A DESCRIPTIVE STUDY OF THE CONTRIBUTION OF SCATTERING BY SEAFLOOR FEATURES TO LONG-RANGE SOUND PROPAGATION IN THE DEEP OCEAN AT 16 HZ

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Abstract

The long-range signal from a loud underwater impulse contains multiple echoes due to scattering by prominent seafloor features such as islands, seamounts and banks. Intensities of such echoes (relative to the intensity of the direct pulse) will vary with the size of the feature, and the horizontal angle by which the feature deflects the path. Signals recorded at New Zealand from two nuclear shots in Fangataufa, French Polynesia contained six echoes whose intensities could be noted. The relative intensities (RI) at 16 Hz ranged between -44 and -25 dB, and the estimated deflection angle (DA) ranged between 20° and 56°. Neglecting the size of the features, the logarithmic regression between RI and DA was found to be RI = –6 - 18 log |DA|, ± 4 dB. Signals from an earthquake in North-west India have been recorded by CTBTO's "H08" North and South hydrophones in the northern Indian Ocean. The effective source of waterborne sound is localized to where the measured reverse azimuth from H08 North (355°) intersects the continental shelf edge, near the earthquake. The signal at H08 South was scattered by a known seamount, located on the measured reverse azimuth from H08 South of 36°. The estimated deflection by the seamount is 45°, where the RI predicted by the Fangataufa regression is -36 dB, in agreement with the measured RI.

Introduction

In many regions of the world's oceans, the seafloor contains prominent features such as islands, atolls, seamounts and banks. In such regions, the underwater signal from a loud impulse at long range will contain pulses that are echoes due to scattering by these features. An algorithm to compute the travel times of these echoes would need to accept the following parameters:

- Seafloor topography
- Source and receiver positions
- Seawater sound-speed profile (SSP)

An algorithm to compute their intensities would also need to accept one or more geo-acoustic models of the seabed.

An important aspect of this 3-D environment is that sound waves are deflected horizontally. Although results for travel times have appeared in the literature, the state of the art is such that no algorithm is available that can be applied to an arbitrary 3-D scenario. (This may be contrasted with propagation in 2-D range-dependent environments along a straight line in the horizontal plane, for which several algorithms exist.) Especially important are cases where a 2-D characterization would predict a particular region to be a shadow zone (such as behind a bank), whereas a 3-D characterization may predict the sideways deflection of sound waves into that zone, if another prominent seafloor feature happens to be in a suitable position.

In discussing the mechanism of deflection by seafloor features, the approach of [1] is followed. This includes the concept of "horizontal rays and vertical modes", in which a ray is not a conventional ray that travels in 3-D and bounces off the seafloor, but instead is a path in the 2-D horizontal plane. Variation of the sound pressure with depth is accounted for by normal modes, whose properties vary with horizontal position. If such a ray is deflected strongly by a seafloor feature (by more than 90°, say), then it will pass through shallow regions where the mode phase speed is high, and the corresponding attenuation is also high.

For a seamount, the inverse is not necessarily the case. A ray approaching the peak may transit the peak with no deflection but high attenuation. As an incoming ray's offset increases from 0, the deflection passes through a maximum but the attenuation decreases monotonically. For a given deflection there are two rays, an inner and an outer ray, and the latter will have less attenuation. This phenomenon of a maximum deflection is the difference between reflection by a seamount on one hand, and an island or a coastline on the other. Hereafter, a deflected ray is assumed to be the outer ray, except for the case of a zero-deflection path directly over a seamount peak.

For both islands and seamounts, the intensity of a deflected ray will decrease as the angle of deflection increases. To measure a deflected ray with a relatively high intensity, it is therefore necessary to avoid large deflection angles. This is generally achieved by measuring the signal at a range much greater than the distance between the source and the feature. In turn, this requires that the sound source be loud, to yield adequate Signal-to-Noise Ratio (SNR). It must also be of short duration, since a direct signal of long duration will mask the deflected signal. Thus there is not a large quantity of data on this phenomenon. An important contribution was the US Navy's CHASE program of detonating shiploads of old explosives. The results of a 1-kT detonation off the

west coast of the USA, as recorded by hydrophones off the north coast of New Zealand, have been reported [2].
There have been numerous nuclear shots in atolls that would have yielded strong waterborne signals, but placing hydrophones at suitable positions to measure the resulting signals is no small task. The New Zealand Defence Technology Agency (DTA) recorded the signals of six nuclear shots fired on Mururoa Atoll (MA) and Fangataufa Atoll (FA) during 1995 and 1996, and have made these recordings available. The measurable pulses contained in the signal arrived at delays (after the main direct pulse) of between 30 and 340 seconds. An analysis of these is presented later in this paper.

