Random Media Effects in Basin-Scale Acoustic Transmissions

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Abstract. Data from a 1000-km acoustic tomography experiment in the eastern North Pacific are used to illustrate the effects of random internal-wave-induced sound-speed fluctuations on basin-scale acoustic transmissions, and by extension to predict random media effects for the Acoustic Thermometry of Ocean Climate (ATOC) transmissions. Monte Carlo numerical simulations of acoustic propagation through random internal-wave sound-speed variations obeying the Garrett-Munk internal-wave spectrum are compared to the data, and they are used to predict internal wave effects for ATOC transmissions. For the 1000-km experiment the simulations are used to infer the range-average internal-wave spectral energy as a function of depth. Such an analysis on the ATOC data can provide useful measurements of internal-wave variability in the eastern North Pacific. In terms of basin-scale acoustic thermometry internal-waves place limitations on the vertical resolution that can be achieved in measuring large-scale ocean heating or cooling.

1. Introduction

The starting point for random media studies in deep ocean acoustics is generally the Helmholtz equation,

$$\frac{\partial^2 \psi(\vec{r}, \omega)}{\partial z^2} + \frac{\omega^2}{(\bar{c} + \delta c)^2} \psi(\vec{r}, \omega) = 0$$

where $\bar{c}$ is the deterministic ocean sound channel and $\delta c$ is a stochastic perturbation. Internal waves change the ocean sound-speed field primarily by their vertical displacements,

$$\delta c(\vec{r}) = \left( \frac{\partial \zeta}{\partial z} \right)_{p} \zeta(\vec{r})$$

where $\delta c(\vec{r})$ is the deviation of sound-speed from the case with no internal waves, $\vec{r}$ is the position, $\zeta$ is the internal-wave displacement and $\left( \frac{\partial \zeta}{\partial z} \right)_{p}$ is the potential gradient of sound speed. As a problem in wave propagation in random media, long range acoustic propagation in the ocean is distinctive in that the random internal-wave medium is combined with the deterministic effect of the ocean waveguide [Flatté, 1983b]. Furthermore, sound speed fluctuations caused by internal waves are anisotropic and inhomogeneous [Flatté et al., 1979].

The focus of this paper will be on describing pulsed signals which propagate over thousands of kilometers in the ocean, and particular emphasis will be placed on the tomographically important quantity of travel time; that is the focus will be primarily on the phase, not on the magnitude, of $\psi$. This emphasis naturally neglects several very interesting wave propagation issues (see Henry and Macaskill, 1996 for a recent review). The focus here is on how internal waves limit the ability to use acoustic travel times as a means of tomographically mapping the large scale heat content changes of the world oceans. Similarly there is an interest in using high-frequency acoustic fluctuations to learn about internal waves themselves at basin scales.

It has been suggested by several investigators that high-frequency acoustic fluctuations can be used to infer properties of the internal-wave field (see Munk [1981], Flatté [1983a], and Usserski [1986]). Experimental trails of these concepts however are few in number. Dushaw et al. [1995] and Brachet and Flatté [1997] used tidal-period wavefront fluctuations from different experiments to measure the deterministic internal tide. Stoughton et al. [1986] have provided the only analysis of wavefront travel-time fluctuations to measure rms internal-wave displacements and horizontal currents.

However, Stoughton’s analysis was limited by the fact that the measured acoustic fluctuations were recorded on a single hydrophone and therefore only one acoustic observable could be used, namely travel-time variance. In this paper several acoustic observables from the SLICE89 experiment (which used a 3-km vertical aperture) and from Monte Carlo simulations of acoustic propagation through random realizations of internal waves obeying the Garrett-Munk (GM) spectrum [Munk, 1981] are used to estimate the average strength of internal waves over a large depth region of the ocean. Because several observables are used, the known inadequacy of the GM model for acoustic energy which samples the upper ocean is revealed.

