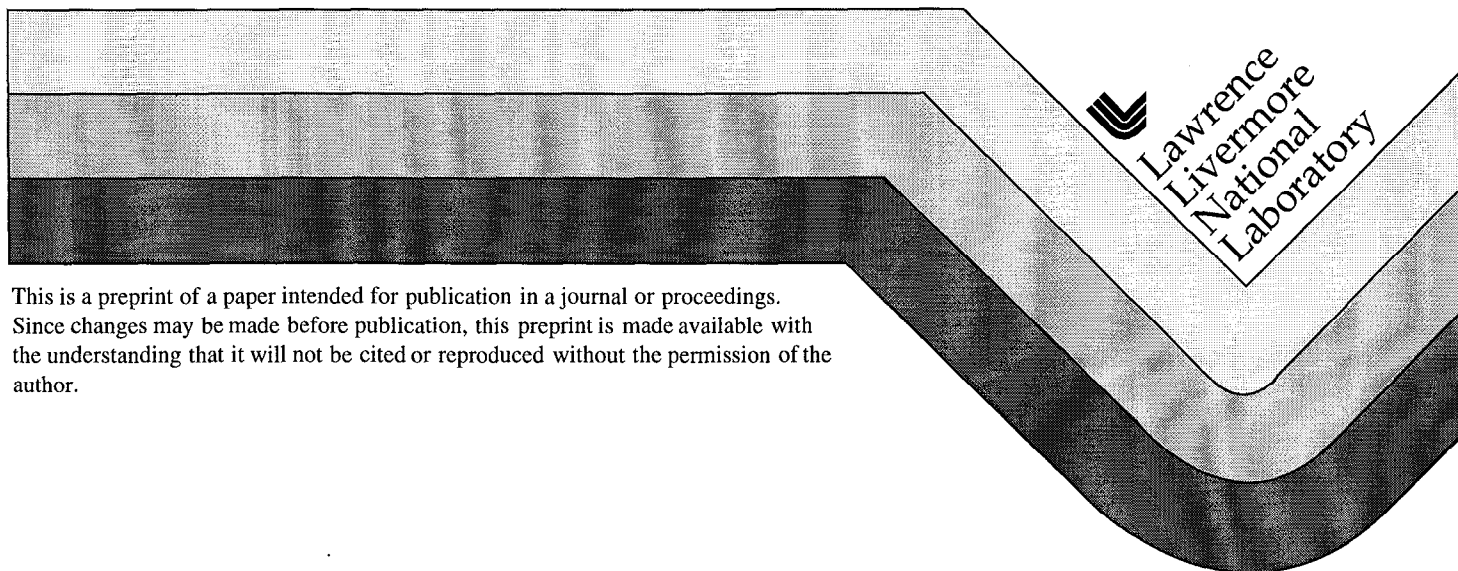# Modeling the Conversion of Hydroacoustic to Seismic Energy at Islands and Continental Margins: Preliminary Analysis of Ascension Island Data

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MODELING THE CONVERSION OF HYDROACOUSTIC TO SEISMIC ENERGY AT ISLANDS AND CONTINENTAL MARGINS: PRELIMINARY ANALYSIS OF ASCENSION ISLAND DATA

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ABSTRACT

Seismic stations at islands and continental margins will be an essential component of the International Monitoring System (IMS) for event location and identification in support of Comprehensive Nuclear-Test-Ban Treaty (CTBT) monitoring. Particularly important will be the detection and analysis of hydroacoustic-to-seismic converted waves (T-phases) at island or continental margins. Acoustic waves generated by sources in or near the ocean propagate for long distances very efficiently due to the ocean sound speed channel (SOFAR) and low attenuation. When ocean propagating acoustic waves strike an island or continental margin they are converted to seismic (elastic) waves.

We are using a finite difference code to model the conversion of hydroacoustic T-waves at an island or continental margin. Although ray-based methods are far more efficient for modeling long-range (> 1000 km) high-frequency hydroacoustic propagation, the finite difference method has the advantage of being able to model both acoustic and elastic wave propagation for a broad range of frequencies. The method allows us to perform simulations of T-phases to relatively high frequencies (≥ 10 Hz). Of particular interest is to identify factors that affect the efficiency of T-phase conversion, such as the topographic slope and roughness at the conversion point and elastic velocity structure within the island or continent. Previous studies have shown that efficient T-phase conversion occurs when the topographic slope at the conversion point is steep (Cansi and Bethoux, 1985; Talandier and Okal, 1998). Another factor impacting T-phase conversion may be the near-shore structure of the sound channel. It is well known that the depth to the sound channel axis decreases in shallow waters. This can weaken the channeled hydroacoustic wave. Elastic velocity structure within the island or continent will impact how the converted seismic wave is refracted to recording stations at the surface and thus impact the T-phase amplitudes.

