Ocean acoustic tomography

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Ocean Acoustic Tomography

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Abstract.
Ocean acoustic tomography (OAT) was proposed in 1979 by Walter Munk and Carl Wunsch as an analogue to x-ray computed axial tomography for the oceans. The oceans are opaque to most electromagnetic radiation, but there is a strong acoustic waveguide, and sound can propagate for 10 Mm and more with distinct multiply-refracted ray paths. Transmitting broadband pulses in the ocean leads to a set of impulsive arrivals at the receiver which characterize the impulse response of the sound channel. The peaks observed at the receiver are assumed to represent the arrival of energy traveling along geometric ray paths. These paths can be distinguished by arrival time, and by arrival angle when a vertical array of receivers is available. Changes in ray arrival time can be used to infer changes in ocean structure. Ray travel time measurements have been a mainstay of long-range acoustic measurements, but the strong sensitivity of ray paths to range-dependent sound speed perturbations makes the ray sampling functions uncertain in real cases. In the ray approximation travel times are sensitive to medium changes only along the corresponding eigenrays. Ray theory is an infinite-frequency approximation, and its eikonal equation has nonlinearities not found in the acoustic wave equation. We build on recent seismology results (kernels for body wave arrivals in the earth) to characterize the kernel for converting sound speed change in the ocean to travel time changes using more complete propagation physics. Wave-theoretic finite frequency kernels may show less sensitivity to small-scale sound speed structure.

1. Introduction
Ocean acoustic tomography was introduced by Munk and Wunsch [1, 2] as a remote-sensing technique for large-scale monitoring of the ocean interior using low-frequency sound. In their original conception, the method works by measuring the travel times of sound pulses between source and receiver, which are separable by time of arrival and reception angle when received on a vertical array. The pulse arrivals are matched to calculated ray arrivals from ray-tracing programs or to pulse arrivals from full-wave propagation models. In the ray approximation, the observed arrival time change can be ascribed to sound speed changes along the ray path corresponding to the arrival. Although the intensity of the arrival peaks can vary strongly due to the influence of internal wave variability, the arrival times are stable and many can be tracked over a year-long experiment. Sound speed is related to ocean temperature (and weakly to
salinity), so the ocean temperature field can be estimated from arrival time measurements over multiple paths. When moorings carry both sources and receivers, simultaneous transmissions in opposite directions between two moorings follow similar “reciprocal” paths, so the effects of sound speed perturbations are identical while the effects of ocean currents are opposite. Forming sums and differences of pairs of reciprocal path travel times allows the isolation of the sound speed changes (sum) and currents (differences). Useful arrivals can be identified and tracked at ranges of thousands of kilometers, so the travel time measurements average over large regions, and have been suggested as a method for monitoring large-scale changes in temperature such as might be expected from climate change. Averaging reduces the effects of small-scale ocean variability such as internal waves and even quasi-geostrophic eddies in long-range transmissions.

At long range some ray paths have been shown to be highly sensitive to small-scale range-dependent changes in sound speed, leading to uncertainty in the kernel that describes the sampling of the ocean by the ray travel time [3][4]. It is hoped that using the full-wave kernels developed in seismic studies ([5, 6], [7–9], [10, 11]) and for the ocean sampling problem [12] will reduce the sensitivity seen with the high-frequency ray approximation.

2. Ocean Acoustics

Sound speed in the ocean increases with increasing pressure and with increasing temperature. The sensitivity of sound speed to salinity is generally much smaller than that of either temperature or pressure. The midlatitude ocean is warmest near the surface and the temperature drops off rapidly in the upper 200 to 700 m, forming the ocean “main thermocline” and greatly decreasing the sound speed. Below the main thermocline, the ocean temperature decreases very slowly with depth and the increasing pressure with depth dominates the sound speed, creating a minimum in sound speed at the bottom of the main thermocline. This ocean sound channel is also called the “SOFAR” channel, from the acronym “Sound Fixing and Ranging”. It has been important in many military applications in the 20th century, including anti-submarine warfare (ASW) and triangulation on downed fliers or equipment [2]. The waveguide allows sound to propagate usefully for long ranges in the ocean, leading to a variety of modern applications, including earthquake monitoring, nuclear testing monitoring, and ocean remote sensing. Near the poles, the oceans are cold at the surface, and so the sound speed minimum is often at or near the surface. Figure 1 shows a schematic of ray arrivals in a typical ocean waveguide.

