Comparison of Teleseismic- and Hydroacoustic-Derived Earthquake Locations along the North-Central Mid-Atlantic Ridge and Equatorial East-Pacific Rise

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1. Introduction

This letter examines earthquake locations derived using independent teleseismic and hydroacoustic datasets for events along the north-central (15-35° N) Mid-Atlantic Ridge (MAR) and equatorial (10°S-10°N) East Pacific Rise (EPR). It represents the first large-scale comparison of such data, providing ground-truth information for the teleseismic locations and insight into the process of T-wave generation. An assessment of location accuracy and catalog completeness in these remote ocean settings is critical for explosion monitoring efforts associated with the Comprehensive Test-Ban Treaty (CTBT), as well as seismo-tectonic studies at mid-ocean ridges (MORs).

During the last decade the value of hydroacoustic studies in monitoring MOR seismicity has been demonstrated through work using the U.S. Navy’s SOund SUrveillance System (SOSUS) (e.g., Fox et al., 1994) and arrays of moored autonomous underwater hydrophones (AUHs) (e.g., Fox et al., 2001; Smith et al., 2002). These studies utilize seismically-generated Tertiary (T) waves that propagate within the ocean’s SOund Fixing And Ranging (SOFAR) channel (Tolstoy and Ewing, 1950). In locating submarine earthquakes, the main advantages of the hydroacoustic method stem from the efficiency of this low-velocity waveguide, which allows for the detection of much smaller events at longer ranges relative to that possible with waves that travel through the
solid earth, as well as from the existence of a well-defined ocean sound-speed model (e.g., Teague et al., 1990). Although T-waves may be produced within regions of shallow sloping bathymetry near oceanic islands, continental shelves and subduction zones (e.g., Shurbet and Ewing, 1957; Johnson and Norris, 1968; Talandier and Okal, 1998), here we examine their generation in association with shallow hypocenter events within a deeper-ocean or abyssal environment (e.g., Johnson, et al., 1968; Fox et al., 1994). In this setting, the scattering of energy from a rough seafloor has emerged as the most likely mechanism for generating T-waves (de Groot-Hedlin and Orcutt, 1999, 2001; Park et al., 2001).

T-wave derived earthquake locations, henceforth referred to as T-wave epicenters, therefore reflect the source region of the scattered energy. Geologic observations in the MOR environment indicate a close spatial association between the epicenters of shallow-hypocenter earthquakes and the locations derived from the time of the peak energy arrival within the T-wave signal. In several instances, T-wave location accuracy has proven sufficient to lead field parties to sites of active seafloor volcanism (e.g., Fox, 1995; Dziak and Fox, 1999) and to facilitate many important structural/tectonic interpretations that would not have been possible using the teleseismic record (e.g., Fox and Dziak, 1999; Dziak et al., 2000; Bohnenstiehl et al., 2002). Fox et al. (2001) has quantified T-wave location errors using a set of Monte Carlo simulations for the equatorial Pacific region; their modeling suggests an accuracy of < 4 km (95% confidence level) in latitude and longitude for acoustic point-sources within the array. This level of accuracy is consistent with observations; however, it should be noted that the source region likely encompasses several km² of seafloor.
Although the International Monitoring System (IMS) is installing both seismic and hydroacoustic sensors in support of the CTBT, the low-density hydroacoustic network, comprising six hydrophones and five island T-phase stations deployed globally, is designed primarily to identify the source signature of explosions (Talandier and Okal, 2001). As such, the accuracy of teleseismic locations will remain critical for event assessment in the oceans. T-wave data may provide a means by which to assess location accuracy in some remote ocean settings, where more conventional ground-truth measures, such as mining explosions or observations of surface deformation, do not exist.

2.0. Data Collection and Processing

2.1. Teleseismic Data

Teleseismic location and arrival data are taken from the Reviewed Event Bulletin (REB), which is currently produced by the International Data Centre (IDC) in Vienna, Austria. Prior to 21 February 2000, this catalog was generated by a predecessor organization, known as the prototype IDC or pIDC. Waveform data were collected by the IMS seismic network, which contains a combination of 3-component seismic stations and multi-instrument seismic arrays (Figure 1). For the latter, signal-to-noise ratios are enhanced through beam forming. A subset of these sensitive array stations routinely detects sub-magnitude 4.0 earthquakes along the northern MAR and equatorial EPR; consequently, the (p)IDC is the agency that locates the largest number of earthquakes in these areas. However, the limited number of array stations in the southern hemisphere often results in a less than optimal azimuthal distribution of arrivals.