Earthquakes make loud waterborne sounds, but the signal can be high for around 2 minutes, and is thus liable to mask any large echoes. A scenario in which an earthquake signal can yield information on a deflected ray is where the hydrophone is in a shadow zone for the direct ray, so that the deflected ray can be measured in isolation. Such a scenario has occurred in the northern central Indian Ocean, near the Great Chagos Bank (GCB). This bank is around 160 km wide and its perimeter is generally around 10 m deep. The UN Comprehensive Nuclear Test-Ban Treaty Organisation (CTBTO) has installed two hydrophone triplets near this bank for its International Monitoring System. One triplet, referred to by CTBTO as “H08 North”, is 25 km west of GCB. The other, denoted by “H08 South”, is located 100 km to the south of GCB, and is 22 km south of the Diego Garcia atoll. Since direct waterborne paths from earthquakes in western India are blocked from reaching H08 South [3], sound that is measured there may be due to scattering by a seafloor feature, if there happens to be one at a suitable position.

Experimental results

1. Mururoa and Fangataufa to New Zealand.
The recordings that have been analysed were of six underground nuclear shots fired between September 1995 and January 1996. Positions, origin times and seismic magnitudes have been summarised by [4]. Two shots (numbers 2 and 6) were fired on FA, at -22.2°, −138.8° (in this paper, a position is described with the latitude followed by the longitude. A negative longitude is “West”). Shot 6 was 3 km west of shot 2. The other four were fired on MA. Shot 1 was at -21.9°, −138.8°, and shots 3, 4, and 5 were in a 6-km cluster, which was 16 km west of Shot 1. Shots 2, 3 and 6 had seismic body-wave magnitudes of 5.3 or 5.4, shot 5 had a magnitude of 5.1, and shots 1 and 4 had a magnitude of 4.8.
The hydrophone was off the east coast of Great Barrier Island (GBI), New Zealand, at a position of approximately −36.2°, 175.5° (the path crossed the 180° meridian). The seafloor depth (SFD) is 75 m. The hydrophone was mounted about 2 m above the sea floor [5] (it was the shallower of the two hydrophones used by [2]). The recordings continued for at least 13 minutes after the arrival of the direct pulse.
The sample rate was 256 per second for all recordings.
Since the direct waterborne pulses from both FA shots overloaded the Analogue-Digital Converter (for 3 seconds), their spectra cannot be described. The MA shots yielded smaller amplitudes and did not cause overloading. Their spectra had maxima that varied between 14 and 25 Hz. They also had a second maximum at 4 Hz, at around 10 dB less than the major peak. By comparing the waveforms of the MA and FA shots, it is estimated that the actual peak pressure of the latter would have been around double the recorded (clipped) pressure. There was a precursor pulse, presumably a ground wave, whose spectrum was similar to the 4-Hz peak in the direct pulse spectrum. The occurrence of the maximum at 4 Hz rather than at a lower frequency may have been due to the presence of an instrumental roll-off at very low frequency, caused by filters intended to prevent sea-swell pressure changes overloading the recording system [5].
The signal from each shot was filtered into octave bands centred on 4, 8, 16, 32, and 63 Hz, and the resulting waveforms converted to mean square pressure (“intensity”) with an integration time of 1 s. The 4- and 8-Hz bands showed poor SNR and few echoes could be seen. This is attributed to the low-frequency cut-off of the shallow water waveguide [2]. Pulses generally had durations of 8 to 11 s. There were many pulses with durations up to 1 s, but these were classed as noise. The time delay of each pulse was the time interval between its peak and the peak of the direct. A pulse was classed as an echo only if it occurred at approximately the same time delay for at least three shots. The signals in the three higher bands contained up to six echoes whose intensities could be read, but of these only the 16-Hz band has been analysed further, since it is the only frequency in common with data obtained in another ocean to be discussed later in this paper.
The time delays of the echoes following each shot are listed in Table 1 (except for shot 5, which was contaminated by many noise spikes). As expected, the stronger shots (2, 3 and 6) yielded higher SNR and hence more observable echoes than did the weaker shots. The three-shot criterion was relaxed for Echo no. 2, since the temporal variations of the two intensity functions for that echo were quite similar over a duration as long as 20 s (probably associated with the Kermadec Ridge).