The ray arrivals are separated at the receiver by their time of arrival and by the angle of propagation, in the case that a vertical line array (VLA) of receivers is available. Figure 2 shows the time record of simulated pulse arrivals at 1000 km range from the integration of a parabolic equation (PE) approximation [13] to the acoustic wave equation. The gray scale gives intensity in decibels, so there is a dynamic range of 1000 for power in the figure.

The “waveguide invariant” [14] (lower panel of figure 2) is a fundamental descriptor of the waveguide [15] which describes the dispersive nature of the propagation. It at least partially controls the geometry of the Fresnel zones (and the ray sampling) [16] and the stability of the ray path to perturbations in launch angle or ocean sound speed structure [4]. At long range, rays with large values of this parameter can exhibit strong sensitivities that have been called “ray chaos” [3]. Since ocean acoustic tomography depends on ray paths to sample the ocean, large uncertainties in path geometry would make the sampling kernel unknown at long range.

3. Ocean sampling by rays

In the ray approximation travel times are sensitive to medium changes only along the corresponding eigenrays. Ray travel times are robust in the presence of internal-wave-induced scattering at least partly because of Fermat’s principle, which states that to first order ray travel times are not affected by ray path changes. This discussion will focus on the use of ray travel time data, since they have been observed to be stable and identifiable in long-range tomography.
Figure 1. Schematic illustration of ray propagation in an ocean waveguide.

experiments and have been used in most tomography experiments to date. The ray geometry controls the sampling (kernels) of observations of pulse arrivals.

3.1. Ray travel time
Representing the ocean sound speed field by \( C(\mathbf{r}) \), and the ocean current vector by \( \mathbf{v}(\mathbf{r}) \), where \( \mathbf{r} \) is position, and assuming that \( |\mathbf{v}|/C \) is small, the travel time \( \tau_i \) of ray \( i \) is

\[
\tau_i = \int_{\Gamma_i} \frac{ds}{C(\mathbf{r}) + \mathbf{v}(\mathbf{r}) \cdot \mathbf{r}'}
\]

where \( \Gamma_i \) is the ray path for ray \( i \) and \( \mathbf{r}' \) is the tangent to the ray at position \( \mathbf{r} \). The sign of \( \mathbf{v} \cdot \mathbf{r}' \) depends on the direction of propagation, and the travel times and ray paths in opposite directions differ because of the effects of currents. (Sound travels faster with a current than against a current.) The eigenrays \( \Gamma_i \) are obtained using a numerical integration.

The “inverse” problem is to compute the sound-speed \( C(\mathbf{r}) \) and current \( \mathbf{v}(\mathbf{r}) \) fields given the measured travel times. The statistics (mean and variance) of \( C(\mathbf{r}) \) in the ocean are available from historical sampling. The interesting problem is therefore to compute the perturbations from an assumed reference state, using the measured perturbations from the travel times computed for the reference state.

Travel times are in general a nonlinear function of the sound-speed and current fields, because
Figure 2. Top: received intensity vs time for a vertical line array of receivers at 1000km range. Bottom: “waveguide invariant” parameter, representing the stability of the ray path with respect to changes in ray launch angle or ocean sound speed structure.

the ray path $\Gamma_i$ depends on $C(r)$ and $v(r)$. Linearize by setting

$$C(r) = C(r, -) + \Delta C(r)$$

$$v(r) = v(r, -) + \Delta v(r)$$

where $C(r, -)$, $v(r, -)$ are the known reference states. The argument $(-)$ denotes the dependence of the variables only on the reference state, independent of the measurements. Normally

$$|\Delta C(r)| < < C(r, -)$$

$$|\Delta v(r)| < < C(r, -).$$
In general, however,
\[ |\Delta \mathbf{v}(\mathbf{r})| > |\mathbf{v}(\mathbf{r},-)| \]
(6)
because the fluctuations in current at a fixed location in the ocean are typically large compared to the time-mean current.

