 3 stations, station CRIS moderately deviates from this intersection, and stations CRUZ and DOOR significantly deviate from this intersection. This is the result of using a uniform structure to determine equation (8) when in reality, irrespective of the epicenter, the collective network

samples a complex structure of variable velocities and crustal thicknesses. This is why avoiding the use of travel times reduces the *a priori* knowledge that contributes to the formal error.

RESULTS

The final epicentral solution and confidence bounds on that solution depend on the variance assigned to the data. Therefore, after determining the backazimuth for each technique at each station, it is necessary to quantify standard deviations to the backazimuth solutions. The determined standard deviations are based on measurable attributes of the signal. The 3-component polarization technique yields three eigenvectors that define an ellipsoid. The eigenvectors associated with the medium (λ_2) and minimum (λ_3) eigenvalues are perpendicular to the backazimuth and dictate the shape of the ellipsoid. If these eigenvalues are large, the ellipsoid approaches sphericity, and thus is poorly resolved. If these eigenvalues are small, the ellipsoid becomes elongate, and the backazimuth solution is well resolved. To incorporate this into a weighting scheme, the eigenvalues are normalized to the largest eigenvalue (λ_1) having a value of 10. The average of the remaining eigenvalues (λ_2 and λ_3) is the standard deviation used to weight the backazimuth result and varies between 10 (not resolved) and 0 (perfectly resolved).

It is important that the standard deviation for the surface wave backazimuth determination have the same value for the same resolution of the 3-component polarization technique. Otherwise, one technique will dominate the weighting scheme and drive the location solution even though the two differently determined backazimuth solutions may share the same resolution. Therefore, the ratio values of the surface wave correlation grid search are normalized so that the maximum ratio value is 10. All ratio values are then averaged. This average approximates the standard deviation and again varies between 10, in which the best correlation could not be resolved, and ≈ 0 , in which the best correlation occurs along the backazimuth and all subsequent ratio values in the grid search are 0. In

other words, the lower the ratio value average, the better resolved, or more "spike" like, the backazimuth solution (station HIZO, Figure 4.4).

In the epicenter inversion, standard deviation is inversely proportional to weighting. Thus, the smaller the standard deviation, the greater the assigned weight. However, in the 90% confidence bounds of the final epicenter error ellipse, variance in the data due to, for example, structure causing off-azimuth solutions, or scattering causing imperfect particle motions resulting in imperfectly polarized data, has not been accounted for. Furthermore, no confidence bounds have been determined for each backazimuth solution that may propagate through the location inversion. Therefore, with only two model parameters (latitude and longitude) to fit, the magnitude of the residual vector ($\mathbf{r}$) can be significantly affected by the number of stations. Unless $N \gg M$, the magnitude of $\mathbf{r}$ is very small. If all 25 stations in the network contributed to the solution, $N=50$. Even in this case, the magnitude of $\mathbf{r}$ is still small and results in an unreasonably small confidence ellipse. Thus, for the solution shown in Figure 4.7, the ellipse has been enlarged, but its shape is still the same.

CONCLUSION

Traditional travel time location inversions are dependent on both knowing the velocity structure *a priori*, and, in some cases, not being able to use a laterally varying structure. Azimuth information is also a function of the velocity structure, but only in that the velocity structure determines off-azimuth deflections from the true back-azimuth. In an effort to avoid the dependence on *a priori* information, azimuth data is used to "focus" a sparse network and determine the best location. However, P-waves and surface waves are sensitive to different regions of the earth, because they have very different ray paths. Thus, the information in surface waves is also used to determine back-azimuth, thereby not

Figure 4.7 A map of Bolivia showing the BANJO stations (white triangles) used in the location inversion. The PDE location of the April 3, 1994, event is shown by the circled star. The epicenter determined by the body wave travel time picks using the BANJO data is shown as the grey ellipse. The epicenter determined by the inversion on backazimuth is shown as the white ellipse.

biasing the location solution towards an off azimuth characteristic of using only one seismic phase.