Table 1. Measured time delays (s) of echoes following the direct pulse of five nuclear shots. Frequency: 16-Hz.

Shot:	1	2	3	4	6
Echo 1		29	25	27	33
2		71			69
3		104	110		101
4	142	151	144		151
5	214	199	220	220	196
6		337	315	317	336

When the delays for the FA shots (2 & 6) are less than those for the MA shots, the feature is south of the path between FA-MA and GBI, and vice versa.

The intensities of the direct pulse (augmented by 6 dB for shots 2 and 6 to allow for the overloading) and the echoes are summarised in Table 2. Their seismic body wave magnitudes (Mb) are also listed.

Table 2. Seismic magnitudes and acoustic intensities of the direct pulse and echoes (dB re μPa^2 /Hz). Frequency: 16-Hz.

Shot:	1	2	3	4	6
Mb	4.8	5.4	5.4	4.8	5.3
Direct	80	99+6	82	83	98+6
Echo 1		80	63	62	78
2		73			70
3		65	51		69
4	67	61	65		69
5	66	68	62	59	74
6		68	58	49	70

To characterise an echo, a useful parameter is its intensity relative to that of the direct pulse. This will give a meaningful measure of scattering by the feature providing that:
(1) the intensity of the radiated waterborne sound wave is constant over the directions to the hydrophone and the scattering features (in practice this requires the distance from the shot to the continental shelf to be constant over those directions); and
(2) the direct and scattered waves are not themselves attenuated by additional features in their paths.
In regard to item (2), there is a seamount peak 2.1 km deep, 40 km WSW from MA that may attenuate paths to GBI and the Tubuai Island group (called 'Australes' on some charts). This is the likely reason that the direct pulse of shot 3 was 23 dB weaker than shot 2, even though their Mbs were the same.

The relative intensities (RI) for echoes 1, 2 and 3 are -23±3, -33±1, and -35±4 dB. For echoes 4 and 5, there was wide scatter in RI (31 and 23 dB respectively), and the MA shots yielded RIs 10 to 30 dB higher than the FA shots. For echo 6, the two MA results differed by 10 dB, and the lower one was similar to the FA results. This variability may be due to the seamount close to MA, combined with seasonal changes in the seawater SSP, and consequent changes in the propagation path over these longer distances between one shot and another. These changes could be either:
(1) transverse horizontal refraction, or
(2) longitudinal movement of convergence zones relative to seamounts.

Estimating the Deflection Angle

There are many prominent seafloor features in the neighbourhood of the path between FA-MA and GBI. The procedure adopted, which is similar to that used by [2], was to select 24 prominent features north of the path, and a further eight south of the path, and compute the total distance travelled by the echo from each feature. The features divide into four groups:
(1) near the east end of the path and south of it (Rapa Island, Macdonald bank),
(2) near the east end but north (Tuamotu Archipelago and the Tubuai, Society and Cook Island groups),
(3) near the west end and north (Kermadec), and
(4) near the west end and south (seamounts c. -38°, -165°)

Distances and azimuths were obtained with the Vincenty algorithm for a spheroid earth [6]. The distances travelled by the direct pulses varied between 4651 km (Shot 6 on FA) and 4671 km (Shot 1 on MA). Time delays were obtained by dividing path differences by the average propagation speed of 1.491 km/s. (The derivations of the various propagation speeds that arise during this study are described in the Appendix.)
The time delays and deflections vary with shot position. Of the 32 features, five were selected that yielded time delays generally consistent with the observed delays of five of the six echoes. None of the features yielded satisfactory delay times for echo no. 3, which was observed with both FA shots and one MA shot. The closest (with discrepancies in time of 13 – 17 s) is a seamount peak 0.4 km deep at -38.4°, -168°. There are neighbouring seamounts, but this one is the largest. The North-South identification was used for rejecting several features.
The features accepted are all islands except for Macdonald Bank and the seamount. Raivavae is in the Tubuai group; un-named "Southernmost" and Curtis are in the Kermadec group; and Atiu is in the Cook group. The corresponding delays are listed in Table 3.