In addition, internal-wave effects on basin-scale acoustic transmissions are very important to large scale ocean acoustic tomography since internal waves impose the ultimate limitations on the measurement technique. This situation is similar to the limitations imposed upon ground based telescopes by atmospheric turbulence. Currently data is being collected by the acoustic thermometry of ocean climate (ATOC) network to provide
a measure of large scale ocean heating or cooling. A review of random media effects observed in the ATOC data will be given to illustrate vertical resolution limitations. An important issue relating to basin-scale tomography which will not be addressed is the issue of internal-wave induced travel-time bias. Finally, the ATOC data can also be used to measure internal-wave variability in the eastern North Pacific. A summary of this effort will be given.

The presentation of this paper is as follows. Section 1 discusses the estimation of internal-wave spectral energies from the SLICE89 experiment. Section 2 examines the tomographic vertical resolution of the ATOC transmissions and there is a discussion of internal-wave climatology. Section 3 has discussion and conclusions.

2. SLICE89 Data

The SLICE89 experiment was conducted in the eastern North Pacific (Figure 1) over a 10-day period in which narrow pulses were transmitted from a moored 250-Hz center frequency source, (source depth = 800 m), to a 3-km-long vertical array of 50 hydrophones 1000 km distant.

Pulses were sent once every hour except for a 21-hour period on day 6 in which pulses were sent every 10 minutes. Data from the SLICE89 experiment and the numerical simulations (Figure 2) are displayed by plotting the acoustic intensity as a function of depth and time. This is called a timefront. A timefront is the signal measured as a wavefront propagates past a vertical array of receivers at a fixed range.

Figure 1. Geometry of the SLICE89 experiment.

The timefronts in Figure 2 show two distinct regions in the arrival pattern. The energy which arrives at the earlier travel times shows fronts that form a double accordion pattern. This is a ray-like region. The ray-like region contains the data for a typical ocean acoustic tomography analysis because every point along the front can be associated with a geometrical optics ray that samples the ocean in a specific way. The earliest arrivals are composed of rays that reflect from the ocean surface and sample the deep ocean. Later fronts are composed of rays that sample the upper 300 meters of the ocean.

When ray-like fronts are no longer apparent at the late arrival times this is called the mode-like region. The mode-like region is composed of acoustic energy that is trapped near the sound-channel axis, or is described by low-order acoustic modes.

The upper panel of Figure 2 shows a timefront from a numerical simulation with internal-wave sound-speed perturbations obeying the GM internal-wave spectrum but at half the reference energy; the center panel is a measured pulse from the SLICE89 experiment; the lower panel is a numerical simulation without internal-wave sound-speed perturbations [Colosi et al., 1994]. For the early ray-like region the three timefronts are qualitatively the same: the data and the simulation with internal waves have small time shifts due to internal waves and the intensity varies along the fronts. However in the mode-like region, the data and the simulation with internal waves show significant broadening of the energy in depth compared with the case of no internal waves. It is in this region that the internal waves show their largest effect; this can be understood in terms of mode coupling [Colosi and Flatté, 1996] or ray refraction (see Simmen et al. [1997] and Duda and Bowlin [1994]). In the mode-like region the wavefront envelope can be used as an acoustic observable [Colosi et al., 1994], or mode arrival patterns can be used as observables [Colosi and Flatté, 1996]. Due to the sparse vertical array used in SLICE89, individual acoustic modes could not be resolved so the wavefront envelope is used. In the ray-like region two different measures of wavefront timing fluctuations are used as data for our estimate of internal-wave displacements.