The (p)IDC uses the ratio between a short-term energy and long-term energy to detect arrivals. Initial arrival time estimates are then refined using a more sophisticated
procedure based on auto-regressive modeling of the signal and/or background energy (International Data Centre, 1999). At array stations, signals are classified as teleseismic P, regional P, regional S or noise based on their slowness and the quality of the frequency-wavenumber analysis. For 3-component stations, these signal types are assigned based on a set of rules that examine the frequency, horizontal-to-vertical ratio and rectilinearity of the signal. Compatible phases are associated using a grid-based approach and can be renamed (beyond P, Pn, Sn) during this procedure (International Data Centre, 1999). Hypocentral locations are determined through an iterative non-linear least-squares inversion of travel time, azimuth and velocity data. An analyst subsequently reviews the results of this automated processing.

REB body-wave magnitudes ($m_b$) will be reported in this manuscript, unless otherwise noted. A negative bias in REB magnitude, relative to the NEIC, has been reported previously (Wuster et al., 2000 and Granville et al. 2002). This bias can be attributed to the (p)IDC’s high-pass filtering procedure and their use of a relatively narrow window length in calculating $m_b$ (Granville et al., 2002).

2.2. Hydroacoustic Data

In mid-1996, an array of six AUHs was moored on the flanks of the EPR, monitoring a region between ~10° S-10° N (Figure 1). In early 1999, a similar array was deployed on the flanks of the MAR, monitoring a region between ~15-35° N (Figure 1). The arrays have been operating nearly continuously since their initial deployments, with semi-annual to annual recovery operations. In November 2000, a seventh hydrophone was deployed near 12° N, 95° W and used in processing arrivals in the Pacific though November 2001. Each instrument consists of a single hydrophone sensor floated within
the sound channel axis at a depth of ~700-1100 m and tethered via an acoustic release to a seafloor anchor. The recording package consists of a filter/amplifier stage designed to prewhiten the ambient noise spectrum, an accurate (< 1 sec/yr drift) clock that is GPS synchronized prior to deployment, a logging computer and multiple hard disks for data storage. The systems deployed were programmed to record 8-12 bit resolution (typically 8 bit) at a sample rate of 100-250 Hz (typically 110 Hz). Recovered and processed data from February 1999-February 2001 within the Atlantic and from June 1996-November 2001 within the Pacific are considered within this report. During this period all six instruments are available continuously in the Atlantic and between five and seven instruments are available in the Pacific. The U.S. National Oceanic and Atmospheric Adminstration and National Science Foundation fund these monitoring efforts, which are ongoing with annual turn-around cruises to recover and re-deploy the instruments. T-wave derived epicenters and hydroacoustic waveform data (CSS3.0 format) can be retrieved via a web-based interface (http://www.pmel.noaa.gov/vents/acoustics/seismicity/seismicity.html).

To associate T-wave arrivals with earthquakes within the REB, hydroacoustic data were aligned to the predicted T-wave arrival time based on the REB location and origin time. To minimize bias associated with the width of the T-wave source region, the peak energy arrival is used as the arrival time of the T-wave signal. It has been suggested, based on the correlation of event depth and rise time in the MOR setting, that this portion of the signal radiates from the near epicentral region, where the amplitude of the scattered energy is largest (Dziak et al., 1995). For SOSUS-detected T-wave events
in the NE Pacific, Slack et al. (1999) has shown that this peak arrival procedure produces no systematic bias relative to regional P-wave derived locations.

Following the identification of T-wave arrivals on four or more hydrophones, an epicentral location and origin time are derived using an iterative nonlinear Gradient-Expansion (Marquardt) algorithm (Fox et al., 2001). This procedure minimizes the following:

\[
\sum_{i=1}^{N} w_i (a_i - b_i)^2
\]

where \( w_i \) is the weight, \( a_i \) is the arrival time and \( b_i \) is the predicted arrival time at hydrophone \( i \) (Fox et al., 2001). \( N \) is the total number of recording hydrophones. The arrival time is calculated as: \( b_i = t + d_i/c_i \), where \( t \) is the origin time, \( d_i \) is the distance between the epicenter and hydrophone, and \( c_i \) is the sound speed along the path derived from the Generalized Digital Environmental Model (GDEM) of oceanic sound speeds (Teague et al., 1990).

Since the number of T-wave arrivals is small (typically \( \leq 6 \)), it is difficult to produce well-constrained error ellipses using standard seismological methods. Consequently, a calibrated Monte Carlo technique has been used to constrain the latitudinal and longitudinal uncertainties for T-wave derived locations (Slack et al., 1999; Fox et al., 2001). This procedure generates model events and randomly induces an arrival time (pick) error, which is drawn from a normal distribution with a one-second standard deviation, at each hydrophone. The error field is defined through repeated simulation within the GDEM-derived sound speed model. The results of these simulations suggest an accuracy of \(< 4 \) km (95% confidence interval) in latitude and
longitude for earthquakes within the two arrays, provided ≥ 4 hydrophones are used in the location (Fox et al., 2001; Smith et al., 2002).