Table 3. Computed time delays for echoes from each seafloor feature accepted as causing an echo.

Echo	Feature	Shot				
		1	2	3	4	6
1	Raivavae	25	32	24	24	32
2	Southernmost	66	68	66	66	68
3	Seamount	93	88	93	94	88
4	Curtis	152	156	152	151	156
5	Macdonald	217	200	222	226	201
6	Atiu	318	335	318	317	335

The resulting deflection angles for the six echoes are as follows (to the nearest integer): -19±1, -35±0, 29±0, -47±0, 57±1, and -50±0°. The deflection angle is positive for features south of the direct path, and negative otherwise.

RI and DA from shots

Although the size of seafloor features is expected to be an important determinant of RI, an examination of it is beyond the scope of this paper. Instead, RI is considered as a function of one variable (DA). If all the data are considered without discrimination, then there is no trend for RI to vary systematically with DA. If the FA data are considered in isolation, RI does tend to decrease as |DA| increases. The logarithmic regression of the FA results is:

RI = –6 - 18 log |DA|, with standard error ±4 dB.

A logarithmic function was selected since it yields a monotonic function of angle that equals 0 dB at a small angle (although not at zero-angle of course), whereas low-order polynomials that pass through the origin are either non-monotonic or give poor fits.

Some of the decrease in RI with increasing DA may be attributed to the increase in distance travelled as DA increases. For example, a time delay of 340 s would correspond to an additional path of 507 km, yielding a 11% increase in distance travelled. If the echo is a reflection (that is, the angular divergence of the rays leaving the feature is the same as the incident divergence), then the increase in spreading loss will be no more than 1 dB. If however the echo is a scattering process (the angular divergence increases), then the spreading loss would depend on the amount of additional divergence, which will depend on the steepness of the feature.

2. India to Diego Garcia

A strong earthquake occurred in North-west India at 03:16 GMT on 26 January 2001 [7]. Its epicentre was at 23.4°, 70.3°. The Harvard Centroid Moment Tensor analysis [8] yielded depth = 18 km, Mb = 7.9, and half-duration = 21 s. The epicentre is at a range of 3291 km, and on a bearing of 358.7°, from the H08 North hydrophone. Characteristics of the signals recorded by single hydrophones in the North and South triplets have been described by [3]. Since the earthquake occurred inland, the measured signal must have travelled through the ground (as a seismic wave) initially and then through the water (as an acoustic wave). The maxima of both spectra occurred at 3 Hz, and the ratio of the two spectra was approximately constant over the frequency band from 3 to 20 Hz. The sampling rate was 250 per second. Details specific to each triplet are given later.

Travel time differences among the individual hydrophones in each triplet have been measured by CTBTO over three successive segments of the signal, each of 50-seconds duration [9] (a total of 2.5 minutes).

H08 North

From the intensity time series presented in [3], it can be seen that the waterborne signal began 34.6 minutes after the origin time of the earthquake. The spectrogram shows that the waterborne wave had a precursor low-frequency ground wave. The maximum intensity occurred at time 35.67 minutes, and thus lay in CTBTO's second time segment. It had a SNR of over 40 dB. By 2.5 minutes after the beginning (the end of the third segment), the intensity had dropped 20 dB below the maximum value.

Since the spectra shown in [3] would have been dominated by the maximum intensity, the following analysis will focus on the second time segment. Based on the travel times at the three hydrophones, the reverse azimuth of the incoming sound wave has been computed to be 355.2°, which is 3.5° west of the geographic bearing to the earthquake. It is shown in the Appendix that the average propagation speed along paths in this region is estimated to be 1.491 km/s. The speeds of P and S seismic waves from the earthquake to the seafloor are approximately 6.2 and 3.6 km/s respectively [10]. It is assumed that the seismic pulse was of short duration, and that the dispersion of the waterborne signal was due to the contributions of multi-paths with different travel times. The effective source of waterborne sound would have been on the measured reverse azimuth, at the position that yielded the measured travel time. This was 2140 s, which was reduced to 2128 s to allow for the rise time of the seismic pulse [3]. If the P wave created the acoustic signal, the effective source would have been located at 21.78°, 68.59°, where SFD is 0.07 km (on the continental shelf). The travel time from this effective source to H08 North would be 2087 s. If the S wave created the acoustic signal, the effective source would have been located at 21.05°, 68.66° (81 km south of the P-location), where SFD is 2.15 km (at the foot of the continental slope). The travel time from this effective source to H08 North would be 2041 s.