Using P-wave and Rayleigh wave back-azimuth information recorded on a broadband array in Bolivia, an iterative least-squares inversion is performed to determine the epicenter. In spite of the complex structure the network covers, the inversion converges on a "best" location. This location is congruent with the location determined using envelope analysis. However, the back-azimuth location is located to the east of the locations determined using traditional travel time residuals.

CHAPTER 5

**AFTERSHOCK SOURCE MECHANISMS OF THE
JUNE 9, 1994, DEEP BOLIVIAN EARTHQUAKE**

INTRODUCTION

The largest (M_w 8.3) deep focus earthquake in modern recorded history occurred on June 9, 1994. This event ruptured the down-going Nazca slab at a depth of 636 km under Bolivia. We use data recorded on the BANJO and SEDA seismic arrays, as well as Global Seismographic Network (GSN) station LPAZ, to study the short-lived aftershock sequence. During this sequence, 19 portable broadband seismic instruments were being operated in Bolivia by the University of Arizona and the Carnegie Institution of Washington (BANJO) as well as by Lawrence Livermore National Laboratory (SEDA). The stations range in epicentral distance from 400 to 900 km (Fig 5.1). We recorded 12 aftershocks, which range in magnitude from 2.8 to 5.6, with good signal to noise ratios for P and S waves.

Aftershocks are invaluable for providing indirect constraints on the source parameters and rupture properties of the main shock. For shallow earthquakes, it is common that aftershocks are closely related to the fault plane and often outline the extent of rupture. The focal mechanisms of aftershocks are a measure of the state of stress in a volume about the fault rupture and provide constraints on the tectonic stress field. However, the mechanism of deep focus earthquakes is still poorly understood. Whether or not they occur in response to brittle failure like shallow, crustal earthquakes, or occur as a result of different processes such as transformational faulting, has yet to be agreed upon primarily because of limited and ambiguous data. Frohlich and Willemann (1987) and Willemann and Frohlich (1987) analyzed numerous deep earthquake sequences and found no preferential clustering of

Figure 5.1 A map of Bolivia showing the BANJO array (white triangles), the SEDA array (inverted grey triangles), GSN station LPAZ (*), and the mechanism and location of the mainshock. The 12,000 foot contour line is also shown.

aftershocks near initial-event nodal planes as would be expected if deep earthquakes were progressive shear failures along a planar surface. In contrast, Wiens et al. (1994) located 9 aftershocks of the March 9, 1994, Tonga deep slab event using a local, portable network and found that the aftershocks align along a nearly vertical plane. Until the Bolivian event, this earthquake ($M=7.6$, depth = 564km) was the best digitally recorded deep focus event. The mechanism of the deep Tonga event ($\theta=7^\circ$, $\delta=77^\circ$, $\lambda=-114^\circ$) indicates either predominate strike-slip motion on a shallow dipping plane to the north, or predominate dip-slip motion on a steeply dipping plane to the east. The aftershocks align on a plane very similar to the steeply dipping plane of the main shock (Wiens et al., 1994).

Unlike the Tonga event, the Bolivian aftershocks are not located along a common plane (Myers et al., 1995). A few aftershocks align with the horizontal plane of the initial event while the bulk of the aftershocks locate below the main shock. Thus, the aftershock locations do not independently constrain which main shock nodal plane ruptured or indicate which plane underwent continued slip during the aftershock sequence. However, a directivity analysis of the P and S waves from the main shock by Beck et al. (1995) strongly indicates that the main shock ruptured to the east along a horizontal plane.

Aftershocks are usually significantly smaller in size than the main shock and require the use of local or regional waves to reveal any information about the source. Focal mechanisms determined from first motions require a detailed knowledge of the velocity structure as well as a dense local network. However, with the advent of portable, three-component, broadband seismometers, local (<100km) waveforms recorded on a sparse network or single station have been successfully inverted for source parameters. Furthermore, intermediate-period regional (<1500km) surface wave spectra have been inverted for source parameters of moderate sized events in the western U.S.