Presently no depth or focal mechanism information is recovered from the T-wave signal. Recent models, however, have shown that the efficiency of T-wave generation through scattering should exhibit mechanism dependence, with strike-slip events being more efficient at generating T-waves than dip-slip events (Park et al., 2001). This has been confirmed observationally using SOSUS data from the northeast Pacific region (Dziak, 2001). The hypocentral depth of near-ridge earthquakes can be reasonably constrained, by a combination of previous seismic studies (e.g., Solomon et al., 1988; Toomey et al., 1988) and thermal/frictional arguments (Cowie et al., 1993), to lie within < 10 km of the seafloor. For the large magnitude earthquakes considered in this study, many of the 8-12 bit hydrophone sensors have clipped T-wave arrivals. Consequently, we will forgo a quantitative discussion of T-wave source strength.

3. Location Comparisons and Statistical Analysis

To ensure the most accurate AUH locations, only REB events located within the arrays were considered, as defined by the following approximate geographic bounds, 15-35° N, 55-35° W in the Atlantic and 10° S-10° N, 110-95° W in the Pacific. During the 2-yr period of the Atlantic deployments, 112 out of 114 REB events could be correlated with a sufficient number of T-wave arrivals to be located. In the Pacific, 400 out of 409 events could be correlated during the ~5.5-yr period considered. In order to better assess the completeness of the T-wave catalog, these uncorrelated events are discussed in Section 4.0.
REB and AUH epicenters are shown in map view for the Atlantic and Pacific in Figures 2 and 3, respectively. In the Atlantic, several events that appear to be intra-plate earthquakes within the REB catalog have AUH locations on or near the MAR axis. In the Pacific, many apparently intraplate events relocate on the EPR transform faults or the intermediate-spreading rate Galapagos Rift. The close association of earthquakes with an active plate-boundary zone of finite width is consistent with numerous morphologic and seismic observations, which suggest that most active faulting occurs within < 15-30 km of the ridge axis (e.g., Edwards et al., 1991; Lee and Solomon, 1995; Searle et al., 1998; Bohnenstiehl and Carbotte, 2001; Smith et al., 2002).

As shown in Figures 4 and 5, the magnitude of the location difference ($\Delta_{\text{loc}}$) is dependant on the size of the earthquake (or equivalently the number of teleseismic detections) and maximum azimuthal gap; however, $\Delta_{\text{loc}}$ does not show a dependence on the number of hydrophone detections, with all events detected on $\geq 4$ AUHs. In the Atlantic, 90% of the REB arrivals lie within 0.75 degrees (equivalent to a 9,273 km$^2$ circular region) of their corresponding AUH locations, with a 50% quantile of 0.20 degrees (an equivalent 2,473 km$^2$ circular region). In the Pacific, 90% of the REB arrivals lie within 2.30 degrees (equivalent to a 28,428 km$^2$ circular region) of their corresponding AUH locations, with a 50% quantile of 0.66 degrees (an equivalent 7,419 km$^2$ circular region). The magnitude dependence of these results is summarized in Table 1. The larger location differences observed in the Pacific relative to the Atlantic are consistent with the larger azimuthal gaps and small number of stations typically available to the IDC when locating earthquakes of a given magnitude within the Pacific (Figures 4 and 5).
Directional bias may be examined using circular statistics (Fisher, 1993), which disregard the magnitude of the relative location vectors and consider only the azimuth between the AUH and REB epicenters. For a population of $M$ azimuths ($\theta_1, \theta_2, \theta_3, \ldots, \theta_M$), the mean resultant azimuth and length are given by:

$$<\theta> = \tan^{-1}\left(\frac{x_r}{y_r}\right) \quad (2) \quad \text{and} \quad \tilde{R} = \frac{\sqrt{x_r^2 + y_r^2}}{M} \quad (3),$$

respectively, with $x_r = \sum_{i} \sin \theta_i$ and $y_r = \sum_{i} \cos \theta_i$, where $\theta_i$ is the azimuth from the AUH to the REB epicenter for event $i$. $\tilde{R}$ has a range from 0 to 1, where 1 indicates that all directions are identical and values $<1.0$ indicate increasing dispersion. The resultant length is related to the circular variance as: $S_o^2 = 100(1-\tilde{R})$, and so larger values of $\tilde{R}$ indicate less variance within the distribution of azimuths.

In the Atlantic the resultant vector has a mean azimuth of $338.2 \pm 15.9^\circ$ (2$\sigma$) and a length of 0.45 ($S_o^2 = 55\%$). In the Pacific, the resultant vector has a mean azimuth of $181.9 \pm 21.8^\circ$ (2$\sigma$) and a length of 0.21 ($S_o^2 = 79\%$). Rayleigh’s trend test (Swan and Sandilands, 1995) indicates a non-random azimuthal distribution for both datasets (at the 95% confidence level); however, the large circular variances do not argue for a systemic correction to the REB locations. The magnitude dependence of the circular statistics is summarized in Table 1.