These effective source positions are both near the edge of the continental shelf (although on opposite sides of it), and are close to being at the shortest distance from the earthquake to this edge. Propagation from the earthquake to the hydrophone via this position would thus have less attenuation than paths that enter the ocean at other positions. Relative to the incoming seismic wave, the waterborne wave that travelled to H08N was deflected horizontally by –49° (P wave) or –38° (S wave) at the effective source position. The mechanism for this deflection was possibly reflection by sloping interfaces in the seafloor topography.

H08 South

The South waterborne signal began 2.5 minutes later than the North signal. The maximum intensity occurred 1.0 minute after the beginning, and was 35 dB weaker than the North maximum. By 2.5 minutes after the beginning, the intensity had dropped 10 dB below the maximum value and the SNR was approximately 8 dB.

In contrast to H08 North, no result for travel time differences among this triplet could be obtained from unfiltered signals, but results were obtained when the signals were filtered through the band from 2 to 16 Hz [9]. The first two time segments exhibited reverse azimuths of 36.6° and 36.2° (no result was obtainable for the third segment). These are 40° east of the direct geographic bearing of 356.0°, which passes through GCB.

According to [11], there is a prominent seamount on a bearing of 36° from H08 South. The seamount peak is 1.5 km deep, at -6.4°, 73.38° (approximately 70 km east of the GCB). It is 3161 and 3080 km from the P and S effective sources and 169 km from H08 South, giving total waterborne path lengths of 3330 or 3249 km.

Since the estimated travel times from the effective sources (of the maximum signal) to H08 North are 2087 and 2041 s, the travel times from the effective sources to H08 South via the seamount would have been a further 150 s, namely 2237 and 2191 s. The corresponding average estimated speeds are thus 3330 /2237 = 1.489 km/s (P), and 3249 /2191 = 1.483 km/s (S). By comparison, the group speeds in deep water of the modes that are dominant at H08 South 1.489 to 1.490 km/s, as derived in the Appendix.

RI and DA from earthquake

For rays that start from the P and S effective sources and end at H08S, the Deflection Angles at this seamount are both 45°. The corresponding RI, predicted from the regression based on the FA data, is −36 dB. The measured RI of -35 dB lies well within the standard error of the regression.

Conclusions

By using propagation speeds based on the average seawater sound-speed profile, the effective source of the maximum level of waterborne sound from an inland earthquake has been shown to occur near where the measured reverse azimuth intersected the edge of the continental shelf or slope.

By neglecting effects due to the size of seafloor features, shots at Fangataufa yielded a regression for the intensity of long-distance echoes scattered by seafloor features as a function of deflection angle, at the frequency of 16 Hz. This regression will be applicable to prominent features. The results from Mururoa did not yield a regression, and this may be attributed to the presence of a seamount close to Mururoa that blocked the direct path and paths to features at small angles.

A further detailed analysis involving the size of the features that give rise to individual echoes would permit the dependence of RI on feature size to be determined. It should be possible to treat this aspect theoretically, using the adiabatic-mode approximation for propagation in a range-dependent environment.

Appendix: Results for wave propagation speeds

A by-product of this study has been the derivation of average propagation speeds of waterborne waves along particular paths. Propagation speed depends on the SSP in the water-column (which generally varies with latitude), SFD, and acoustic frequency. For a given scenario, the variation of propagation speed with frequency can be computed, since it corresponds to the group speed of the appropriate normal modes. Deep-water SSPs contain a minimum, referred to as the SOFAR axis. If SFD is double the axis depth or more, the sound-speed minimum forms an acoustic waveguide, along which sound propagates with little loss. The speed of waves propagating in the SOFAR waveguide will generally increase as SFD increases, and approach an asymptote when SFD is too large to affect their properties (around 3 or 4 km, at most latitudes). This asymptote will depend on frequency, and at high frequencies (100 Hz or more) will be in the neighbourhood of the minimum sound-speed.