For this paper we will focus on validating the finite difference method for modeling T-phases in the ocean and land environments and on modeling T-phases observed by the May 1999 Ascension Island Experiment. A network of broadband seismometers on Ascension Island recorded a large number of offshore airgun shots. The shots occurred at all azimuths around the island and at ranges from 1-45 km. Measurements of signal amplitude and duration will be made to understand the variability of T-phase behavior on Ascension Island. The sensitivity to topographic slope and island structure will also be investigated.

OBJECTIVE:

Two types of recording stations are being used for hydroacoustic monitoring of the Comprehensive Nuclear-Test-Ban Treaty (CTBT): hydroacoustic and T-phase. Hydroacoustic stations consist of hydrophones in the ocean sound channel. T-phase stations are seismic stations on islands or near continental margins that record acoustic signals, which convert to elastic waves (the "T-phase") at the ocean-land boundary. T-phase stations are much cheaper to install and maintain than the hydrophone stations. However, this cost-savings is offset by the reduction in signal quality due to energy loss at the acoustic-to-seismic conversion and subsequent propagation of the T-phase through the solid Earth. Because of local site conditions, T-phase amplitudes will vary with back-azimuthal (direction-of-approach).

The objective of this research is to perform numerical simulations of T-phase conversion to identify the factors that most strongly impact T-phase signals and to gain predictive capabilities of T-phase behavior. Simulations of T-phases will be compared with onshore recordings of airgun blasts from the May 1999 Ascension Island Hydroacoustic Experiment (Harben et al., *this volume*). We performed simulations of acoustic-to-seismic conversions using a two-dimensional finite difference code that is accurate for both the acoustic and elastic wave propagation.

RESEARCH ACCOMPLISHED

Background

The propagation of acoustic waves in the ocean has been studied for some time. Acoustic waves in the ocean travel very efficiently for long distances due to propagation in the sound speed channel and low attenuation. The sound speed profile of the ocean decreases to a minimum at about 1000 m depth then increases to the ocean floor (Figure 1a). The sound speed profile for Ascension Island is derived from the World Ocean Atlas (Levitus et al., 1994) using the formula developed by Mackenzie (1981). Notice that the sound speed does not vary much (~6% peak-to-peak). Nonetheless, the low-velocity region of the profile acts as a waveguide to channel acoustic waves without energy loss due to reflection at the ocean surface or floor. When ocean propagating acoustic waves strike an island or continental margin they are converted to elastic (seismic) waves. Such hydroacoustic-to-seismic conversions can be observed by on-shore seismic stations and are called "T-phases". It is known that the efficiency of T-phase conversion is impacted by such factors as local bathymetry (especially topographic slope) and the land path to the seismic station (e.g. Cansi and Bethoux, 1985; Talandier and Okal, 1998).

The Ascension Island Experiment

The International Monitoring System (IMS) specified in the CTBT calls for a network of eleven (11) hydroacoustic stations to detect and locate possible explosion events in or near the oceans. Six of these stations will consist of hydrophones in the ocean and five will be seismic stations recording T-phases. One of the hydrophone stations currently operating is Ascension Island (station code ASCH) in the southern Atlantic Ocean. A field experiment in May of 1999 recorded offshore air-gun blasts with both ocean hydrophones and T-phase seismic stations on-shore. This experiment, described in detail by Harben et al. (1999, *this volume*), provided a unique data set for which to investigate T-phase behavior. Air-gun blasts were used as active sources for an investigation of crustal structure in and around Ascension Island. These shots occurred at distances from 1-45 km and circled the island. Analysis of the shots recorded as acoustic waves in the ocean and T-phases recorded on shore is presented in Harben et al. (1999, *this volume*).

Numerical Experiments

The following sections will describe a series of numerical experiments we performed to investigate various aspects of the modeling procedures and hydroacoustic-to-seismic energy conversion.

Validation of the Modeling: Particle Motion and the Convergence Zone

In order to validate the finite difference method for modeling hydroacoustic waves, we simulated the response of the ocean to an explosive source. We then investigated the particle motions and compared the convergence zones inferred from both finite difference and ray-tracing simulations. The convergence zone is the distance where wave amplitudes are maximal due to constructive inference of refracted arrivals. Multiple refracted arrivals occur because energy is refracted below the SOFAR channel. This creates a zone of focused energy tens of kilometers from the source. The pattern is repeated away from the source due to efficient propagation of sound in the ocean, although the coherence degrades with range.

Figure 1b shows the ray-tracing result for the Ascension Island sound speed model. Using a source depth of 800 m (near the SOFAR channel axis) we traced both upward and downward propagating rays that are refracted back towards the horizontal. These are the rays that can propagate long distances because they are not subjected to energy loss by surface or seafloor reflection. The convergence zone can be seen as the distance at which many rays arrive, about 40 km in Figure 1b.