In Figure 6, $\Delta_{loc}$ is depicted as a function of time using a five-event median average for earthquakes with REB $m_b \geq 4.0$. As the IMS network has grown through time and the IDC continues to refine methods, models and procedures, significant increases in location accuracy have been noted within some regions of the world (e.g., Wuster et al., 2000). Inspection of Figure 6, however, suggests only a marginally
significant decrease in $\Delta_{\text{loc}}$ within the EPR and MAR regions during the periods
examined. The addition of IMS hydroacoustic stations within the Atlantic and Pacific
Oceans, with real time data transfer to the IDC, will facilitate joint seismic-hydroacoustic
processing of earthquake/explosion data and may provide for future improvements in
location accuracy within these regions. In the Pacific, IMS T-phase seismic stations (see
Okal, 2001) are planned on Socorro and Queen Charlotte Islands, with a hydrophone
stations near Wake and Juan Fernandez Islands. IMS stations in the Atlantic will include
island T-phase stations on the Flores, Guadaloupe and Tristan da Cunha Islands, and a
hydrophone installation near Ascension Island.

4. REB Earthquakes Not Detected by the Hydroacoustic Arrays

4.1. Atlantic Array

Out of the 114 earthquakes considered in the Atlantic, two REB earthquakes
could not be detected on a sufficient number of hydrophones to be located. One of these
events is associated with a pair of earthquakes having close temporal ($\sim$20 s) and spatial
separation (located near 16.2° N, 46.7 W). In this instance, inspection of the hydrophone
data shows a composite T-wave packet, where the analyst could not identify two distinct
arrivals.

The second uncorrelated event was located slightly to the west of the array near
28° 30’ N, 54° 20’ W by the (p)IDC. Although both P and T-wave arrivals were detected
on the central-west (CW) hydrophone, the T-wave strength is noticeably less than
typically observed for a 4.0 $m_b$ event at a distance of 5.0° (Figure 7). The next closest
instruments, the NW and CE hydrophones, show no detectable arrivals. This event differs
from the near-ridge earthquakes examined in that it occurs on ~ 75 Ma lithosphere, where
deeper hypocentral depths are permissible and a heavily-sedimented seafloor resides at a depth >5,500 m below sea level. These factors may combine to suppress the amplitude of the generated T-wave (de Groot-Hedlin, 1999, 2001; Park and Odom, 2001). We have identified at least one additional off-axis event of this nature in the February 2001-2002 Atlantic data, which is shown in Figure 7d (Note, due to multiple instrument failures these 2001-2002 data are not included in our location comparisons).

4.2. Pacific Array

Out of the 409 earthquakes considered in the Pacific, nine earthquakes could not be detected on a sufficient number of hydrophones to be located. One of these events is associated with a pair of earthquakes that occurred ~19 s apart and were nearly co-located with an epicenter near ~2º N, 97º W. This situation is analogous to that previously described in the Atlantic, and a composite T-wave packet is observed.

Two of the unmatched events where subsequently reviewed by the International Seismological Centre (ISC, 2001) and located well outside of the array. Of the remaining six events, which were not reviewed, two are accompanied by “low confidence location” comments in the REB catalog, suggesting they also may be significantly mislocated. The other four were located using a combination of P and PKP arrivals recorded at only 3-5 stations. Due to the faster spreading rate of the EPR, the seafloor is notably younger (<20 Ma) within the array, with less sediment cover and shallower depths (<4000 m) relative to the 53-55º W area within the Atlantic array. Moreover, no P or S-wave arrivals are detected on the closest hydrophones, suggesting that our failure to correlate these events within the AUH record may reflect their poor seismic locations, rather than the suppression of T-wave generation—as observed in the Atlantic.
5. Consistency between AUH Locations and Teleseismic Arrival Time Data

To examine the consistency between teleseismic arrival time observations and AUH-derived locations, the P-wave travel times from each AUH epicenter to each detecting seismic station were calculated using the IASP91 velocity model. Then, for each event, independent estimates of the origin time ($OT_{ij}$) were calculated based on these travel times ($TT_{ij}$) and the arrival times ($AT_{ij}$) listed in the REB catalog.

$$\text{OT}_{(ij)} = \text{AT}_{(ij)} - \text{TT}_{(ij)} \quad (4),$$

where $i$ indicates the $i_{th}$ event and $j$ the $j_{th}$ station. The standard deviation ($std$) of these origin time estimates and the station residuals ($res$) were tabulated as:

$$std(OT)_i = \sqrt{\frac{\sum_{j=1}^{N} (\text{OT}_{(ij)} - \langle \text{OT}_{(ij)} \rangle)^2}{N - 1}} \quad (5)$$

and

$$res(OT)_j = \text{OT}_{(ij)} - \langle \text{OT}_{(ij)} \rangle \quad (6),$$

respectively.