The average SSP for latitude -27° in the western South Pacific Ocean, has a minimum of 1.485 km/s and the SOFAR axis is 1.2 km deep. The average SSP for latitude 5° in the central Indian Ocean has a minimum of 1.492 km/s, and the SOFAR axis is 1.8 km deep. These sound-speeds were computed using Mackenzie's [12] function of temperature, salinity and depth; and the appropriate temperatures and salinities were obtained from the Levitus World Ocean Atlas [13].

MA-FA to New Zealand

Since the start times of the recordings of the nuclear shots in 1995/96 were not noted precisely [5], they cannot be used to determine a propagation speed. In 1987 however, nuclear shots at MA were recorded at GBI with accurate timing, and the travel time of the westernmost shot (the least inland transit) was observed to be 52.18 minutes. Since the distance was 4668 km, the average propagation speed was 1.491 km/s. From bathymetry data for the region [11], the average SFD along the path is approximately 5 km.

Properties of the normal modes have been computed with the ORCA normal-mode algorithm [14], at the (spectral peak) frequency of 20 Hz. The scenario was defined as follows:

- SSP for latitude -27°
- SFD of 5 km.
- seabed comprised of sand (at 5 km, the effect of this will be negligible)

The group speed is 1.484 for Mode #1, increases with Mode Number to 1.491 (at Mode #23), and then decreases for higher modes. Thus the speed derived from the measured travel time of the beginning of the direct pulse is the same as the group speed of the fastest mode.

H08 North

The second CTBTO time segment of the earthquake signal at the North site yielded a local propagation speed of 1.474 km/s. Among the three hydrophones, SFD varies between 2.3 and 2.4 km.

For ORCA, the scenario was defined as follows:

- SSP for latitude 5°
- SFD of 2.35 km.
- seabed comprised of a 100-m mud layer over a basalt basement [15]

At this SFD, the group speed of Mode #1 increases from 1.487 at 4 Hz to 1.493 at 16 Hz. The modes that have group speeds closest to the observed value (1.474) are #2 at 4 Hz, #5 at 8 Hz, and #9 at 16 Hz. It is not known at present why these modes would be dominant.

Effective Source Position to H08 North.

Properties of the normal modes have again been computed with ORCA, at frequencies of 4, 8 and 16 Hz. The scenario was defined as follows:

- SSP for latitude 5°
- SFDs of 0.2 to 5 km.
- seabed comprised of sand

For this SSP (with a blunter minimum than the Pacific SSP), group speed decreases monotonically with mode number. For each frequency, the theoretical group speed of the important mode is high for SFD less than 0.2 km and has a minimum in the region of 0.5 km. Once the SFD reaches 3.5 km, the group speed is close to its asymptote. In water deeper than 3.5 km, the group speeds of the modes that appear to be dominant at H08 North are all 1.491 km/s.

In computing the travel time of the waterborne wave from trial effective source locations to H08 North, the group speed at 4 Hz was selected, as being the strongest and most comprehensive. For each trial position, the SFD was obtained from TOPEX topography [11] along the 355.2° line from H08 North.

H08 South

CTBTO's two segments of the earthquake signal at the South site yielded local propagation speeds of 1.478 and 1.479 km/s. Among the three hydrophones, SFD varies between 1.8 and 1.9 km.

Properties of the normal modes have again been computed, at frequencies of 4, 8 and 16 Hz. The scenario was the same as used at H08 North, except that SFD was set to 1.85 km. The modes that have group speeds closest to the observed value (1.479) are #1 at 4 Hz, #3 at 8 Hz, and #6 at 16 Hz. Again, it is not known why these particular modes would be dominant.

Effective Source Position to H08 South via Seamount.

The average SFD for the 3149 km from the Effective Source to the seamount is 3.6 km, and for the 176 km from the Seamount to H08 South is 2.9 km, giving an overall average of 3.5 km.

Frequency (Hz)	4	8	16
Dominant mode at H08S	#1	#3	#6
Group speed at SFD = 1.85 km	1.480	1.484	1.481
Group speed at SFD = 3.5 km	1.492	1.492	1.492

Acknowledgements

The New Zealand data were acquired by Ralph Marrett of the Defence Technology Agency, Auckland, New Zealand; and provided by both him and Lindsay Hall.

For the Indian Ocean data, Martin Lawrence of CTBTO provided precise hydrophone positions. Travel time differences for both North and South hydrophone triplets were produced by John Newton, and provided by Martin Lawrence.

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