Using the sound speed profile for the Ascension Island region we simulated the hydroacoustic wave propagation with a two-dimensional finite difference algorithm. The finite difference algorithm requires that the density (ρ) and compressional (v_P, P-wave) and shear (v_S, S-wave) velocity structure be specified on a grid of points. We used the sound speed velocity (Figure 1a) for the P-wave velocities and set the S-wave velocities to zero in the ocean. An absorbing sediment layer (0.5 km thickness) and ocean crust (3 km thickness) were placed below the ocean (Table 1).

Table 1. Velocity Structure for Modeling Validation

Layer	Thickness (km)	v_P, P-wave velocity (km/s)	v_S, S-wave velocity (km/s)	ρ, Density (gm/cc)	Q_P, P-wave attenuation	Q_S, S-wave attenuation
ocean	3.5	Figure 1a	0.0	1.0	50000	-
sediment	0.5	2.0	1.15	2.0	50	22
crust	3.0	6.0	3.46	2.9	1000	444

An explosive source with peak frequency 2.0 Hz was placed near the SOFAR channel axis at a depth of 800 m. The finite difference algorithm solves the equations of motion on the grid by stepping through discrete time intervals. The wavefield can then be interrogated at any point within the grid and written out as time-series of the horizontal and vertical velocity. A description of the code can be found in Larsen and Schultz (1995) and Larsen and Greiger (1998).

Figure 2a shows the vertical and radial velocity records at a distance of 50 km for the ocean model. The marks indicate move-out velocities (2 and 1.5 km/s). The large arrival at approximately 1.5 km/s is the hydroacoustic T-phase that travels to the receiver along purely oceanic paths. The energy arriving after the T-phase is probably from multiply reflected rays (reverberations). Notice that energy arrives before the T-phase, mostly on the vertical component. We analyzed the particle motions before and after the T-phase arrival. Figure 3b shows the particle motion of the T-phase precursor and Figure 3c shows the T-phase particle motion. Notice that the precursor is mostly vertically polarized while the T-phase is horizontally polarized. The precursor results from energy that travels into the sediment and/or hard rock layers as a head wave and then travels upward through the ocean. Energy along these paths proceeds the T-phase because the elastic layers of the seafloor are faster than the ocean. The T-phase is horizontally polarized along the propagation path as is expected from a guided acoustic wave.

Figure 3 shows the simulated T-phase amplitude (pressure in arbitrary units) as a function of distance from the source. The receivers were placed at a depth of 800 m. Notice that the amplitude decreases away from the source then increases to a peak near 35-40 km. This is consistent with the convergence zone estimated from ray-tracing (Figure 1).

The above analysis demonstrates that the finite difference algorithm can model the basic properties of hydroacoustic T-phase propagation. We next proceed to modeling the conversion of hydroacoustic to seismic energy at ocean-land interfaces.

Modeling Hydroacoustic-to-Seismic Conversions

We simulated seismic T-phases with the finite difference code by propagating an explosion generated hydroacoustic signal through the ocean and into an island. For these simulations we used the Ascension Island ocean sound speed profile (Figure 1). Islands were put in the model with varying bathymetric slopes at the island (continental margin). Velocity structure of the islands was similar to the hard rock layer of the model in Table 1, however in this case we added a velocity gradient to the crust. Figure 4 shows an example of the ocean-island environment for the simulations. We sampled the wavefield in the ocean as well as on the island. The location of the source and stations is indicated. In this case the slope of the island margin is 20°.

The resulting response of the model to an explosion source is shown in Figure 5. The source was specified with a central frequency of 2 Hz at a depth of 800 m. For this simulation the source was a distance of 40 km offshore. This distance is not representative of potential CTBT monitoring cases where the sources will be thousands of kilometers away. Because of the near-shore distance, energy that propagates through the elastic crust arrives with observable amplitude (Figure 5). This energy is marked as the 'precursors' in Figure 5. The response for the island stations ($\Delta > 40$ km) shows strong T-phases. Notice that the vertical velocity amplitude at the island stations is reduced by approximately a factor of two. Strictly speaking we should compare the hydrophone pressure to seismic amplitudes. In the future we will attempt to model the absolute pressure and seismic amplitudes. Notice that the T-phases on the island propagate faster than the expected ~ 1.5 km/s moveout for ocean propagation. Modeling of the island propagation delay times for the Ascension Island Experiment will be a presented at the meeting. Determination of this delay time will improve location estimates with the T-phase network.

The amplitude reduction caused by hydroacoustic-to-seismic conversion and propagation on the island is represented in Figure 6. Here we plot the smoothed envelopes of the velocities at the island stations for source depths of 800 m and 100 m. The source depth of 800 m generates a much stronger hydroacoustic T-phase with more impulsive onset. The T-phase for the 100 m source depth is reduced by a factor of 3 compared to the source in the SOFAR channel. The shape of the envelope and the position of the peak amplitude is shifted to later times for the source depth at 100 m, although the onset time is similar for both source depths. The differences in the hydroacoustic T-phases are propagated into the island. Notice the differences in the envelopes for the island seismic stations. The onset times of the seismic T-phases for the near-surface shot are more ambiguous than those for the SOFAR shot (800 m).