For comparison, identical calculations were made using the REB epicenters. Since we are interested in the relative consistency of the calculated origin times and travel time data and not attempting to relocate the events, earthquake depth was fixed at 0.0 km (IDC default depth), and no station corrections were applied in either calculation. By examining the standard deviation of the origin time estimates and station residuals, potentially erroneous phase associations can be identified. Note that T-wave origin times were not considered in the above calculations, since they represent the time at which energy was radiated into the water column, rather than the time of the earthquake’s initiation.

5.1. Atlantic Array
Figure 8a shows a histogram of $std(OT)$, with origin times calculated using the AUH-constrained locations of 112 earthquakes within the Atlantic. The mean of these origin time standard deviations is 1.40 s. The $std(OT)$ values are observed to be lower, with a mean of 0.94 s, when origin times are calculated using the full set of P-wave arrivals and the REB-constrained locations. This is not surprising, since the (p)IDC determined the REB locations by minimizing seismic travel times within the IASP91 model. Still, the AUH locations, which minimize the residual of the T-wave arrivals, result in only a slight increase in the residuals of the seismic observations in most cases. When the maximum azimuthal gap is large, large $\Delta_{loc}$ values can be accommodated with little increase in the $std(OT)$ of an event. This reflects the trade-off between origin time and location for teleseismic events with large azimuthal gaps.

Outliers appear to result from poorly associated phase(s) at azimuthally isolated stations, as we demonstrate for the event with the largest AUH-constrained $std(OT)$ (~12 s) and largest $\Delta_{loc}$ (~440 km) in the Atlantic. The REB location places the event to the west of the plate boundary zone ($34.71^\circ$ N, $43.92^\circ$ W) on the North American Plate (Figure 1c). The pattern of T-wave arrivals across the array is inconsistent with this location, and the AUH-derived epicenter places the earthquake on axis ($31.81^\circ$ N, $40.70^\circ$ W).

Using the five P-wave arrivals within the REB, large residual errors are recovered when the origin time of the event is calculated using the AUH-derived locations and IASP91 travel times. A large positive residual (late calculated origin time, relative to the event mean) is found for the European station GERES, while negative residuals (early calculated origin times) are associated with the remaining four North American stations.
By excluding the GERES station, the AUH location and teleseismic arrival times appear reasonably consistent, with \( std(OT) = 1.25 \text{ s} \) (Table 2). In this instance, the association of a phase arrival at GERES served to shift the event’s epicenter to the west of the ridge axis and resulted in a later estimated origin time. During ISC review, seismic phases from two additional North American stations were associated with this event. However, this had little effect on the computed seismic location, since their inclusion did not increase significantly the spatial distribution of stations. For the AUH-constrained calculations (Figure 8a), the remaining outliers with \( std(OT) > 3.0 \text{ s} \) can be attributed to azimuthally isolated phase picks and circumstances similar to that described above.

5.2. Pacific Array

Figure 8b shows a histogram of \( std(OT) \), with origin times calculated using the AUH-constrained locations of 400 earthquakes within the Pacific. The mean of these origin time standard deviations is 2.2 s. As in the Atlantic, for many events the AUH locations result in only a slight increase in the \( std(OT) \), relative to that calculated using fixed REB epicenters (0.94 s mean). Outliers, which are noticeably larger relative to the Atlantic, also appear to result from incorrectly associated phases. As evident in Figure 5a, there is a concentration of large \( \Delta_{loc} \) values in the \(~200^\circ\) gap range. These represent a set of phase picks from North American stations, which are dominantly array detections, combined typically with a single arrival from a 3-component station in South America.

For example, one of the largest \( std(OT) \) values for an AUH-constrained location (26.05 s) is associated with a 3.7 \( m_b \) earthquake on 20 September 1999 (Figure 1b). P-waves were reported for three North American array stations and one three-
component South American (BDFB) station. Using the AUH location, a large positive origin time residual (late calculated origin time, relative to the event mean) is found for the South American station, with negative residuals for the North American stations (Figure 1; Table 3). Excluding the BDFB station, teleseismic and hydroacoustic data are consistent with one another, suggesting that the association of this phase has shifted the event’s epicenter westward from its AUH location on the western Galapagos Ridge onto the Pacific plate.

A second example of this phenomenon is given in Table 4 (Figure 1). Many of the large \( \text{std}(OT) \) values in the Pacific (Figure 8b) can be attributed to similar circumstances with inconsistent phase arrivals at azimuthally isolated South American stations.