2.1. Measurement of Internal Waves: The Final Arrival

The effects of mesoscale sound-speed perturbations on the wavefront envelope are treated first. Figure 3 shows the wavefront envelope for the SLICE89 data, a simulation using a range-independent sound-speed profile, and a simulation with the range-independent profile plus mesoscale perturbations. The mesoscale perturbation field was determined by from an objective mapping of environmental data from CTDs, XBTs, and AXBTs (see Worcester et al. [1994] and Cornuelle et al. [1993]). The range-independent sound-speed profile was calculated from the average of 11 CTD casts taken during the experiment (10 CTD’s went to 2000-m depth and one went to 4000-m depth). For the SLICE89 data and the numerical simulations with internal waves the wavefront envelope is defined to be the location in time/depth coordinates of the intensity level which is 2% of the maximum intensity of the incoherent average of all of the
Figure 2. Time fronts from the numerical simulations and the SLICE89 experiment. The upper/lower panels are simulations with/without internal-wave sound-speed fluctuations. The middle panel is a measured pulse from the SLICE89 experiment.

data (110 pulses are averaged for the SLICE89 data and 33 pulses are averaged for the simulation). The calculation of the wavefront envelope for the simulation using the range-independent profile and the mesoscale perturbations used the same reference intensity level as the case with internal waves.

Figure 3 shows that the observed mesoscale sound-speed field for the SLICE89 experiment has very little effect on the wavefront envelope, and therefore the mismatch between the data and the simulations without internal waves is a purely internal-wave effect. Since the eastern North Pacific is known to be a very quiet region for mesoscale activity, this result may not hold in other parts of the ocean.

Figure 4 shows the wavefront envelopes for the SLICE89 data, and the numerical simulations at 0.5 and 2.0 GM. It is clear that the 0.5 GM curve matches the data curve very well indicating a 0.5 GM level near the axis.

2.2. Measurement of Internal Waves: Wavefronts

For the ray-like arrivals travel-time variances have been calculated for individual fronts [Duda et al., 1992]. The broadband variance $\sigma^2(z_n)$ is a measure of travel time variations which take place on a time scale of less than 10 minutes, while the wander variance $\sigma^2_w(z_n)$ is a measure of the travel time variations which take place on longer timescales ($\approx$ 1 hour). Since ray travel times are very sensitive to internal-wave sound-speed changes at the upper apex of the ray [Platé and Stoughton, 1986] each wavefront segment is identified with the av-
Figure 3. Wavefront envelope observed from the SLICE89 experiment (solid), a calculated envelope using a range-independent sound-speed profile (dash-dot), and a calculated envelope using the range-independent profile plus range-dependent mesoscale perturbations (large dash).

Figure 4. Wavefront envelope observed from the SLICE89 experiment (solid), a calculated envelope using a range-independent sound-speed profile plus internal-wave realizations at 0.5 GM (dash-dot), and at 2.0 GM (large dash).

\[ k_m = 1 \text{ (cpkm)}, \text{ and } W_j(k, z) \text{ are solutions to the } \]
\[ \text{internal-wave eigenmode equation. Figure 5 shows the estimates of } \xi_{b, w}^2(z_u) \text{ and } \xi_{1, 0GM}^2(z_u) \text{ has been plotted for reference. For acoustic energy with turning points below 100 m, the variance estimates indicate internal-wave displacements that are from 0.4 to 1.0, the reference GM level, but for energy that samples the upper 100 meters the two different variance measures imply very different internal-wave displacements. This result shows that the spectrum of sound-speed variations that existed in the ocean near the surface was significantly different from the GM spectrum modeled in the numerical simulations.} \]

\[ \xi_{b, w}^2(z_u) = A \xi_{2, 0GM}^2(z_u) + B \xi_{0, 5GM}^2(z_u) \quad (3) \]

\[ A = \frac{\sigma_{1, b, w}^2(z_u) - \sigma_{0, b, w}^2(0.5, z_u)}{\sigma_{2, b, w}^2(2.0, z_u) - \sigma_{0, b, w}^2(0.5, z_u)} \]

\[ B = 1 - A \]

where \( \sigma_{1, b, w}^2(0.5, z_u) \) and \( \sigma_{2, b, w}^2(2.0, z_u) \) are the broadband \( (b) \), and wander \( (w) \) variances calculated from the numerical simulations at 0.5 and 2.0 GM, and \( \sigma_{0, b, w}^2(z_u) \) are the measured variances from the experiment. The quantities \( \xi_{2, 0GM}^2(z_u) \) and \( \xi_{0, 5GM}^2(z_u) \) are the variances of internal-wave displacement at 2.0 and 0.5 GM, calculated from