This analysis can be used to model the amplitude attenuation within the island that results from both elastic (such as velocity gradients and scattering) and anelastic effects.

CONCLUSIONS AND RECOMMENDATIONS

The results of the previous year's work, although preliminary, promise to advance our understanding of T-phase behavior. The finite difference method is well suited for modeling the short-range propagation we presented above. Ideally we hope to gain some predictive capability of T-phase amplitudes so that we can properly account for azimuthal variations in T-phase travel times and amplitudes. We will be especially interested in modeling absolute amplitudes of hydroacoustic wave pressures and seismic ground motion amplitudes.

We showed that the source depth significantly impacts the amplitudes and waveform character of both hydroacoustic and seismic T-phases. This has implications for T-phase station calibration procedures. It remains to be seen if offshore airgun shots have sufficient amplitude to generate onshore T-phases at realistic ranges. If airgun shots are too weak to generate we will need to develop alternative sources for deeper shots, such as imploding spheres.

In the future we will run many simulations to investigate the effects of ocean-island margin structure on T-phase amplitudes and travel times. For the meeting we will analyze the T-phases recorded at Ascension Island for the Hydroacoustic Experiment and for large earthquakes in the Atlantic Basin. We hope to obtain a model of bathymetry and topography for Ascension Island. This will allow us to test the effects of bathymetry and topography on T-phase conversions.