The BANJO and SEDA arrays recorded high quality regional broadband waveform data for the Bolivian aftershocks, yet none of the above techniques could successfully be applied. First, the Bolivian event was very deep and thus generated no surface waves. Secondly, there was no dense seismic network or adequate detailing of the velocity structure to constrain a reasonably unique first motion solution. Finally, waveform inversions would be difficult on all but the largest aftershock due to poor signal to noise ratios for raw broadband waveforms. However, for twelve events we recovered clear P, SV, and SH waveforms when the data were band pass filtered at 1 to 5 Hz. Thus, we use a grid search technique using P and S amplitude ratios to determine focal mechanisms. In all cases the mechanisms are dominated by a nearly horizontal east-west trending T-axes that, for 75% of the solutions, are complemented by near vertical P-axes, while the other 25% of the solutions are strike-slip.

DATA

The BANJO and SEDA arrays are located approximately one focal depth south of the epicenter along azimuths spanning from 150° to 190°. The BANJO array has 12 stations arranged on an approximate east-west line and contains the eastern-most station. The SEDA array consists of 7 stations aligned on a roughly north-south transect that intersects the BANJO array (Fig. 5.1). The northern-most station of the SEDA array, the western most station of the BANJO array, and GSN station LPAZ, have approximately the same azimuth from the main shock epicenter, the only difference being distance.

MAGNITUDE DETERMINATION

Magnitudes were determined by calculating the amplitude ratio to a reference event, in this case, the largest aftershock (event #4, see Table 5.1). The change in magnitude (ΔM)

is equal to the logarithm of the ratio of the amplitude of the first motion for the known event (A_m) to the amplitude of the first motion for the unknown event (A) at a single station.

$$\Delta M = \log (A_m / A). \quad (1)$$

We averaged the change in magnitude determined from four stations between which there was good consistency. This yielded a maximum standard deviation of 0.48 for event #6 and a minimum standard deviation of 0.08 for event #3.

TABLE 5.1. The event numbers and corresponding origin times, fault plane solutions, magnitudes, and the locations (locations reported by Myers et al., 1995). There was no location reported for Event 8. Therefore, we used the mainshock location to generate the synthetics.

Event	Date	Origin time	Strike	Dip	Rake	Mag.	Latitude (°S)	Longitude (°W)	Depth (km)
1	June 9	01:35	10°	50°	-40°	3.5	13.695°	67.271°	654.4
2	June 9	02:49	230°	50°	-80°	3.1	13.380°	68.310°	629.8
3	June 9	04:23	110°	40°	-100°	2.8	13.595°	67.135°	661.7
4	June 9	05:26	260°	50°	-110°	5.3	13.880°	67.250°	638.0
5	June 9	09:51	120°	40°	-70°	2.8	13.739°	67.366°	636.5
6	June 9	12:12	290°	50°	-110°	3.1	13.570°	67.346°	648.4
7	June 9	18:35	280°	50°	-100°	2.9	13.790°	67.150°	633.1
8	June 9	20:22	110°	40°	-90°	2.9			
9	June 9	21:53	110°	40°	-100°	2.8	13.537°	67.050°	656.4
10	June 16	01:52	100°	40°	-60°	3.1	13.666°	67.327°	633.1
11	June 17	03:06	290°	50°	-100°	3.5	13.570°	67.581°	622.8
12	August 8	07:54	110°	50°	-100°	5.6	13.470°	68.340°	594.0

In terms of the grid search mechanism determination, for the largest aftershock we used data recorded at 9 stations; for the smallest events we only used 4 stations. The number of stations used was contingent upon how confidently we could pick the S wave arrival. Although it is usually much smaller in amplitude than the P-wave, the S-wave is clearly distinguished on a 3-component seismogram. Furthermore, within the azimuthal coverage

of both the BANJO and SEDA arrays, there is a measurable difference in the relative amplitudes of the primary body wave arrivals as a function of location about the source.

