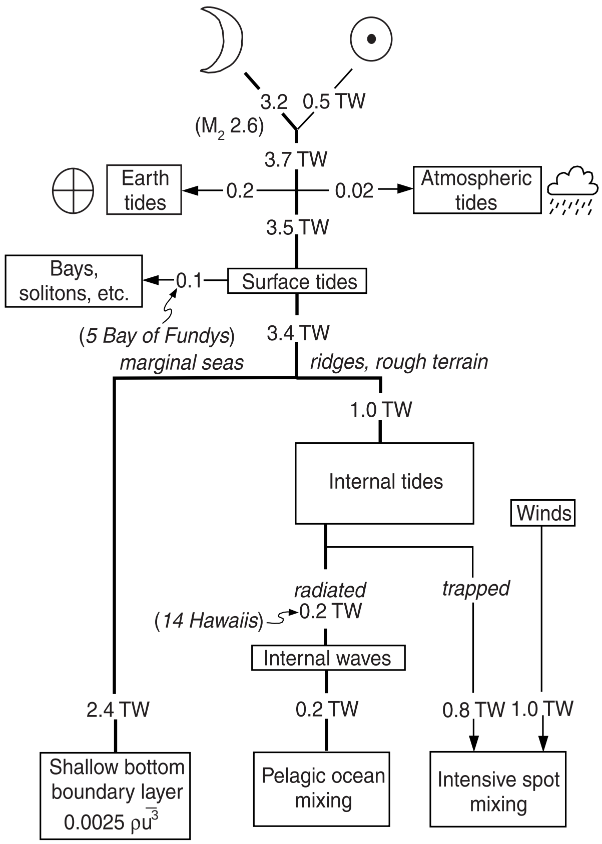

Figure 2 Sketch of proposed flux of tidal energy (modified from Munk and Wunsch, 1997). The traditional sink is in the turbulent boundary layer of marginal seas. Scattering into internal tides over ocean ridges (by the equivalent of 14 Hawaii's) and subsequent degradation into the internal wave continuum feeds the pelagic turbulence at a level consistent with $\kappa_{\text{pelagic}} = 10^{-5}\,\text{m}^2\,\text{s}^{-1}$. Most of the ocean mixing is associated with a few concentrated areas of surface to internal mode convergence over regions of extreme bottom roughness and with severe wind events. Light lines represent speculation with no observational support.

See also

Dispersion and Diffusion in the Deep Ocean. Internal Tides. Internal Waves. Tides. Turbulence in the Benthic Boundary Layer.

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INTERNAL TIDES

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Introduction

Oceanic internal tides are internal waves with tidal periodicities. They are ubiquitous throughout the ocean, although generally more pronounced near large bathymetric features such as mid-ocean ridges and continental slopes. The internal vertical displacements associated with these waves can be extraordinarily large. Near some shelf breaks where the surface tides are strong, internal displacements (e.g., of an isothermal surface) can exceed 200 m. Displacements of 10 m in the open ocean are not uncommon. The associated current velocities are usually comparable to or larger than the currents of the surface tide. Internal tides can occasionally generate packets of internal solitons which are detectable in remote sensing imagery. Other common nonlinear features are generation of higher harmonics (e.g., 6 h waves) and wave breaking. Internal tides are known to be an important energy source for mixing of shelf waters. Recent research suggests that they may also be a significant energy source for deep-ocean mixing.

Internal tides were first recognized in the early part of the twentieth century, yet as late as the 1950s arguments were still being waged over what causes them. Their wavelengths, generally shorter than 200 km, are poorly matched to the planetary-scale astronomical tidal potential, so the generation mechanism for surface tides appears inapplicable. Various theories invoking hypothetical resonances at inertial latitudes (where tidal and Coriolis frequencies are equal) were put forward, but they are not compelling, not least because the inertial latitude for the dominant tide M_2 is in the far polar latitudes (74.5°). The now accepted explanation for internal tides is that they are generated by the interaction of the barotropic surface tide with bottom topography. As the tide sweeps stratified water over topographic features, it disrupts normal (equilibrium) isopycnal layers, setting up pressure gradients that induce secondary internal motions at the same frequency as the tide.

Since internal tides are a special kind of internal wave, much of our knowledge of internal waves is immediately applicable. For example, an internal tide always displays current shear – i.e., the associated horizontal current velocities change with depth – whereas the surface tide's horizontal current is independent of depth. And like other internal waves in smoothly varying density stratification, an internal tide displays the seemingly odd property that its group velocity is in the same vertical plane but perpendicular to its phase velocity. The fundamental properties of internal tides, including whether or not they even exist, are controlled by the relative magnitudes of three basic frequencies: the tidal frequency ω, the local Brunt-Väisälä or buoyancy frequency N, and the local Coriolis frequency f. Depending on which of these frequencies is highest and which lowest, internal tides may propagate freely away from their generation point, they may be reflected in some manner, or they may be evanescent. For mid-latitude semidiurnal tides, typically $f < \omega < N$, a regime allowing free propagation.