\[ \xi_{1, 0GM}^2(z) = \xi_e^2 \frac{N_G^2(z)}{N_G^2(z_e)} \quad (4) \]

\[ N_G^2 = \sum_{j=1}^{50} \sum_{k=k_0}^{k_m} |G_j(k)|^2 W_j^2(k, z) dk \]

where \( z_e \) is a reference depth at which the buoyancy frequency is 3 cpfh, \( |G_j(k)|^2 \) is the Garrett-Munk internal-wave displacement spectrum [Munk et al., 1981] in terms of vertical mode number \( j \), and horizontal wavenumber \( k \), \( \xi_e = 7.3 \text{ (m)} \), \( k_0 = dk = 1/32 \text{ (cpkm)} \), and \( k_m = 1 \text{ (cpkm)} \), and \( W_j(k, z) \) are solutions to the internal-wave eigenmode equation. Figure 5 shows the estimates of \( \xi_{b, w}^2(z_u) \) and \( \xi_{1, 0GM}^2(z_u) \) has been plotted for reference. For acoustic energy with turning points below 100 m, the variance estimates indicate internal-wave displacements that are from 0.4 to 1.0, the reference GM level, but for energy that samples the upper 100 meters the two different variance measures imply very different internal-wave displacements. This result shows that the spectrum of sound-speed variations that existed in the ocean near the surface was significantly different from the GM spectrum modeled in the numerical simulations.
2.3. Discussion

Stoughton et al. [1986] measured internal-wave displacements near Bermuda and found the standard GM level for rays which sample the upper 200 m of the ocean, and 0.5 GM for rays whose upper turning points sample the depth region 600 to 900 m. The Bermuda experiment did not have a vertical array nor were the transmissions rapid enough to separate the broadband and wander variances. Therefore to compare our results with the Bermuda experiment one must use the average of the wander and the broadband variance. Following the curves in Figure 5, a similar result to the Bermuda experiment is found that is near 0.5 GM for the main thermocline and near 1.0 GM in the upper ocean. However, because both the broadband and wander variances are measured, the inconsistency of applying the GM model to the upper 100 m of the ocean is revealed. Furthermore, the use of the wavefront envelope extends the measurement of internal waves into the deep ocean, giving more vertical resolution than was achieved by Stoughton. Clearly, however, the use of the wavefront envelope is sub-optimal and depends on the precise intensity cut off level used. Therefore in future experiments like the ATOC measurements, use of mode resolving arrays is essential so that modal travel times can be used as acoustic observables.

3. ATOC Data

Since December 1995 the Acoustic Thermometry of Ocean Climate (ATOC) program has been transmitting pulses at a center frequency of 75 Hz from a bottom mounted source (depth = 930m) on Pioneer seamount to several receivers throughout the North Pacific (See Figure 6). Receivers include two moored 40-element vertical-line arrays spanning 1400m (VLA1 and VLA2) and several bottom-mounted horizontal arrays operated by the Navy (alphabetical labels). Signals are sent every 4 hours during transmission periods established by the ATOC Marine Mammal program. A single pulse consists of 40 repeated transmissions which are coherently averaged to increase signal to noise levels.

The ATOC timeseries of basin scale travel times is being gathered as a means to test large-scale climate models, and as a measurement technique it is suited to complement satellite remote sensing of the ocean sea surface [Munk and Wunsch, 1982]. The promise of acoustic thermometry is to provide vertical resolution of ocean climate variability—resolution that a satellite cannot provide (horizontal resolution for satellites is excellent while for acoustic methods it is poor). Therefore in discussing internal-wave effects I focus on the issue of the limitations in vertical resolution imposed by internal waves.