GRID SEARCH TECHNIQUE

We employ a grid search technique similar to Schwartz et al. (1995) that evaluates the misfit between the observed P, SV, and SH amplitude ratios and corresponding synthetic ratios to solve for the best double couple focal mechanism. Over the observed 40° span in azimuth there are large variations in the ratios, indicating that some stations were in the proximity of a nodal plane (Fig. 5.2). The synthetics are created by first generating Green's functions for the three fundamental fault orientations using a reflectivity code based on the formulation of Randall (1994). Radiation pattern coefficients are then generated for all possible combinations of strike, dip, and rake at 10° increments (11,700 different possibilities). Then, the Green's functions are multiplied by their corresponding radiation pattern coefficients and filtered to the same pass band as that used to observe the data. Finally, the maximum swing of the P wave on the vertical component and the maximum swings of the SH and SV waves on the tangential and radial components are determined. Although we are at regional distances, the synthetics are very simple because the ray paths are primarily vertical, traveling up through the mantle, to the station. This results in simple seismograms with only 2 identifiable arrivals (direct P and S) (Fig. 5.3).

Due to the importance of these phases for the grid search technique, difficulties can arise from the thick and complex crustal structure of the South American Altiplano in evaluating the reliability of the synthetics. The ratioing technique does not require that the synthetics exactly fit the data. The relative amplitudes are what is important, and they are dependent on assumptions of the crustal structure. We use an 11 layer velocity model based on IASPI91 to approximate the mantle and crust, which for the Altiplano we thicken to 70km. We include attenuation and require that $Q_\alpha = 9/4 Q_\beta$. However, if the S-wave

Figure 5.2 The isolated P-wave and SH wave arrivals for the largest aftershock at 6 stations plotted with a constant y-axis. However, the y-axis for the SH arrivals is magnified then times that of the P-wave yaxis. For example, if the SH amplitude appears smaller than the P-wave amplitude, the P/SH ratio would be greater than ten. Notice the variation in amplitude as a function of azimuth.

Event 4

Figure 5.3 The vertical, radial, and tangential components of station 8 for the largest aftershock in black. The reflectivity synthetics are shown in grey. The y-axis is of constant scale. Note the amplitude of the P-wave arrival relative to the S-wave arrival (at approximately 88 seconds).

attenuation is very high, it would influence P to S amplitude ratios, thereby making them more than just a function of station location about the focal sphere. We assume synthetics with the lowest misfit represent a certain source mechanism. Minor iterations of the velocity structure might yield a smaller misfit to the error, but in general, the mechanisms will still be very similar to those obtained by the original structure. In other words, this technique yields fairly robust mechanisms.

The error for the misfit between the observed ratios and the 11,700 synthetics for each station is evaluated by a simple L2 norm. For each of the three ratios (P/SV, P/SH, and SV/SH), the synthetic ratio (S) is subtracted from the observed ratio (O) and defined by the following function where N is the number of stations

$$\text{error} = \sqrt{\frac{\sum_{i=1}^N \left| O \frac{P}{SV} - S \frac{P}{SV} \right|^2 + \left| O \frac{P}{SH} - S \frac{P}{SH} \right|^2 + \left| O \frac{SV}{SH} - S \frac{SV}{SH} \right|^2}{N}} \quad (2)$$

Once the error value is determined for all values of strike, dip, and rake, it is graphically represented in terms of the maximum (P) and minimum (T) principal stresses to evaluate the model space. In Figure 5.4, the P and T axis for the 15 mechanism pairs with the lowest error are plotted for Event #1. The errors range in value from 5.6 to greater than 3000 for all 11,700 possible solutions, and from 5.6 to 100 for the 7500 lowest error solutions. The 60 solutions with the lowest error, which is about 1% of the error for the 7000 "best" solutions, range in error value from 5.6 to 6.8. The 30 solutions with the lowest error are evaluated for consistency by plotting the trend and plunge of the P and T axis (Fig. 5.4). Fault-plane mechanisms "pair" due to a redundancy about rake (two mechanisms will have the same error and the same nodal planes, the only difference being a change in rake of 180°), therefore, P and T axes cannot be distinguished. Thus, in Figure 5.4, two points represent two mechanisms having opposite rake. For one mechanism, one