Given that the generation and propagation of internal tides depend strongly on the stratification, it is not surprising that most observations have found internal tides to be highly variable, sometimes with pronounced seasonal variations. In some places they appear only during spring tides (when solar and lunar tides are at maximum). In other places they appear randomly intermittent, evident for several days and then disappearing. Some observations, primarily from the open ocean, have revealed a component that does remain temporally coherent with the astronomical potential (see below), but the dominant characteristic of internal tides in most regions is one of incoherence, both spatially and temporally.

Modes and Beams

Two complementary dynamical frameworks are used for analyzing internal tides: decomposition into vertical modes and propagation along characteristics. Generally, the latter description is more useful near generation points, and the modal description more useful elsewhere, but in any particular situation one or the other approach may be advantageous.

Both approaches require knowledge of the stratification, usually parameterized by the buoyancy frequency N. This is the frequency with which a vertically displaced fluid element would oscillate because of restoring buoyancy forces. It is given by $N = \sqrt{(-g\rho^{-1}\partial\rho/\partial z)}$, where g is the acceleration of gravity and ρ is the average potential density, a function of position and depth. The Coriolis frequency $f = 2\Omega \sin\phi$, where ϕ is the latitude and Ω is the Earth's sidereal rotation frequency $(7.2921 \times 10^{-5}\mathrm{s}^{-1})$.

Modes

The governing dynamical equations for internal tides, under linear, hydrostatic, inviscid, Boussinesq and flat-bottom assumptions, and neglecting the horizonal currents, may be solved by separation of variables. The equation for the vertical displacement leads to an eigenvector problem with eigenvalue α_n^2:

$$\frac{\partial^2}{\partial z^2} G_n(z) + \alpha_n^2 N^2(z) G_n(z) = 0$$

The eigenvectors $G_n(z)$, ordered so that $\alpha_{n+1} > \alpha_n$, provide a complete, orthogonal basis for the internal vertical displacements. The corresponding equations for horizontal dependence yield expressions for the horizontal wave number k, phase velocity $c_p = \omega/k$, and group velocity $c_g = \mathrm{d}\omega/\mathrm{d}k$ in terms of α_n:

$$\begin{aligned} k^2 &= \alpha_n^2(\omega^2 - f^2) \\ c_p^2 &= \omega^2/[\alpha_n^2(\omega^2 - f^2)] \\ c_g &= (c_p \alpha_n^2)^{-1} \end{aligned}$$

If N is taken constant, then G_n can be found analytically: $G_n(z) = a \sin(n\pi z/D)$ for an ocean depth

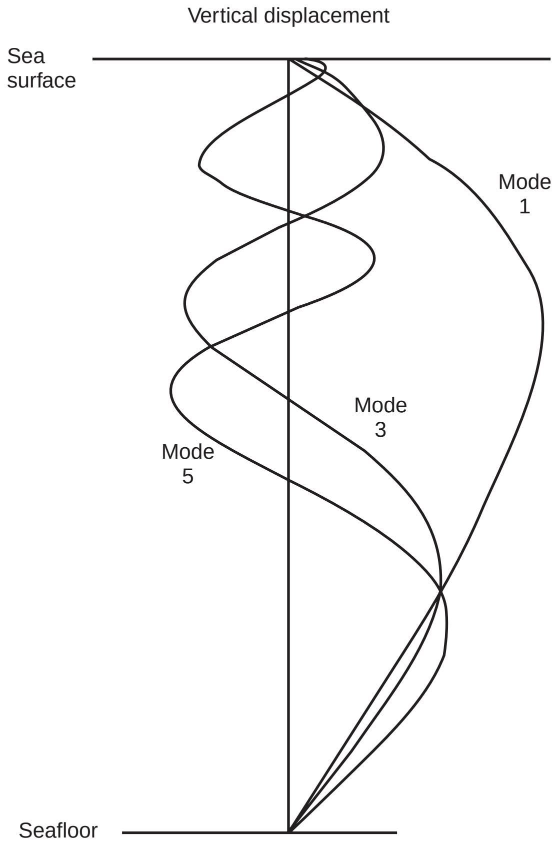

Figure 1 Baroclinic displacement modes 1, 3, and 5, computed for a buoyancy frequency profile from the deep ocean.

D. If N is taken more representative of deep-ocean conditions, with a peak at the pycnocline, then the oscillations in G_n are shifted upward (**Figure 1**). Notice that the displacements are small or zero at the top and bottom of the water column, and that each G_n has $n-1$ crossings of the origin. Horizontal velocity modes are given by $\partial G_n/\partial z$, and they have n crossings and hence nonzero shear for all modes. Most observations of internal tides (except those very near the generation point) are adequately described by a superposition of a few low order modes.