6. Summary and Conclusions

In general, AUH locations are more closely associated with the plate boundary, relative to REB locations. In the Atlantic, most large magnitude earthquakes (i.e., those of a sufficient size to be detected by land-based seismic stations) are associated with the slow-spreading MAR axis. In the Pacific they are associated with major transforms along the fast-spreading EPR and the axis of the intermediate spreading rate Galapagos Ridge. The near absence of teleseismic events along the EPR crest reflects the thinness of the brittle layer in this region (Cowie et al., 1993).

The difference between AUH and EPR epicenters increases with increasing azimuthal gap and decreasing seismic magnitude (or number of seismic stations used in the location). The AUH locations thus appear to be more accurate, as inferred from previous
error simulations (Slack et al., 1999; Fox et al., 2001) and geologic observations (e.g., Fox, 1995; Bohnenstiehl et al., 2002).

In some instances the AUH locations appear inconsistent with a single REB arrival, which may represent an incorrectly associated phase. North American array stations account for a large number of the P-wave arrivals in both the Atlantic and Pacific; however, they provide similar azimuthal paths. When these otherwise large azimuthal gaps are split by a single station in either Europe/Africa for an Atlantic event or South America for a Pacific event, the quality of that single station arrival can play a critical role in determining the accuracy of the seismic location. Relative to the Atlantic, Pacific earthquakes are characterized typically by a larger azimuth gap and a smaller number of stations used in determining the REB location. Consequently, larger location differences are observed in the Pacific.

Azimuthal differences are non-random in both the Atlantic and Pacific regions, with resultant vectors trending to the north-northeast (338.2 ± 15.7°) of the AUH locations in the Atlantic and to the south (181.9± 21.8°) in the Pacific. However, the circular variances are large, 55% and 79% respectively, and therefore a constant direction of offset should not be assumed and no simple correction to the seismic location can be applied.

Although all near-axis REB earthquakes could be located using the AUH data, observations associated with two intra-plate earthquakes in the Atlantic suggest that these events may generate relatively weaker T-wave signals. This is consistent with recent airgun observations in the Indian Ocean, which indicate that a rough seafloor is required to efficiently scatter energy into the SOFAR within the abyssal setting (Harben et al.,
Greater seafloor and potentially hypocentral depths also may have contributed to the suppression of T-wave amplitude (de Groot-Hedlin and Orcutt, 2001; Park et al., 2001).

The rise time and length of the T-wave coda may inhibit the hydroacoustic detection of some large magnitude events with small (<20 s) inter-event times. When such events are nearly co-located, a composite T-wave packet may be observed and only one event will be reported in the T-wave catalog. This situation does not occur frequently, being observed in <0.5% of our observations. For smaller magnitude events, the T-wave coda is shorter and individual events can be discerned clearly from earthquake pairs with small inter-event times. The distribution of inter-event times in the Atlantic is discussed in Bohnenstiehl et al. (2003).
Acknowledgements: NOAA and NSF funded AUH data collection in the Pacific and Atlantic. H. Matsumoto and A. Lau developed much of the AUH hardware and software. These instruments were serviced at sea by NOAA/PMEL technicians, M. Fowler, J. Haxel and P. Will. E. Chapp assisted in reprocessing AUH data and provided comments on this manuscript. Discussions and collaborations with C. Fox, R. Dziak, D. Smith, P. Richards, W-Y. Kim, J. Granville, F. Graeber and P. Fribas were extremely helpful in completing this project. Comments by S. Hough were helpfully in improving the content and clarity of this manuscript. Work supported by CTBTO contract 01/2/20/258 (M.T.). This is LDEO contribution number 6489.
References Cited


FIGURE CAPTIONS

**Figure 1.** a) Map of IMS 3-component (triangles) and array (white circle) stations contributing phase arrivals for events in the study areas (gray open boxes). Hydrophone locations are shown as solid gray circles. Mid-ocean ridge spreading centers are shown as solid black lines. Only stations relevant to the discussion in Section 5 are labeled. Lower panels (b and c) show AUH (black dots) and REB (crosses) locations, as well as REB error ellipses (95% confidence) for earthquakes discussed in Section 5.0. These events are a 4.1 m$_b$ earthquake on 23 September 1999 on the north-central Mid-Atlantic Ridge (Table 2), a magnitude 3.7 m$_b$ earthquake on 1 August 1999 on the western Galapagos Ridge (Table 3), and a magnitude 3.8 mb earthquake on 23 July 1996 on the equatorial East Pacific Rise (Table 4).

**Figure 2.** Comparison of REB and AUH locations in the Atlantic. Solid lines drawn from REB location (crosses) to AUH location for 112 correlated events between February 1999 and February 2001. Open circles are hydrophone instrument locations. Slow-spreading (~25 mm/yr full rate) Mid-Atlantic Ridge plate boundary is shown in gray.