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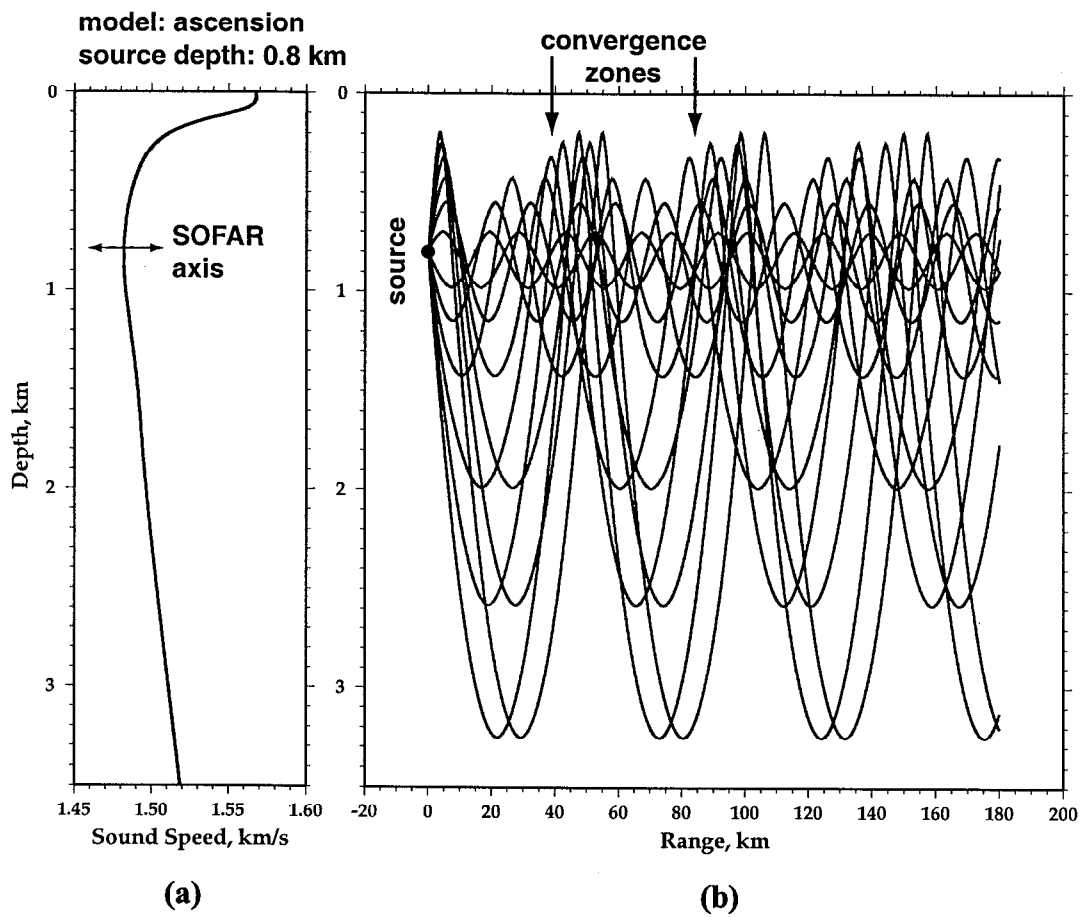
Figure 1. (a) The sound speed profile for Ascension Island. (b) Vertically exaggerated plot of refracted rays for a source at 800 m depth, near the SOFAR channel axis. Note the convergence zones at approximately 40 km intervals.
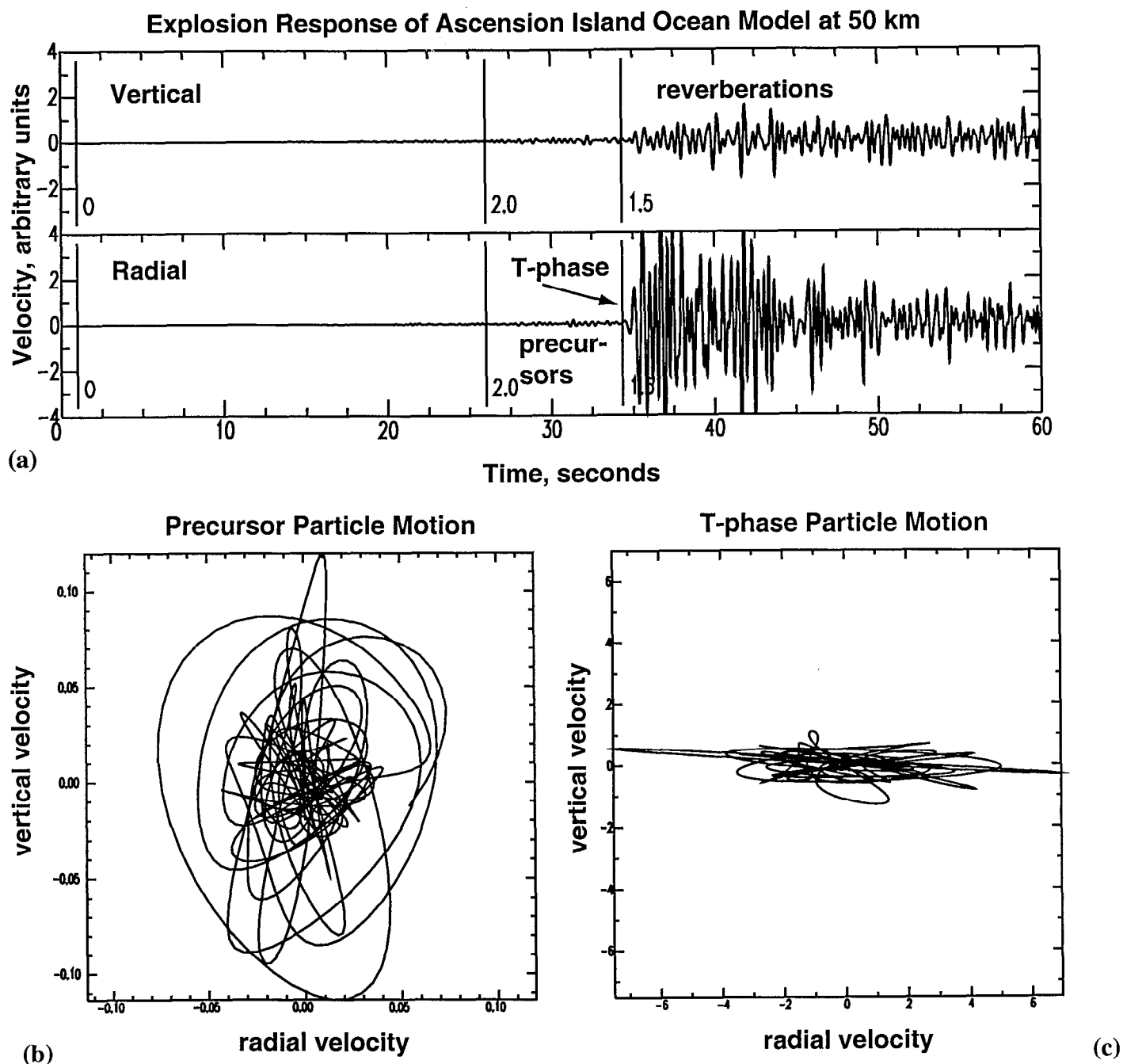
Figure 2. (a) Synthetic vertical (top) and radial (bottom) velocity responses for the Ascension Island ocean model (Table 1) at a range of 50 km. The source and receiver were both at 800 m depth. Marks indicate move-out velocity in km/s. The particle motions for the T-phase precursor (b) and T-phase (c). Note the different scales for the particle motion plots.
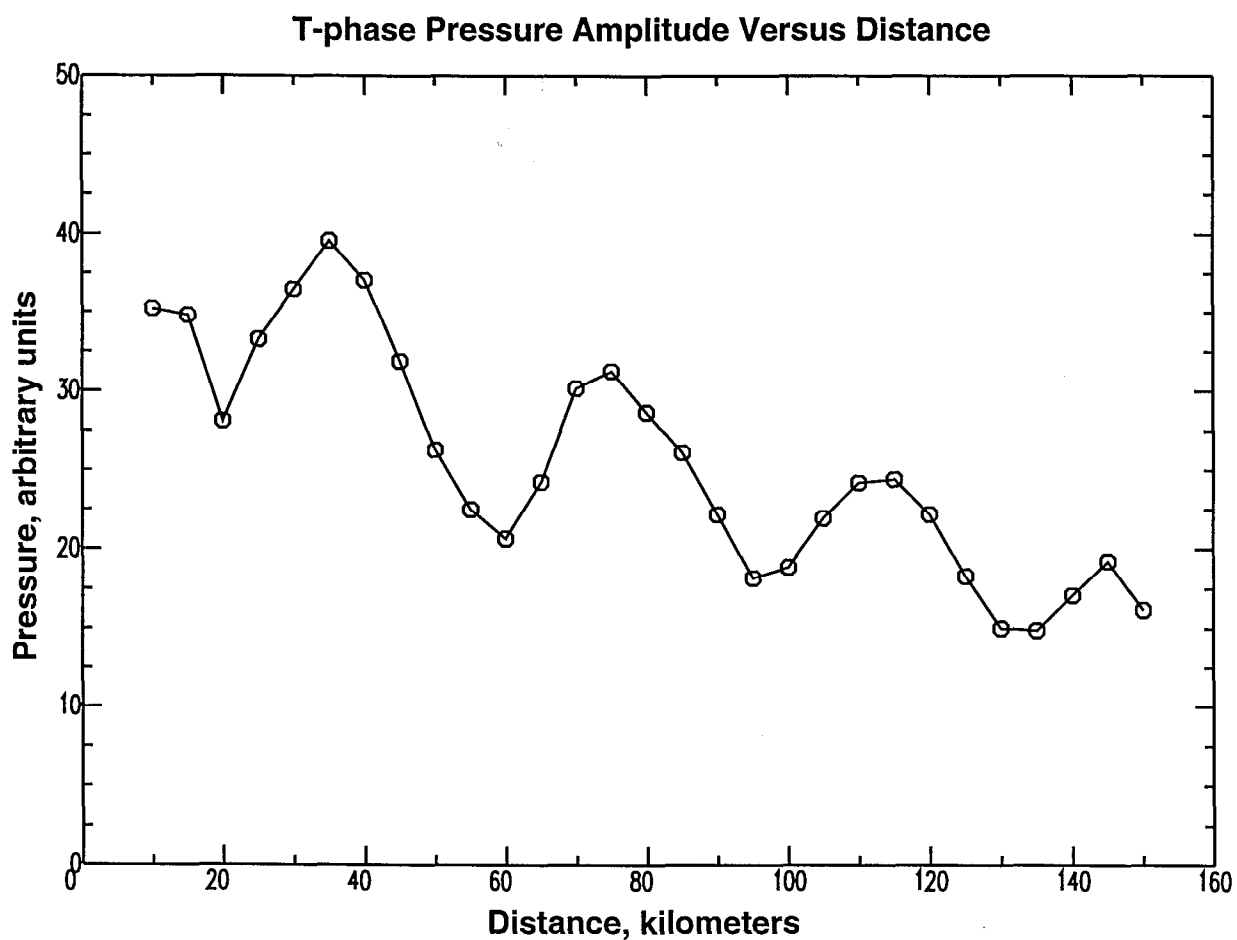