Figure 5.4 A plot of the trend of the P and T axis for 30 mechanisms (15 pairs) with the lowest error for Event #1. Notice the 2 cluster of points, one with a trend of approximately 260° and shallow plunge, the other with a trend of approximately 350° and variable plunge. This is indicative of a predominantly dip-slip mechanism, with the horizontal stress axes oriented nearly east-west and the variable sub-vertical stress axes oriented to the north. Inspection of the first-motions indicate normal motion with the sub-vertical stress being the P-axis and the horizontal stress being the T-axis.

of the two points is the P-axis. For the corresponding mechanism of opposite rake, that same point is the T-axis. In other words, 15 pairs of points in Figure 5.4 represent 15 pairs of mechanisms. The bulk of the stress axes in Figure 5.4 are consistent with a predominantly dip-slip mechanism because of the two clusters of axes; one with low plunge, the other with high plunge. Furthermore, each is convergent on a single trend. Therefore, it is inferred that one cluster represents the T-axes and the other cluster represents the P-axes. The redundancy in rake, i.e., whether or not the mechanism shown in Figure 5.4 is a normal or thrust mechanism, is resolved by inspecting the first motion data from the portable stations and teleseismic stations when available. The first swing of the P-wave (up or down) at each station must be consistent with the quadrant of the focal mechanism (compressional or dilatational) in which the station is located based on azimuth and take-off angle. Using this information, the preferred mechanism is chosen.

RESULTS

We evaluated 12 aftershocks, 9 of which occurred within one day of the main shock, using this grid search technique. The 11 immediate aftershocks occurring within one week of the main shock ranged in depth between 620 km and 660 km (Myers et al., 1995). The grid search technique for most of the aftershocks found well developed minima, although a few were slightly variable in terms of rake. In general, we observe 2 groups of mechanisms. Nine of the twelve solutions are characterized by a dominant dip-slip component and the remaining three solutions are predominantly strike-slip. Based on the first motions of the portable data and where available, teleseismic data, we found the dip-slip mechanisms are normal. The source-receiver raypaths leave the source traveling up out of the focal sphere with takeoff angles ranging from 41° to 54° from vertical and ranging in azimuths from 150° to 190° . For many events, the first motions plot on the focal sphere very near a nodal plane. However, the dip of the nodal plane trades off with assumptions

about the velocity structure, particularly, the mantle source region. Inserting a low or high velocity zone immediately above the source alters the take-off angle thereby flattening or steepening the nodal plane. In terms of the attenuation structure, increasing the entire crustal S-wave attenuation by a factor of 2 had a minor, if observable, effect (depending on the event) and would not alter the conclusions drawn from our solutions. Figure 5.5 shows a comparison between a teleseismic solution (Wu et al., 1995) and the grid search solution for Event 4 as well as the three solutions with the next smallest error (2,3,4). Figure 5.5 also shows the grid search results for Event 4 using the lower S-wave crustal Q structure. It should be noted that Event 4 had the most pronounced difference in solutions between the two Q structures. Even so, the three solutions with the smallest error obtained from the lower Q structure are very similar to the solutions obtained from the original structure. Thus, in Figure 5.6, we present the mechanisms determined from the original 11 layer mantle structure with the caveat that alterations to the velocities and attenuation values could slightly rotate some of the nodal-planes.