**Figure 3.** Comparison of REB and AUH locations in the Pacific. Solid lines drawn from REB location (crosses) to AUH location for 400 correlated events between June 1996 and November 2001. Open circles are hydrophone instrument locations. A seventh hydrophone near 12° N, 95° W was used to located events during the period November 2000- November 2001. The fast-spreading (~110 mm/yr full rate) East-Pacific Rise trends north-south and the intermediate spreading (~60 mm/yr full rate) Galapagos Ridge trends east-west, as shown by the solid gray line.
Figure 4. The absolute value of the location difference between the AUH and REB epicenters ($\Delta_{loc}$) in the Atlantic is plotted against REB location parameters: a) maximum azimuthal gap, b) $m_b$, c) the number of stations used in the location, d) minimum distance to a station. Symbols listed in key indicate the number of AUH instruments used in the hydroacoustic location of each event.

Figure 5. The absolute value of the location difference between the AUH and REB epicenters ($\Delta_{loc}$) in the Pacific is plotted against REB location parameters: a) maximum azimuthal gap, b) $m_b$, c) the number of stations used in the location, d) minimum distance to a station. Symbols listed in key indicate the number of AUH instruments used in the hydroacoustic location of each event.

Figure 6. Median location difference for five-event windows of $>4.0$ $m_b$ earthquakes shifted by one event in time. Only a slight decrease in $\Delta_{loc}$ is observed during the period examined. Best fitting slopes of $0.028 \pm 0.022$ (2$\sigma$) and $0.009\pm0.059$ (2$\sigma$) shown for the Pacific (crosses) and Atlantic (circles) data, respectively.

Figure 7. a) Signal traces and spectrograms for the three closest hydrophones to a 4.0 $m_b$ off-axis event near, 28.5° N, 54.7° W. P- and T-wave arrivals are detected on the center-west (CW) hydrophone at a distance of ~5.0°; however, the signal is not detected on other hydrophones within the array. The impulsive signal seen on the CW and north-west (NW) hydrophones is seismic airgunning. For comparison, b) and c) show higher amplitude T-waves generated by similar magnitude on-axis earthquakes recorded at similar distances. d) Signal trace and spectrogram for the closest hydrophones to a 4.2 $m_b$ (4.7 $m_b$ NEIC) off-axis event near 15.3° N, 53.7° W. P and T wave arrivals are detected with a low amplitude T-wave arrival.
Figure 8. Standard deviation of (a) Atlantic and (b) Pacific origin time estimates based on AUH-constrained epicenters, calculated travel times within the IASP91 velocity model and REB P-wave arrival times. The mean is shown for each group, with its standard deviation in brackets. Identical calculations were done using fixed REB epicenters, with mean values of 0.94 [0.36] s and 0.94[0.41] s observed in the Atlantic and Pacific.
Table 1: Statistical Summary of Location Comparison

<table>
<thead>
<tr>
<th></th>
<th>Atlantic</th>
<th></th>
<th></th>
<th>Pacific</th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
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</thead>
<tbody>
<tr>
<td></td>
<td>90-Q</td>
<td>50-Q</td>
<td>&lt;θ&gt;</td>
<td>R</td>
<td>r*</td>
<td>N</td>
<td>90-Q</td>
<td>50-Q</td>
<td>&lt;θ&gt;</td>
<td>R</td>
<td>r*</td>
<td>N</td>
<td>90-Q</td>
<td>50-Q</td>
<td>&lt;θ&gt;</td>
<td>R</td>
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<tr>
<td>All</td>
<td>0.75</td>
<td>0.20</td>
<td>338.2</td>
<td>±15.9</td>
<td>0.45</td>
<td>N 112</td>
<td>2.30</td>
<td>0.66</td>
<td>181.9</td>
<td>±21.8</td>
<td>0.21</td>
<td>N 400</td>
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<tr>
<td>≥ 4.5</td>
<td>0.26</td>
<td>0.15</td>
<td>267.4</td>
<td>±263.6</td>
<td>0.13</td>
<td>Y  6</td>
<td>0.44</td>
<td>0.14</td>
<td>70.3</td>
<td>±133.8</td>
<td>0.14</td>
<td>Y  18</td>
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<td></td>
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<tr>
<td>4.0-4.4</td>
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<td>0.14</td>
<td>340.6</td>
<td>±34.73</td>
<td>0.35</td>
<td>N 41</td>
<td>1.10</td>
<td>0.35</td>
<td>159.1</td>
<td>±82.9</td>
<td>0.09</td>
<td>Y 117</td>
<td></td>
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<tr>
<td>&lt; 4.0</td>
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<td>0.21</td>
<td>348.4</td>
<td>±17.6</td>
<td>0.54</td>
<td>N 65</td>
<td>2.69</td>
<td>0.91</td>
<td>187.0</td>
<td>±17.5</td>
<td>0.28</td>
<td>N 265</td>
<td></td>
<td></td>
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</tbody>
</table>

Q- percent quantile values of Δloc given in degrees
<θ> - resultant mean azimuth from AUH to REB epicenters, ± 2-σ limits (circular statistic)
R – resultant length of circular data (circular statistic)
r* – statistically random as determined by Rayleigh’s trend test (circular statistic, 95% confidence)
N – number of events

Table 2. Teleseismic arrival data for 20 September 1999 earthquake, north-central MAR region.