Setting \( \mathbf{v}(\mathbf{r},-) \equiv 0, \Delta \mathbf{v}(\mathbf{r}) \equiv \mathbf{v}(\mathbf{r}) \), forming perturbation travel times, and linearizing to first order in \( \Delta C/C \) and \( |\mathbf{v}|/C \) gives
\[
\begin{align*}
\Delta \tau_i^+ &= \tau_i^+ - \tau_i(-) = -\int_{\Gamma_i(-)} \frac{[\Delta C(\mathbf{r}) + \mathbf{v}(\mathbf{r}) \cdot \mathbf{r}'(-)]}{C^2(\mathbf{r},-)} ds \\
\Delta \tau_i^- &= \tau_i^- - \tau_i(-) = -\int_{\Gamma_i(-)} \frac{[\Delta C(\mathbf{r}) - \mathbf{v}(\mathbf{r}) \cdot \mathbf{r}'(-)]}{C^2(\mathbf{r},-)} ds 
\end{align*}
\]
(7)
(8)
where \( \Gamma_i(-), \mathbf{r}'(-) \) are the ray path and tangent vector for the reference state. The superscript plus (minus) refers to propagation in the + (−) direction. The reference travel time is
\[
\tau_i(-) = \int_{\Gamma_i(-)} \frac{ds}{C(\mathbf{r},-)}.
\]
(9)

The sum of the travel time perturbations
\[
\Delta s_i = \frac{1}{2} (\Delta \tau_i^+ + \Delta \tau_i^-) = -\int_{\Gamma_i(-)} \frac{ds}{C^2(\mathbf{r},-)} \Delta C(\mathbf{r})
\]
(10)
depends only on the sound-speed perturbation \( \Delta C(\mathbf{r}) \). The difference
\[
\Delta d_i = \frac{1}{2} (\Delta \tau_i^+ - \Delta \tau_i^-) = -\int_{\Gamma_i(-)} \frac{ds}{C^2(\mathbf{r},-)} \mathbf{v}(\mathbf{r}) \cdot \mathbf{r}'(-)
\]
(11)
depends only on the water velocity \( \mathbf{v} \cdot \mathbf{r}' \) along the ray path. Forming sum and difference travel times separates the effects of \( \Delta C \) and \( \mathbf{v} \). This separation is crucial for measuring \( \mathbf{v} \), because \( |\mathbf{v}| \) is usually much smaller than \( \Delta C \). It is not crucial for measuring \( \Delta C \), however, and one-way, rather than sum, travel time perturbations are often used for this purpose.

The data used in the inverse problem can therefore either be the one-way travel time perturbations, e.g., \( \Delta \tau_i^+ \), or the sum and difference travel time perturbations, \( \Delta s_i \) and \( \Delta d_i \). The use of one-way travel time measurements to estimate \( \Delta C \) is sometimes given the special name of acoustic thermometry, reflecting the fact that sound-speed perturbations depend mostly on temperature.

4. Full wave kernel
As an alternative to the singular (and sensitive) ray approximation, the kernel for travel times can be determined from more physical approximations. Many examples exist in the literature for kernels derived from a variety of propagation codes, including finite differencing, spectral elements [10], normal modes, and the Parabolic Equation (PE) approximation used commonly in ocean acoustics [13]. These are examples of adjoint methods, which have been widely used with ocean and atmospheric general circulation models [17–19].

The adjoint can be derived from conversion of the propagation model code [20], or more typically by using the Born approximation. Two types of observables are considered here, the travel-time and intensity sensitivity of an arrival peak to changes in sound-speed. The sensitivity kernel for acoustic intensity is the same as the raytube calculated by Bowlin ([21]).
The Green’s function $G(r|\mathbf{r}_s, \omega)$ is the response at an arbitrary point $r$ of the wave equation,

$$
\left[ \nabla^2 + \frac{\omega^2}{c^2(r)} \right] G(r|\mathbf{r}_s, \omega) = -\delta(\mathbf{r} - \mathbf{r}_s) \, ,
$$

(12)
to a harmonic point source at $\mathbf{r}_s$ with frequency $\omega$, in a medium with variable sound speed $c(r)$. The pressure waveform can be synthesized as the real part of

$$
p(t) = \frac{1}{2\pi} \int_{-\infty}^{\infty} G(r|\mathbf{r}_s, \omega) P_s(\omega) e^{j\omega t} d\omega \, ,
$$