In the deep ocean typical phase speeds c_p are of order $3\,\mathrm{m\,s^{-1}}$ for $n=1$. Corresponding wavelengths $\lambda = 2\pi/k$ are between 100 and 200 km. Higher order modes have speeds and wavelengths given roughly by c_p/n and λ/n, respectively. On continental shelves both speeds and wavelengths may be an order of magnitude smaller. These values are for semidiurnal tides; wavelengths of diurnal tides are approximately twice as large.

From the above expressions for k and c_p it is apparent that internal tides cannot freely propagate unless $\omega > f$. They are 'evanescent' (exponentially damped)' polewards of the critical latitudes where $\omega = f$. Freely propagating waves for diurnal tides are therefore confined to the region between latitudes $\pm 30°$. (In fact, unambiguous observations of diurnal tides are fairly rare, but this is partly due to relatively weak barotropic forcing and higher background noise levels.)

Beams

A complementary approach to modal analyses stems from the equation for the two-dimensional stream function, which is hyperbolic in spatial coordinates and may therefore be solved by the method of characteristics. The resulting solution consists of narrow beams of intense motion embedded in an otherwise resting ocean. The group velocity, and hence the energy propagation, follow the characteristics, which are along lines of slope:

$$c = \tan\theta = \pm\left(\frac{\omega^2 - f^2}{N^2 - \omega^2}\right)^{1/2}$$

From a given internal tide generation point, energy thus propagates along beams at the angle θ relative to horizontal, the angle depending (for a given f and N) only on the tidal frequency. Well defined beams comprise a large number of modes, with modal cancellations occurring outside the allowed beam. A numerical example of beam-like propagation from a shelf break is shown in **Figure 2**.

Generation of internal tides is apparently especially efficient when the seafloor slopes at precisely the critical value c. Barotropic flow is then coincident with the motion plane for free internal waves, resulting in near-resonant conditions in which even quite small surface tides can generate internal tides. With nominal values of $N \sim 50$ cpd, $\omega \sim 2$ cpd, $f \sim 0.6$ cpd, then θ is $2°$. Continental slopes commonly exceed this, so c would be attained near the shelf break, as depicted in **Figure 2.**

When an internal wave is reflected from the ocean bottom or ocean surface, energy propagation is still confined to the angle θ, which makes for a curious variation on the usual laws of reflection. If the wave is incident upon bathymetry that is steeper than θ (supercritical case), energy is reflected backwards into deeper water. If the bathymetry is less steep (subcritical case), energy is reflected forward toward shallower water (see **Figure 3**). Ocean observations of this behavior are not easy to obtain, since mooring instruments must be precisely placed (depending on the ambient N); yet measurements of internal tides in the Bay of Biscay have not only observed the downward energy propagation from the generation