<table>
<thead>
<tr>
<th>Stat.</th>
<th>Dist. (deg)</th>
<th>Az. (deg)</th>
<th>Arrival time</th>
<th>SNR</th>
<th>mb</th>
<th>Origin time residual (s) a</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
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<td></td>
<td></td>
<td></td>
<td></td>
<td>REB</td>
</tr>
<tr>
<td>GERES</td>
<td>44.19</td>
<td>53.2</td>
<td>12:32:14.9</td>
<td>13.5</td>
<td>m</td>
<td>4.8</td>
</tr>
<tr>
<td>TXAR</td>
<td>50.31</td>
<td>281.4</td>
<td>12:33:02.4</td>
<td>---</td>
<td>m</td>
<td>---</td>
</tr>
<tr>
<td>PDAR</td>
<td>50.67</td>
<td>299.9</td>
<td>12:33:06.0</td>
<td>20</td>
<td>a</td>
<td>4.3</td>
</tr>
<tr>
<td>NVAR</td>
<td>58.33</td>
<td>297.3</td>
<td>12:34:01.6</td>
<td>7</td>
<td>a</td>
<td>3.6</td>
</tr>
<tr>
<td>ILAR</td>
<td>64.35</td>
<td>332.4</td>
<td>12:34:40.8</td>
<td>2.1</td>
<td>m</td>
<td>3.7</td>
</tr>
</tbody>
</table>

std(OT) 0.50 12.72 1.25

* Origin times calculated using REB or AUH epicenters, and IASP91 velocity model.

Table 3. Teleseismic arrival data for 1 August 1999 earthquake, western Galapagos Ridge.

<table>
<thead>
<tr>
<th>Stat.</th>
<th>Dist. (deg)</th>
<th>Az. (deg)</th>
<th>Arrival time</th>
<th>SNR</th>
<th>mb</th>
<th>Origin time residual (s) a</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>REB</td>
</tr>
<tr>
<td>TXAR</td>
<td>25.31</td>
<td>1.6</td>
<td>10:12:24.2</td>
<td>7.7</td>
<td>a</td>
<td>3.4</td>
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<tr>
<td>PDAR</td>
<td>38.97</td>
<td>354.0</td>
<td>10:14:25.0</td>
<td>7.4</td>
<td>a</td>
<td>3.5</td>
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<tr>
<td>BDFB</td>
<td>59.13</td>
<td>110.7</td>
<td>10:16:59.1</td>
<td>3.1</td>
<td>a</td>
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<tr>
<td>ILAR</td>
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<td>342.8</td>
<td>10:17:54.4</td>
<td>6.2</td>
<td>a</td>
<td>--</td>
</tr>
</tbody>
</table>

std(OT) 1.28 26.05 1.16

* Origin times calculated using REB or AUH epicenters, and IASP91 velocity model.
Table 4. Teleseismic arrival data for 23 July 1996 earthquake, equatorial EPR region.

<table>
<thead>
<tr>
<th>Stat.</th>
<th>Dist. (deg)</th>
<th>Az. (deg)</th>
<th>Arrival time</th>
<th>SNR</th>
<th>mb</th>
<th>Origin time residual (s) a</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
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<td></td>
<td></td>
<td></td>
<td>REB</td>
</tr>
<tr>
<td>TXAR</td>
<td>34.87</td>
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<td>LPAZ</td>
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<td>354</td>
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<td>a</td>
<td>-1.27</td>
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<tr>
<td>YKA</td>
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<td>a</td>
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<td>20:24:02.7</td>
<td>4.6</td>
<td>a</td>
<td>-0.51</td>
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</table>

std(OT) 1.26 13.59 0.48

a Origin times calculated using REB or AUH epicenters, and IASP91 velocity model.
Figure 2: Bohnenstiehl and Tolstoy
Figure 3: Bohnenstiehl and Tolstoy
Figure 4: Bohnenstiehl and Tolstoy
Figure 5: Bohnenstiehl and Tolstoy
Figure 6: Bohnenstiehl and Tolstoy
Figure 7: Bohnenstiehl and Tolstoy
Fixed AUH epicenters - MAR
\(<std(OT) >=1.40 [1.27]\)

Fixed AUH epicenters - EPR
\(<std(OT) >=2.22 [3.04]\)

Table 2 event
Table 4 event
Table 3 event

Figure 8: Bohnenstiehl and Tolstoy