Figure 3. Simulated T-phase pressure amplitude (arbitrary units) as a function of distance for the Ascension Island Ocean model. Pressure amplitude was taken as the root-mean-square (rms) velocity of the radial and vertical components for a window defined by the 1.5 and 1.35 km/s velocities.
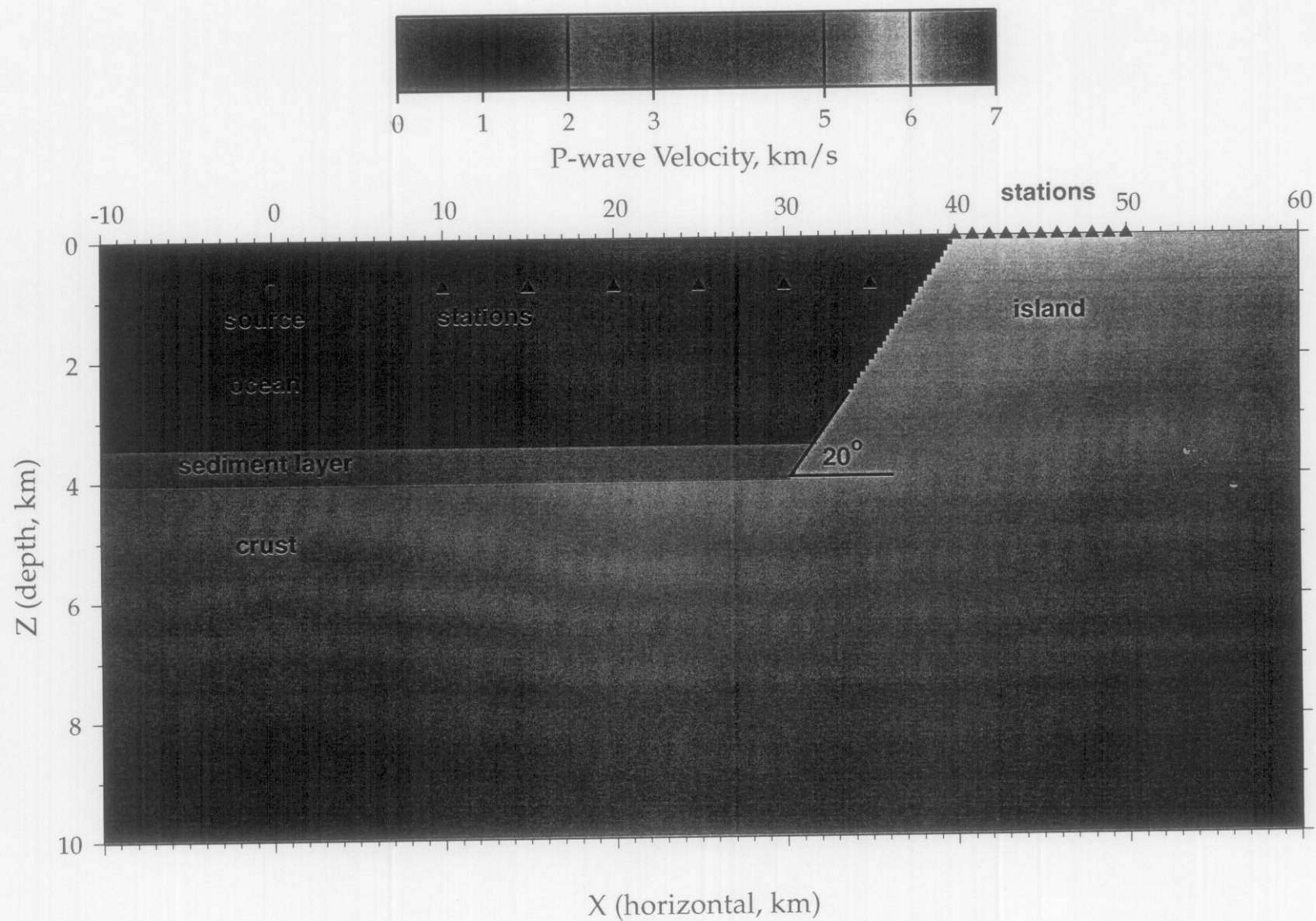
Figure 4. Plot of the ocean-island environment for the simulations of hydroacoustic-to-seismic conversion. The bathymetric slope angle is shown. The source (red circle) and receivers (black triangles) are shown. The compressional (P-wave) velocities are indicated by color bar. Sound speed in the ocean is poorly represented by the color because it varies so little (see the sound speed profile shown in Figure 1). The velocity gradient in the crust is 0.1 km/s/km.