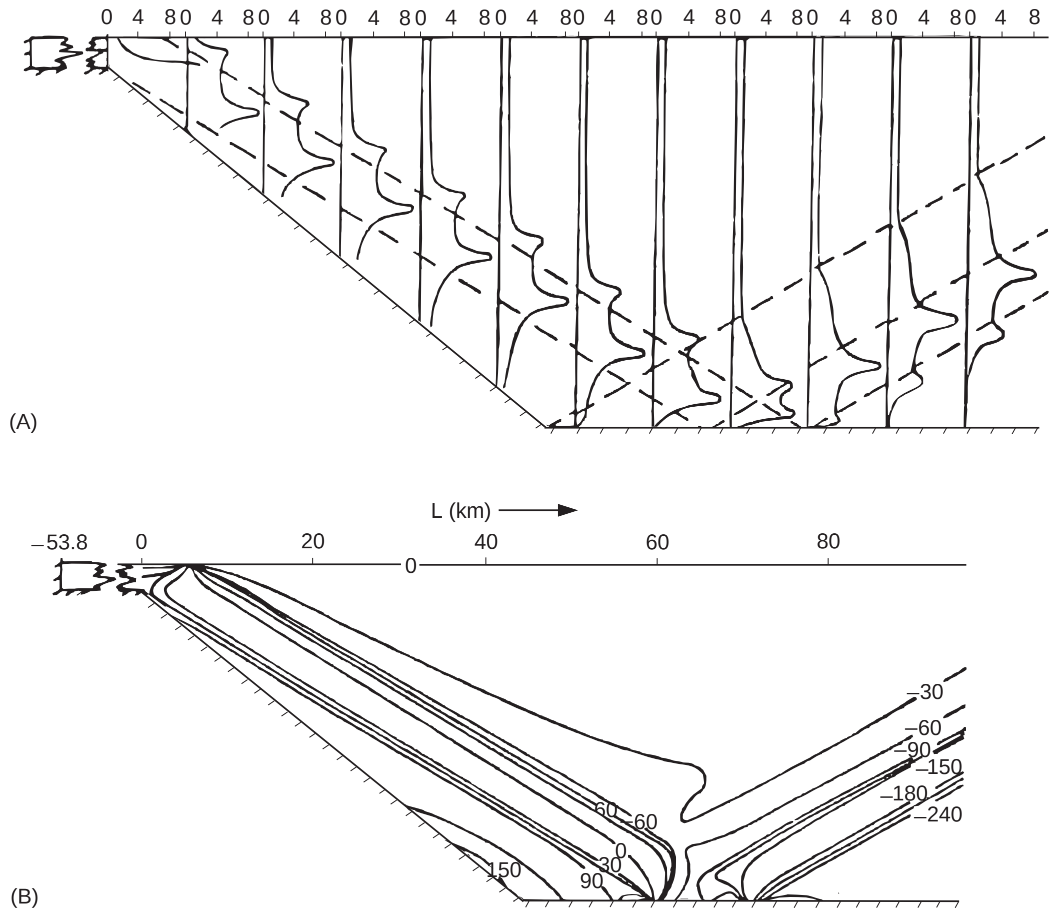

Figure 2 Theoretical internal wave beam from a shelf edge in a constant N ocean. (A) vertical displacements following beam at constant slope tan θ; (B) phase contours (in degrees) of the vertical displacements relative to the surface tide. Notice how phase propagation is at right angles to the beam; i.e., the phase velocity is perpendicular to group velocity (and to the direction of energy propagation). (Reproduced with permission from Prinsenberg SJ and Rattray M (1975) Effects of continental slope and variable Brunt-Väisälä frequency on the coastal generation of internal tides. *Deep Sea Research* 22: 251–263.)

point, but also the subsequent reflection from the ocean bottom. In the Bay of Biscay, as in most places, N diminishes with depth, so θ grows larger and the beams become steeper as they approach the bottom (**Figures 2** and **3** are drawn for constant N.)

With such reflection properties, internal waves incident on a subcritical sloping bottom will be focused into the shallows (as in **Figure 3A**), with energy density correspondingly intensified. The same mechanism tends to trap internal wave energy within steep (supercritical) canyons, where the canyon sides reflect energy ever deeper, focusing it toward the canyon floor. If the floor is subcritical, then energy is further focused toward the canyon head. Intense internal tide currents and large kinetic energy densities have indeed been observed in canyons, and especially near canyon heads.

In the presence of internal viscosity or other dissipative mechanisms, internal tidal beams widen. Because group velocities are smaller and decay scales shorter for higher order modes, beams tend to disintegrate rapidly into the few low order modes that are most commonly observed.

Observations

Internal tides have been observed with a great multitude of instruments and technique, both *in situ* and remote. Four distinctly different examples are given here which serve to highlight a number of characteristic features of internal tides. Except for the first example, emphasis is given to deep-sea tides.

Vertical Profilers

Vertical profilers, ranging from echo sounders to repeated hydrographic casts to special yo-yo instruments, provide some of the clearest pictures of internal tides. An especially dramatic example from the

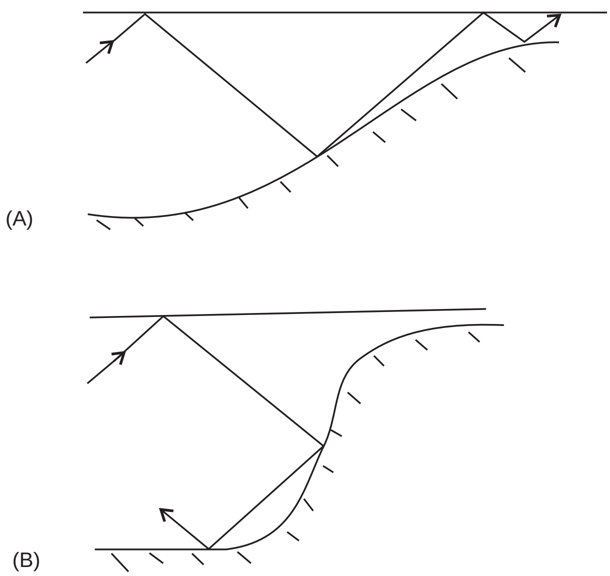

Figure 3 Successive reflections of an internal wave along (A) a subcritical seafloor and (B) a supercritical seafloor. Arrows denote direction of energy propagation.

continental shelf off Oregon is shown in **Figure 4**. It shows a clear semidiurnal signal in the isotherm displacements, somewhat distorted into a bore-like shape (akin to the nonlinear distortion seen in shoaling surface tides in very shallow water). Along its downward edge the tide has generated packets of internal solitons. These solitary waves, starting from an initial 7 m depth, have extraordinarily nonlinear isotherm displacements of 25 m. Although usually less dramatic, this phenomenon is not uncommon in high tide regions on a continental shelf. The internal tide, generated at the shelf break, propagates shorewards and becomes progressively more nonlinear until the bore disintegrates into a group of solitons, the leading one usually of largest amplitude.