In August of 1996 VLA1 and VLA2 were recovered so that their data could be analyzed. This section will focus on numerical simulations for VLA1 whose range from the source was 3515 km. Comparisons between the numerical predictions and the actual data will be discussed in future publications.

3.1. ATOC Timefronts

Two simulated timefronts for VLA1 are shown in Figure 7. The simulations are calculated using range dependent sound-speed profiles derived from the annual Levitus 1994 database. Internal-wave sound-speed fluctuations were calculated using the method of Colosi and Brown [1997], but variations in the internal-wave field due to changes in the buoyancy frequency profile and the local Coriolis Figure 6 the center frequency is 84 Hz and the 3 dB bandwidth is 15 Hz (for the ATOC source the center frequency is 75 Hz and the bandwidth is 37.5 Hz). These source parameters were chosen to compare with an alternate source which is discussed in Sec. 3.4. These parameters, however, are close enough to the ATOC source to make a comparison. The general character of the arrival pattern is very similar to the SLICE89 pattern (Figure 2) with early arriving stable ray-like energy with small travel time fluctuations and late-arriving, highly scattered mode-like energy.

3.2. Vertical Resolution: Rays

Acoustic thermometry relies on tracking the path of acoustic energy through the ocean from source to receiver. When wavefronts can be resolved in the ray-like region this acoustic energy path can be identified as the geometrical optics ray path. Each point on the wavefront has a specific ray path associated with it and these paths have a specific number of turning points.
Figure 7. Time fronts from numerical simulations at a center frequency of 84 Hz for the ATOC VLA1. The upper panel is a simulation using the Levitus 1994 sound-speed database plus a random realization of internal-wave sound-speed perturbations at the GM reference level, and the lower panel is a simulation without internal waves.

Figure 8. Vertical sampling of resolved eigenrays at VLA1 determined from numerical simulation. Solid curves give the range average upper and lower turning depths of eigenrays as a function of the number of ray turning points. Dash-dot curves give the minimum upper and maximum lower turning depths.

The effective acoustic path of mode 1 follows the sound channel axis, therefore this point serves as a useful constraint on sound-speed structure in this region (700-1000 m depth range). Figure 9 shows simulated mode 1 and mode 5 arrivals for VLA1. For mode 1 the simulation without internal waves shows a sharp peak where as the simulation with internal waves shows some time spread ($\approx 0.1$ s). This time spreading is the result of mode coupling caused by internal waves; that is mode 1 does not propagate from the source to the re-
Figure 9. A realization of the arrival patterns for modes 1 and 5 at VLA1 computed from the simulation data shown in Figure 7. The dash-dot/solid lines are simulated arrival patterns without/with internal-wave sound-speed perturbations.

A receiver in such a way as to maintain its identity. Mode 1 measured at the receiver is a complex superposition of energy which has propagated as other modes numbers having different group velocities. But for mode propagation in the deep ocean, mode 1 has the smallest group velocity, and therefore the last arriving energy in the mode 1 arrival pattern must be energy which has mostly traveled as mode 1 from source to receiver. Following Headrick [1997], this point can be called the pseudo-adiabatic mode 1 (PAM1) arrival which can be used in an adiabatic acoustic inverse [Worcester et al., 1997]. This PAM1 arrival is not strictly adiabatic, there is a travel time bias associated with it [Colosi and Flatté, 1996] and the bias is on the order of 0.1 s.

Use of higher order modes would also increase the vertical resolution of acoustic thermometry. From Figure 9 it is clear that mode 5 shows dramatic time spreading (≈ 1.0 s). A method for doing the acoustic inverse problem in situations of strong random mode coupling or non-adiabaticity remains a difficult research topic, and therefore making use of the higher modes at this frequency is a future endeavor.

Whether the ATOC data shows such dramatic mode coupling as predicted by these simulations will be revealed in later work.