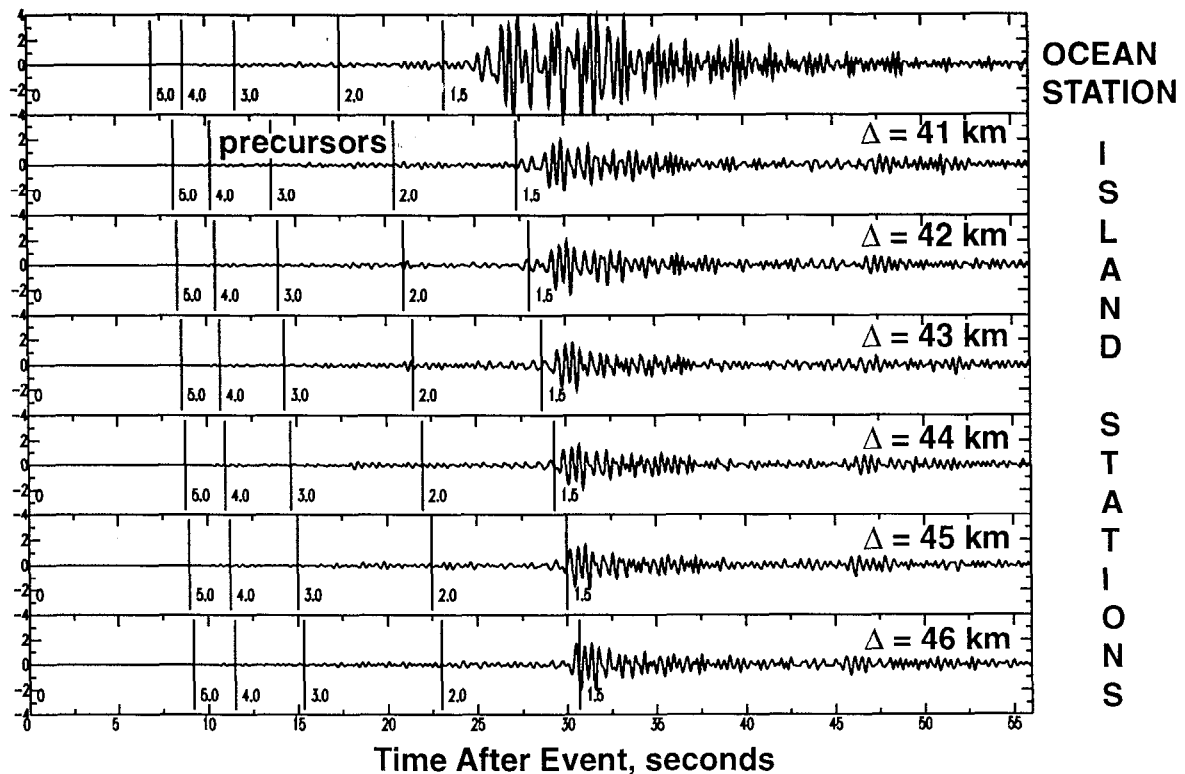## Response of Ocean-Island Model - vertical component velocities