Moored Current Meters

Because of their widespread deployments, current meters provide perhaps the most common method for observing, or at least detecting, internal tides, especially in the open ocean. Sufficient vertical sampling is required for decoupling the internal modes from the surface tide (and unfortunately sufficient sampling is not common). **Figure 5** is an example of marginally adequate vertical sampling; it shows tidal current estimates extracted from moored meters near 110°W on the Pacific equator. Estimates are given for each of 10 months, at 10 depths throughout the water column. The current ellipses are fairly uniform below 1000 m; these depths are dominated by the stable, depth-independent currents of the surface tide. In contrast, large temporal variation, and occasionally much larger amplitudes, are evident in the shallower estimates; in these depths, where the buoyancy frequency (and its change) is maximum, the tidal signal is dominated by the internal tide. Modal analysis reveals that the internal tide is essentially random, isotropic, and without a dominant mode for these 10 months.

Such observations are characteristic of *in situ* observations of internal tides; but in a few locations in the deep ocean, a component of the internal tide has been observed that is not so variable and that maintains phase lock with the astronomical tide. The famous MODE experiment in the western Atlantic found that approximately 50% of the internal tide variance was temporally coherent with the astronomical tide. Such observations imply a nearly constant ocean stratification, at least to the extent that it determines generation and propagation properties.

Satellite Altimetry

Recently satellite altimetry has been shown capable of providing a near-global view of the coherent component of internal tides. It does this by detecting the very small surface displacements associated with internal tides. These are given roughly by the tide's internal displacements scaled by $\Delta\rho/\rho$, the fractional difference in water density, typically of order 0.2%, thus implying surface displacements of a few centimeters for internal displacement of tens of meters. Altimetry detects such small waves as modulations (with wavelengths 100–200 km for internal mode 1) of the surface tide as estimated along satellite tracks. Because tides can be estimated from altimeter data only by gathering multi-year time series of elevations at a particular site, only the coherent component of the internal tide which maintains phase lock with the surface tide is capable of being detected.

Figure 6 gives an example of the first detection of such waves, near the Hawaiian Ridge. The waves are roughly 5 cm amplitude near the ridge and decay slowly with distance, but are still detectable 1000 km away. Phase estimates (not shown) reveal clearly that the waves are propagating away from the ridge. Evidently they are created by the barotropic tide striking the ridge (at nearly right angle from the north) and generating an internal tide that propagates both northwards and southwards. The picture reveals three important aspects of deep-ocean internal tides: (1) that in some locations they maintain temporal coherence over several years, thus allowing altimetry to measure them, (2) that they maintain spatial coherence over a wide area, and (3) that they are capable of propagating hundreds to thousands of kilometers before being dissipated. All three aspects contrast sharply with the usual picture of incoherence obtained from *in situ* observations.

Waves similar to those in **Figure 6** have been detected in many regions throughout the global