(13)
where $P_s(\omega)$ is the frequency-dependent sound source excitation. Using the first-order Born approximation, the change in the Green’s function due to a sound-speed perturbation is

$$
\Delta G(r|\mathbf{r}_s, \omega, \Delta c) = -2\omega^2 \int_V G(r|r', \omega) G(r'|\mathbf{r}_s, \omega) \frac{\Delta c(r')}{c^2(r')} dV(r')
$$

(14)
and the resultant pressure perturbation is

$$
\Delta p(t, \Delta c) = \frac{1}{2\pi} \int_{-\infty}^{\infty} \Delta G(r|\mathbf{r}_s, \omega, \Delta c) P_s(\omega) e^{j\omega t} d\omega \, .
$$

(15)

Given the $dp/dc$ kernel (dependence of pressure at a particular frequency on sound speed), the sensitivity of any observable can be found by using the chain rule. Likewise, we can ”propagate our sensitivity” through whatever processing method we choose, including beamforming, by applying the processing in question to $dp/dc$ just as it would be applied to the $p$.

As explained in Skarsoulis and Cornuelle[12], the travel-time perturbation of the $i$-th demodulated arrival peak can then be written as

$$
\Delta \tau_i = \int_V K_i^T(r'|\mathbf{r}_s, \mathbf{r}_r) \Delta c(r') dV(r')
$$

(16)
where the travel-time sensitivity kernel or TSK, $K_i$, is

$$
K_i^T(r'|\mathbf{r}_s, \mathbf{r}_r) = \Re \left\{ \frac{u_i - jv_i}{2\pi b_i} \int_{-\infty}^{\infty} j\omega Q(r|\mathbf{r}_s, \mathbf{r}_r, \omega) e^{j\omega \tau_i} d\omega + \frac{\dot{u}_i - j\dot{v}_i}{2\pi b_i} \int_{-\infty}^{\infty} Q(r|\mathbf{r}_s, \mathbf{r}_r, \omega) e^{j\omega \tau_i} d\omega \right\}
$$

(17)
and

$$
Q(r|\mathbf{r}_s, \mathbf{r}_r, \omega) = G(r|\mathbf{r}_s, \omega) G(r|\mathbf{r}_r, \omega) \frac{2\omega^2 P_s(\omega)}{c^4(r)} \, .
$$

(18)
The normalization

$$
b_i = \dot{u}_i^2 + u_i\ddot{u}_i + \dot{v}_i^2 + v_i\ddot{v}_i
$$

(19)
is defined in terms of the complex pressure $p_i = u_i + jv_i$ measured at the peak of the unperturbed arrival.

The first-order intensity perturbation in response to a pressure perturbation is

$$
I + \Delta I = (u + \Delta u)^2 + (v + \Delta v)^2
$$

$$
\Delta I \approx 2(u - jv)(\Delta u + j\Delta v) = 2(u - jv)\Delta p \, .
$$

(20)
Using the chain rule to relate this to a sound-speed perturbation

\[ \Delta I(\tau, \Delta c) = \int_V K^I_i(r'|r_s, r_r) \Delta c(r') dV(r') \]  

(21)

where the intensity sensitivity kernel or ISK, \( K^I_i \), is

\[ K^I_i(r|r_s, r_r) = \Re \left\{ 2(u_i - jv_i) \int_{-\infty}^{\infty} Q(r|r_s, r_r, \omega)e^{j\omega \tau_i} d\omega \right\} . \]  

(22)

An actual observation of intensity would occur at the perturbed arrival peak, so we must use a Taylor expansion

\[ I(\tau_i + \Delta \tau, \Delta c) = I(\tau_i, 0) + \Delta I(\tau_i, \Delta c) + \dot{I}(\tau_i, 0) \Delta \tau_i \]  

(23)

to find the intensity. Since the time derivative at the unperturbed peak is zero, the measured intensity at the new perturbed arrival time

\[ \Delta I(\tau_i + \Delta \tau, \Delta c) = \Delta I(\tau_i, \Delta c) \]  

(24)

is the same as at the unperturbed travel time.