Figure 5. Vertical component velocity response of the Ocean-Island model (Figure 4) foa a source at 800 m depth. Notice that there are precursors to the T-phase that propagate through the elastic crust. The T-phases on the island are reduced in amplitude by a factor of approximately 2. Also note that the T-phases propagate on the island the a velocity faster than the 1.5 km/s moveout.

Velocity Amplitude Envelopes for Hydrophone and Island Stations
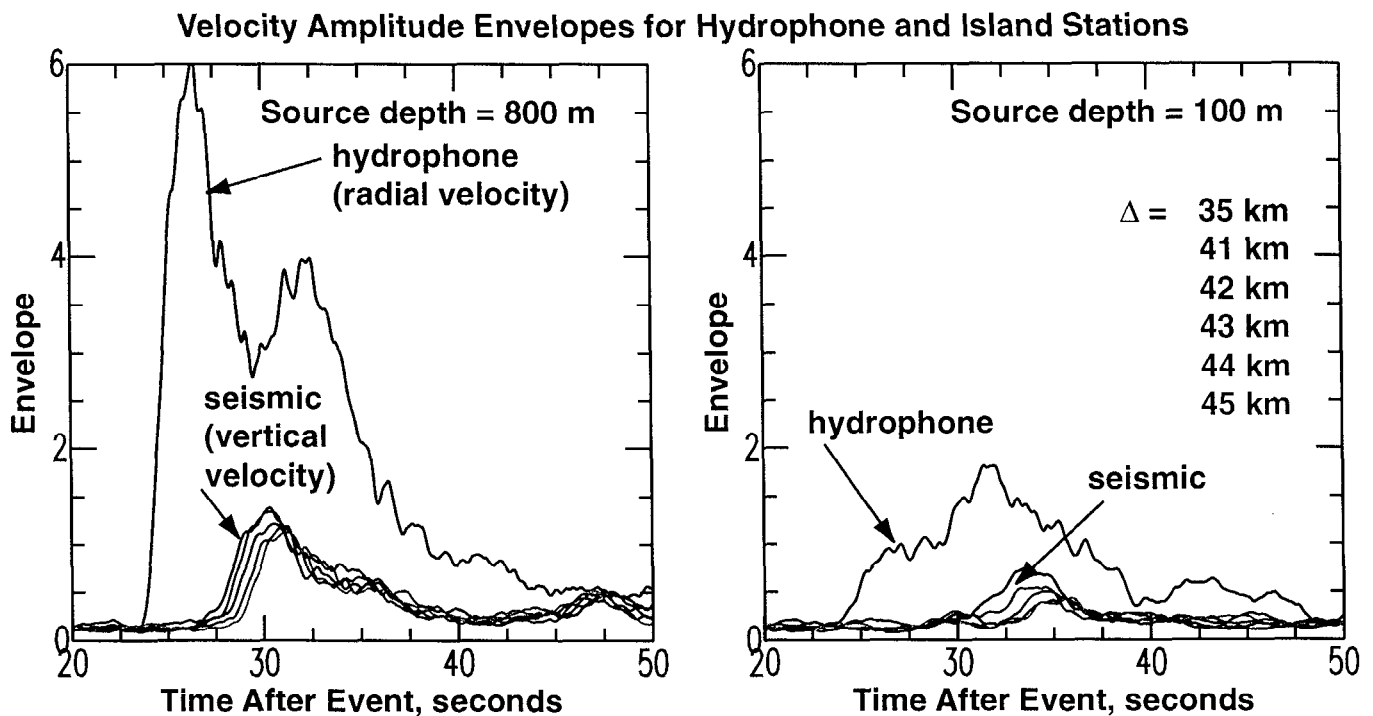

Figure 6. Velocity amplitude envelopes of island T-phases for source depths of 800 m (left) and 100m (right). Note that the hydroacoustic T-phase amplitude is reduced by a factor of 3 for the near-surface shot and the envelope is emergent compared to the shot in the SOFAR channel.