DISCUSSION

The directivity analysis of the main shock (Beck et al., 1995) indicates that rupture occurred on a nearly horizontal fault plane where the top of the subducting Nazca Plate moved north relative to the base of the slab. Some aftershocks were located along this horizontal plane, but the majority were located below this plane (Myers et al., 1995). Thus, the aftershocks define a volume of the slab rather than delineate the main shock fault plane. Although it has been established for some shallow earthquakes that aftershocks defining the main shock fault plane can have variable mechanisms (Beroza and Zoback, 1993), we observe the opposite. Our aftershocks do not form a plane yet do share similar stress orientations. All of our source mechanisms share a nearly horizontal T-axis that is roughly oriented east-west. Nine of our 12 mechanisms have a P-axis that approaches vertical,

Figure 5.5 A comparison of focal mechanism solutions between a teleseismic wave-form technique and our grid search technique for Event 4. Also shown are the next three solutions (2,3,4) with the smallest error obtained from an 11 layer model. These are compared to the three solutions with the smallest error from a model incorporating higher S-wave attenuation. Event 4 had the largest disparity between solutions obtained from the two models.

Figure 5.6 A grey-coded, hypocentral plot of the mainshock and the 8 aftershocks with locations from Myers et al. (1995). The hypocenters (*) are projected onto 4 faces of the cube. The horizontal faces show the locations in map view as a function of latitude and longitude. The front and side vertical faces are parallel and perpendicular to the strike and dip of the slab respectively. The sides of the cube show the vertical and horizontal distribution of the aftershocks within the slab. The vertical dimension is scaled to depth. The focal mechanisms are plotted both in map view and their corresponding cross-sectional view oriented parallel to the dip of the slab. The relative magnitudes are indicated under each event number. No location was reported for Events 2, 8, and 12.

indicative of normal slip, while the remaining three events have a near horizontal P-axis indicative of strike-slip. Wu et al. (1995) have determined the main shock has a P-axis plunging 53° to the north. Therefore, the post-main shock stress field (vertical compression) responsible for the aftershocks has a different orientation than the main shock stress field (approximately down-dip compression). A possible explanation for the rotation of the stress field is that the horizontal plane of the main shock ruptured the elastic width of the slab, reducing the shear stress on the fault to near zero and rotating the maximum compressive stress near the fault plane to a fault-normal orientation. This explanation is consistent with and provides independent corroboration of the main shock rupture characteristics determined from directivity studies (Beck et al., 1995).

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CHAPTER 6

CONCLUSION

More than 50 years since TRINITY, the first nuclear weapons test, the world has finally agreed to ban the testing of all destructive nuclear weapons in a agreement known as the Comprehensive Test Ban Treaty (CTBT). The CTBT finally came to fruition due to many variables, one of which was the ability of seismology to effectively enforce it. This requires the detection and identification of thousands of low magnitude events each year. Therefore, seismologists must appeal to regional distances (<1500 km) in order to record these events above background noise. However, the regional wavetrain is extremely complicated and possesses many unique characteristics for various regions of the earth. The primary goal of this dissertation was to explore certain aspects of the regional wavefield, namely, its effect on seismic phases commonly used in discriminants and if regional complexities may be avoided and yet still determine accurate locations.

Seismic discriminants must be transportable if a seismically uncharacterized region of the world is to be monitored. The Non-Proliferation Experiment demonstrated the discriminant P_n/L_g is affected by both the geologic terrain and the large (gross orientation) and small (topography) scale Moho structure. Knowing the gross behavior of P_n and L_g increases the effectiveness of the discriminant P_n/L_g anywhere in the world.

The most important first order discriminant is location. However, seismically determined regional locations are a function of the regional structure. A location scheme is developed that requires minimum *a priori* knowledge of the regional structure. Using only backazimuth information contained in P-waves and Rayleigh waves, an earthquake is efficiently located in one of the most complex regional environments on earth, the South American Andes.

The ratio of P to S energy is a fundamental component of seismic discriminants. The effects of the South American Andes on P and S waves is evaluated by using P to S wave energy ratios to determine source characteristics for aftershocks of the great Bolivian earthquake of 1994. Finally, a teleseismic relative location scheme is used to locate France's final 6 nuclear tests and evaluate their testing facility in the Pacific.

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