3.4. Vertical Resolution: Lower Frequencies

In an effort to mitigate the effects of internal waves, an alternate source test was conducted in May 1996, in which dual frequencies of 28 Hz and 84 Hz were transmitted from a source suspended from a ship near Pioneer seamount. The 3-dB bandwidth for this source at both frequencies was about 10 Hz. Lower acoustic frequency means a larger acoustic wavelength which more effectively averages out small-scale internal-wave variability. Simulations of the arrival pattern at VLA1 for the 28-Hz signal are shown in Figure 10, which can be directly compared to Figure 7.

The ray arrivals for the low frequency case are about as clean as the higher frequency case, yielding a similar identification of wavefronts (Figure 8). The biggest difference between the two cases is for the modes.

Figure 10. Time fronts from numerical simulations at a center frequency of 30 Hz for the ATOC VLA1. The upper panel is a simulation using the Levitus 1994 sound-speed database plus internal-wave sound-speed perturbations at the GM reference level, and the lower panel is a simulation without internal waves.
Figure 11 shows the mode 1 and 5 arrival patterns at VLA1 for the 28-Hz signals. Comparing to Figure 9 it is clear that mode coupling effects are significantly reduced.

The mode 1 arrival is almost completely adiabatic, showing very little time spread. The mode 5 arrival shows fluctuations but the basic envelope of the calculation with internal waves matches the pattern of the no internal-wave calculation. Therefore, there is evidence that low-mode arrivals at 28 Hz are much more robust to internal waves than the 84 Hz case!

Again, whether the ATOC data shows this dramatic reduction in mode coupling will be revealed in later work.

3.5. Internal Wave Climatology

Clearly internal wave effects are very pronounced in basin-scale acoustic transmissions (particularly in the final arrival), and therefore high frequency acoustic fluctuations can be a powerful tool in understanding internal-wave variability in the North Pacific ocean basin.

The VLA1 and VLA2 timseries from December 1995 to August 1996 can provide many acoustic observables for an internal wave analysis. The observables include modal travel time spread from low mode numbers, identified ray travel time variances, and time and depth coherence. As discussed by Flatté [1983a] and Flatté and Stoughton [1986], travel time variance, time and depth coherence weight the internal wave spectrum in different ways, therefore these observables provide constraints on model spectra. For example the travel time variance for a ray can be written [Flatté and Stoughton 1986],

$$\tau^2 = \left(1/c_0^2\right) \int_0^R dx <\mu^2(z_r)> L_n(\theta, z_r)$$

(5)

where the integral is over a geometrical optics path, $c_0$ is a representative sound speed, $<\mu^2(z_r)>$ is the fractional sound speed variance from internal waves, and $L_n(\theta, z_r)$ is an effective internal wave correlation length along a ray path tangent.

Also, following Flatté and Stoughton [1986], in the saturated region the coherence between two points 1 and 2 is given closely by

$$<\psi^*(2)\psi(1)> \approx \exp\left[-\frac{1}{2}D(2,1)\right]$$

(6)

where $D(2,1)$ is the phase structure function which can be computed from

$$D(2,1) = q^2 \int_0^R dx <\mu^2(z_r)> L_n(\theta, z_r)$$

$$\times \left\{1 - \cos(\omega \Delta t) \cos(k_n \Delta z)\right\}$$

(7)

$\Delta t$ is the time separation between receptions, and $\Delta z$ are range-dependent separations between rays to the

two receivers. Internal wave frequency and vertical wavenumber are $\omega$ and $k_n$. Curly braces denote averages over the internal wave spectrum as in Esswein and Flatté [1981]. Let $\psi(0)$ be the observed acoustic field at a given reference depth and time. The coherence of the field is described by second moments, and the ones that can be treated for the ATOC dataset are

$$<\psi^*(\Delta z)\psi(0)> \approx \exp\left[-\frac{1}{2}(\frac{\Delta z}{z_0})^2\right]$$