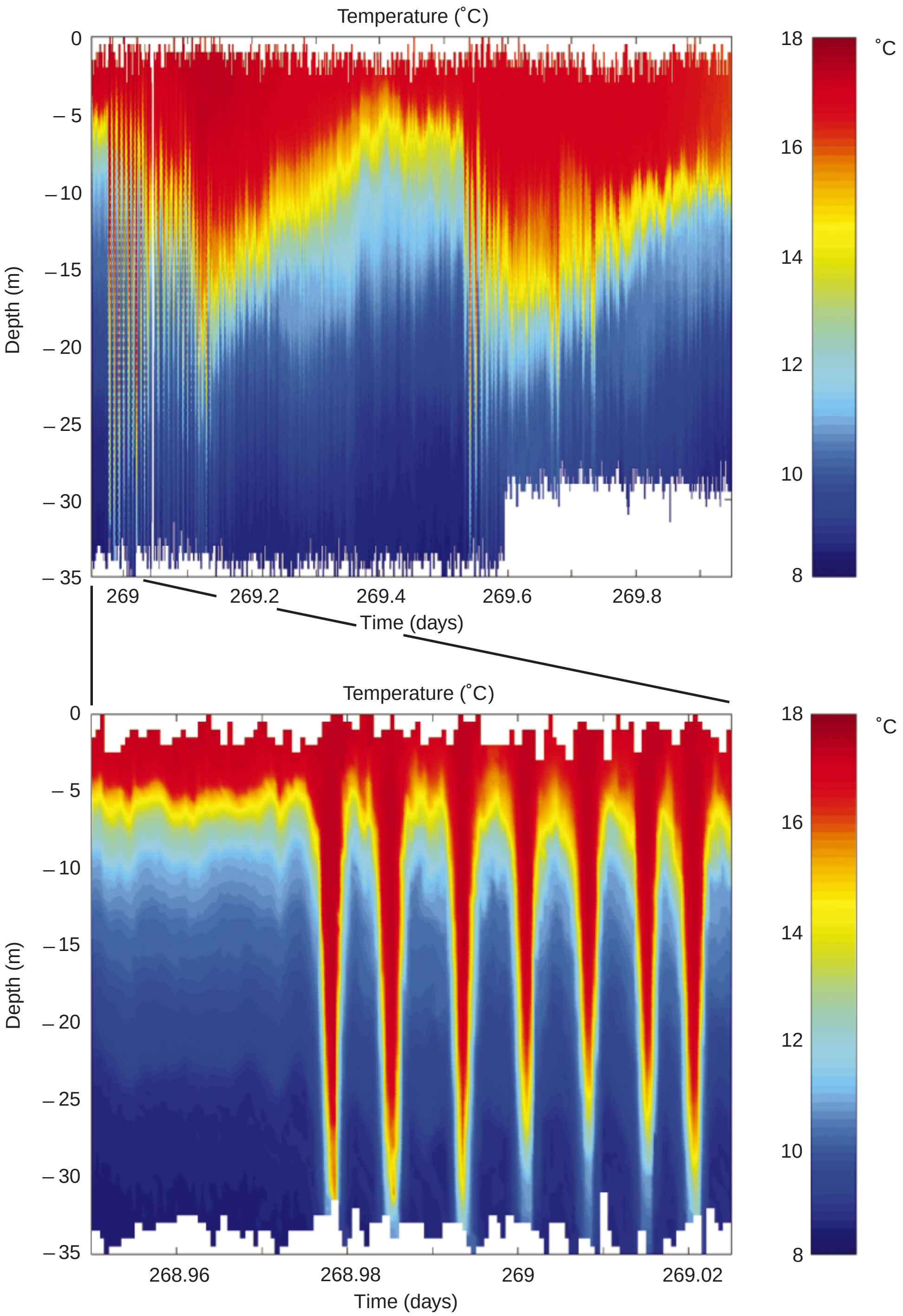

Figure 4 Color contour time series of temperature profiles from the surface to 35 m depth, obtained by repeated 80 s raising and free-falling of the loose-tethered microstructure profiler, deployed offshore Tillamook, Oregon in October 1995. Top: the semidiurnal internal tide displacement (most clearly seen along the yellow 13.8°C isotherm) for a 24 h period. Bottom: a zoom view of a 1.7 h period showing the start of the first soliton displacements. The solitons are separated by roughly 10 min. (Reproduced with permission from Stanton TP and Ostrovsky LA (1998) Observations of highly nonlinear internal solitons over the continental shelf. *Geophysical Research Letters* 25: 1695–1698.)