5. TSK examples

The arrival shown in the upper panel of figure 2 were generated by for a source with center frequency 75 Hz and a bandwidth of 18.75 Hz (Q=4). The definition of Q is the ratio of the carrier frequency to the bandwidth, but it is also the number of carrier cycles in a pulse. The source is at a depth of about 1200m and the sound-speed profile (not shown) is an analytical profile used by [15] to examine ray stability. The horizontal line at 1200m indicates the depth of the receiver, and the dots indicate the arrivals selected for sensitivity calculations. The dots labeled “S” and “U” indicate “stable” and “unstable” arrivals as measured by the waveguide invariant parameter (lower panel). Large absolute values of \( \alpha \) indicate low path stability.

Figure 3 shows the TSK of the arrivals in color the eigenray that links source and receiver at the same travel time in black. The TSK structure is as seen in previous work [12]. There is a blue central section indicating decreased travel-time with faster sound-speeds and there is a Fresnel zone structure around the central ray, where the number of red-blue variations (4 in this case) matches the number of cycles in the acoustic pulse. The kernel shown has been integrated in the direction perpendicular to the vertical plane between source and receiver so that there is no "doughnut hole" along the eigenray path itself as seen in 3-dimensional kernels. Figure 3 shows the TSK for the stable (top panel) and unstable (bottom panel) rays, respectively. The TSK of the unstable ray is clearly smeared in comparison the stable arrival. This is interpreted as a full-wave representation of the uncertainty of the sampling by the unstable ray in the presence of inhomogeneities.

6. Examples

Tomographic methods have been used in a number of studies covering a range of ocean processes and all latitudes. Measurements have been made at scales ranging from a few tens of kilometers (e.g., to measure the transport through the Strait of Gibraltar) to thousands of kilometers (e.g., to measure the heat content in the northeast Pacific Ocean). This review concludes by summarizing a few examples to illustrate the use of tomographic observations of the ocean.
6.1. Barotropic and Baroclinic Tides

Sum and difference travel times from long-range reciprocal transmissions provide precise measurements of the sound-speed (temperature) changes associated with baroclinic (internal) tidal displacements and of barotropic tidal currents, respectively.

The availability of global sea-surface elevation data from satellite altimeter measurements has made possible the development of improved global tidal models. Tomographic measurements of tidal currents made in both the central North Pacific and western North Atlantic Oceans have shown that the harmonic constants for current derived from a recent global tidal model (TPXO.2) are accurate to a fraction of a millimeter per second in amplitude and a few degrees in phase in open ocean regions. The integrating nature of the tomographic measurements suppresses short-scale internal waves and internal tides, providing tidal current measurements that are substantially more accurate than those derived from current meter data.

Tomographic measurements of sound-speed fluctuations at tidal frequencies from the same experiments revealed large-scale internal tides that are phase-locked to the barotropic tides. Prior to these measurements it had been commonly assumed that midocean internal tides are not phase-locked to the barotropic tides (except for locally-forced internal tides) and have correlation length scales of only order 100–200 km. These observations were subsequently confirmed from satellite altimeter data. The acoustical observations of the baroclinic tide average in range and depth, suppressing internal wave noise and providing enhanced estimates of the deterministic part of the internal tide signal compared to point measurements.
Figure 4. ATOC and NPAL sources and receivers and ray paths superposed on a plot of North Pacific bathymetry.

6.2. Heat Content
Acoustic methods have been used to measure the heat content of the ocean and its variability on basin scales, taking advantage of the integrating nature of acoustic transmissions to rapidly and repeatedly make range- and depth-averaged temperature measurements at ranges out to about 5000 km.

Measurements of basin-scale heat content in the Northeast Pacific were made intermittently from 1983 through 1989 using transmissions from an acoustic source located near Kaneohe, Hawaii, and more recently from 1996 through 2005 during the Acoustic Thermometry of Ocean Climate (ATOC) and North Pacific Acoustic Laboratory (NPAL) projects using sources located off central California and north of Kauai, Hawaii (see figure 4). Data from these projects have shown that ray travel times may be used for acoustic thermometry at least out to ranges of about 5000 km. The estimated uncertainty in range- and depth-averaged temperature estimates made from the acoustic data at these ranges is only about 10 m°C.

References