(8)

with $z_0^{-2} = q^2 \ln(\Phi) \int_0^R dx <\mu^2(z_r)> L_n(\theta, z_r)$

$$\times \{k_n^2\} |\zeta_i(x)|^2$$

(9)

and

$$<\psi^*(\Delta t)\psi(0)> \approx \exp\left[-\frac{1}{2}(\frac{\Delta t}{t_0})^2\right]$$

(10)

with

$$t_0^{-2} = q^2 \int_0^R dx <\mu^2(z_r)> L_n(\theta, z_r) \{\omega^2\}$$

(11)

The quantity $\Phi = \sigma\tau, q = \sigma/c_0, \sigma$ is acoustic frequency, and $\zeta_i(x)$ is a raytube function whose properties are described in Esswein and Flatté [1981].

Therefore the observations of $z_0$ and $t_0$ complement those of $\tau$ in that the weighting function for $z_0$ includes a factor of $\{k_n^2\}$ thereby increasing the sensitivity to small vertical scales, and the weighting function for $t_0$ includes a factor of $\{\omega^2\}$ which increases the sensitivity to high internal-wave frequencies. It must be emphasized, however, that the accuracy of these formulae for the frequencies and ranges of ATOC remains to be established. Clearly Monte Carlo numerical simulations will play a significant role in establishing the limitations of the above formulae.
At this point there is no analytic theory for mode travel time spread, so there is no known weighting function. Numerical simulations must be done to estimate the sensitivity of mode spread to variations in internal-wave spectral parameters. In addition there is the issue of mesoscale effects on modal spread. For SLICE89, the mesoscale field was very weak, and the pulse finale was unaffected by these perturbations. This situation may be different for the ATOC data.

4. Summary

A number of acoustic observables that are available in a pulse-transmission experiment have been used to estimate the average strength of the internal-wave field in the North Pacific during July 1989. The Garrett-Munk internal-wave model with half the reference displacement variance is consistent with all of the acoustic observables that are associated with acoustic energy that samples the ocean below 100 m depth. For acoustic energy that samples the upper 100 m, dramatic differences are seen between the measurements and the acoustic predictions using the GM model. Since resolved ray arrivals for the basin scale transmissions of ATOC fall in this category, (see Figure 8), a model for upper ocean sound-speed fluctuations is clearly needed.

Internal waves cause wavefront travel-time variations and also cause mode travel time spreading which is evident through a diffusion in depth of the latest-arriving acoustic energy in a pulse. Therefore, measurements of these acoustic observables can be used to determine internal-wave displacement variance. Additional acoustic observables like time and depth coherence can provide information on the high-frequency and high vertical wavenumber components of the internal-wave spectrum.

At present, the only method for connecting most of these acoustic observables to internal-wave displacement variance is numerical simulation, and the calculation is so demanding that supercomputers are required. An analytic theory exists for travel-time wander [Flatté et al., 1979], and signal time and depth coherence [Flatté and Stoughlin, 1986], but the validity of these expressions remains to be established for the frequencies and ranges of ATOC. Expanding our analytic techniques to include other acoustic observables like mode travel time spread and the broadband travel time variance, thus avoiding the need for supercomputer calculations is of course highly desirable.

Finally, basin scale acoustic transmissions appear likely to resolve large-scale heating and cooling differences in the upper 250 m of the ocean. Progress in enhancing the vertical resolution of basin-scale acoustic thermometry can be made either by going to lower frequencies or by using improved signal processing techniques. These advances will allow acoustic thermometry to be used even more effectively. At 75 Hz, advanced beamforming techniques show some promise of extending the identification of wavefronts into the highly scattered mode-like region [Worcester, personal communication]. The use of lower frequencies can also mitigate the effects of internal waves since the associated larger Fresnel zone of the acoustic wave more effectively averages out the internal-wave sound-speed fluctuations. Preliminary results from a two-frequency transmission test at 28 Hz and 84 Hz for the ATOC network shows promise for perhaps resolving more wavefronts and achieving less mode travel-time spread (i.e., modes are more adiabatic) [Birdsall, personal communication]. In short several breakthroughs appear forthcoming.

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References