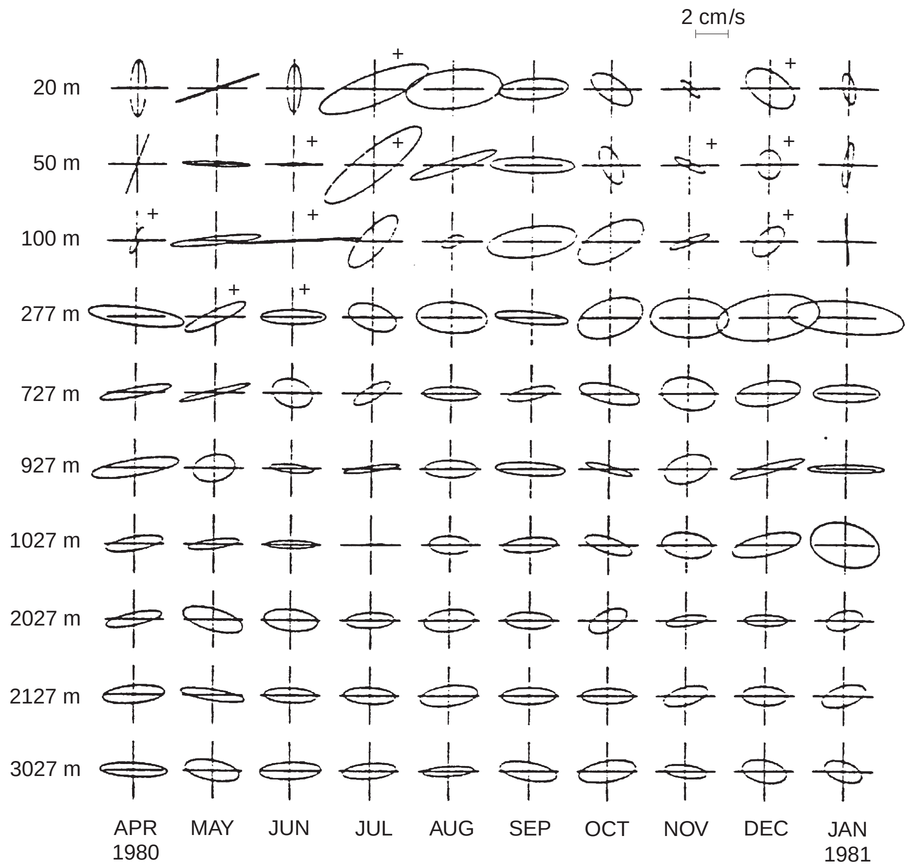

Figure 5 M_2 tidal current ellipses by month at each of 10 depths obtained from moored current meters near 0°, 110°W. Scale bar for velocity is at upper right. Each ellipse indicates how the direction and magnitude of the horizontal current velocity evolves over one tidal cycle. All ellipses are polarized clockwise except those marked with a plus sign. (Reproduced with permission from Weisberg RH, Halpern D, Tang T, Hwang SM (1987) M2 tidal currents in the eastern equatorial Pacific Ocean. *Journal of Geophysical Research* 92: 3821–3826.)

ocean. However, altimetry is incapable of detecting internal tides in a region where they are temporally incoherent. Such is apparently the case, for example, off the northwest European shelf, a region known for some of the largest internal tides in the world, but where the coherent signals in altimeter data are extremely weak. Internal tide studies with satellite altimetry are relatively new, and further work should reveal new facets from a global perspective.

Acoustic Methods

A second example of a powerful, but unconventional, technique for studying coherent internal tides is acoustic tomography. Differences in two-way acoustic travel times between reciprocal transceivers are sensitive to barotropic tidal currents within the acoustic path. Similarly, since vertical isotherm displacements perturb the sound speed within the path, the sums (or averages) of the travel times are sensitive to the internal tide. From a sufficiently long time series the mean tidal characteristics along a given path can be determined. The seemingly coarse spatial resolution is actually an advantage, because it suppresses short-scale internal waves and other noise that typically plague current meter measurements. And, in fact, an array of acoustic transceivers can act as a very sensitive directional antenna for spatially coherent internal tides. Such an array in the central Pacific, consisting of acoustic paths roughly 1000 km long and located just north of the area shown in **Figure 6**, has measured the same coherent internal tide field seen in the altimetry and indicates that the primary source is the Hawaiian Ridge, even at that great distance.

Implications for Energetics and Mixing

Internal tides are an important energy source for vertical mixing, especially in coastal waters where

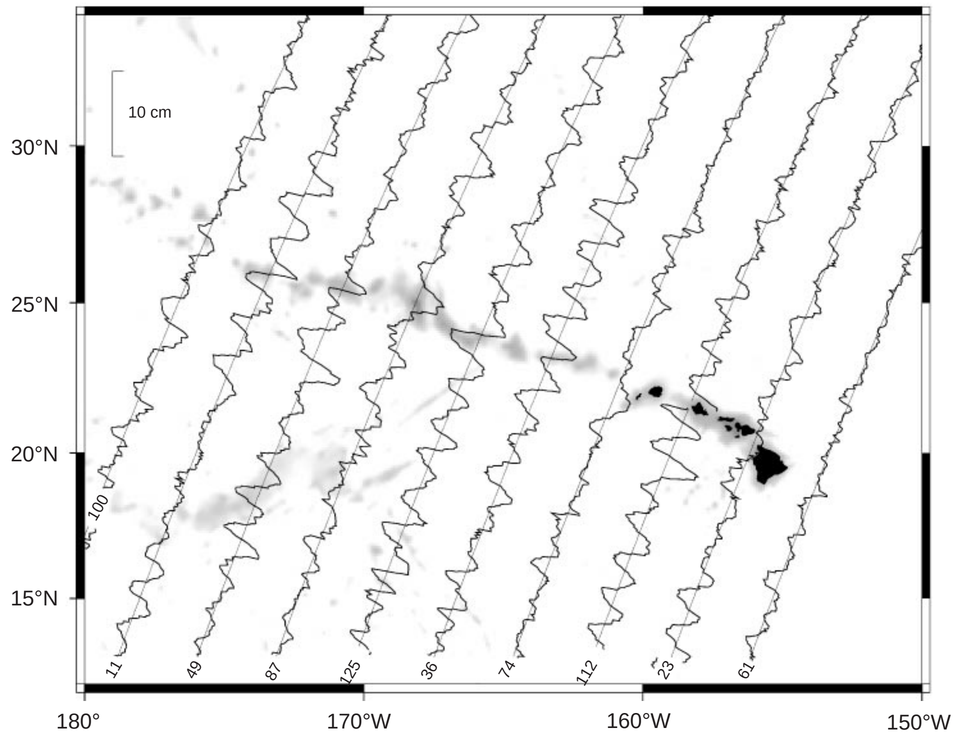

Figure 6 Mean elevations at the sea surface of internal tides near Hawaii, deduced from altimeter measurements of the Topex/Poseidon satellite. Positive values (north of the trackline) indicate that the internal tide's surface elevation is in phase with the barotropic tide's elevation. Scale bar for elevations at upper left. Background shading corresponds to bathymetry, with darker shading denoting shallower depths and the main axis of the Hawaiian Ridge. Only internal tides that are coherent with the surface tide over the entire measurement period (here 3.5 years) can be detected in this manner.

they help maintain nutrient fluxes from deep water to euphotic zones on the shelf. A good example is the Scotian shelf off Nova Scotia where internal tides are responsible for a strip of enhanced concentrations of nutrients and biomass along the shelf break. During each tidal cycle one or two strong (50 m) internal solitons (compare **Figure 4**) are generated near the shelf edge, moving shoreward but dissipating rapidly, possibly within 10 km. Estimated energy fluxes of $500\,\mathrm{W\,m^{-1}}$ appear more than adequate to maintain observed nutrient supply to the mixed layer. Similar mixing mechanisms have been observed in the Celtic Sea and elsewhere.

In the open ocean it seems reasonable that internal tides dissipate by transferring energy into the internal wave continuum or by directly generating pelagic turbulence, but the associated energy fluxes, and even the dominant mechanisms, are unclear. Nonlinearity is a common feature of internal tides (e.g., occurrences of higher harmonics), so 'diffusion' into the continuum is conceivable via nonlinear (resonant triad) interactions, but the evidence for this is so far more anecdotal than convincing. Bottom scattering of low mode tides into higher modes may play a role, as well as wave reflections off sloping bottoms, which tend to intensify kinetic energy densities and may lead to shear instabilities and wave breaking. The traditional view is that both the internal wave continuum and the pelagic turbulence and mixing are maintained by nontidal mechanisms such as wind generation; whether internal tides play a major or minor role in this is not resolved. At a minimum, improved quantitative estimates are needed for the global internal tide energy budget.

The internal tide energy budget also has a bearing on a longstanding geophysical problem: finding the energy sink for the global surface tide. If the generation/dissipation rate for internal tides is sufficiently large, then internal tide generation conceivably supplements the traditional sink of botton friction in shallow seas. Dissipation rates for the surface tide are well determined by space geodesy (e.g., lunar laser ranging) at 3.7 TW, with 2.5 TW for the principal tide M_2. How much of this is accounted for by conversion into internal tides is not well determined; published estimates range from < 100 GW (0.1 TW) to > 1 TW. There is fairly wide agreement that generation of internal tides at continental slopes provides a fairly small energy sink. Both models and measurements suggest that typical energy fluxes at shelf breaks are of order $100\,\mathrm{W\,m^{-1}}$, leading to a global total of order 15 GW. This is perhaps an underestimate, because it may not fully account for shelf canyons and other three-dimensional features, but the order of magnitude seems reliable. Internal tide generation by deep-ocean topography, however, may be far more important. Recent research based on global tide models as well as on empirical estimates of tidal dissipation deduced from satellite altimetry suggests that generation of internal tides by deep-sea ridges and seamounts could account for 1 TW of tidal power. Refining such estimates, and understanding the role that internal tides play in generation of the background internal wave continuum, in vertical mixing, and in maintenance of the abyssal stratification, are some of the outstanding issues of current research.

See also

Acoustics in Marine Sediments. Internal Tidal Mixing. Internal Waves. Satellite Altimetry. Tides.

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INTERNAL WAVES

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Introduction

Waves at the sea surface are a matter of common experience. Surface tension is the dominant restoring force for waves with a wavelength less than 17 mm or so; longer waves are more affected by gravity. They have periods up to about 20 s and amplitudes that may be many meters.

Given the stable density stratification of the ocean, it is not surprising that there are also 'internal gravity waves,' with a water parcel displaced vertically feeding a gravitational restoring force. The wave periods depend on the degree of stratification but may be as short as several minutes and can be long enough that the Coriolis force plays a major role in the dynamics. Vertical displacements are typically of the order of ten meters or so, with horizontal excursions of several hundred meters. The associated horizontal currents are typically several tens of millimeters per second. An interesting difference from the surface wave field is that internal waves always seem to be present, without the intense storms or periods of calm that exist at the surface.

The existence of internal waves complicates the mapping of average currents and depths of particular density surfaces. They have also been the objective of intensive military-funded research because of the possibility that wakes of internal waves generated by submarines might be detectable by remote sensing, thus betraying the submarine's location. More conventional acoustic means of submarine detection are complicated by the deflection of acoustic rays by the rather random variations in sound speed induced by internal waves. In civilian activities, the currents and buoyancy changes associated with internal waves are a matter of concern in offshore oil drilling.

Most importantly, perhaps, the current shear of internal waves, including those of tidal frequency, can lead to instability and turbulence, and so the waves are the main agent for vertical mixing in the ocean interior. This mixing plays a major role in determining the strength of ocean circulation, and hence the poleward heat flux and climate. The mixing, along with the associated circulation, also provides nutrient fluxes into the sunlit upper ocean where primary biological production occurs. Under-