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COASTAL TRAPPED WAVES

J. M. Huthnance, CCMS Proudman Oceanographic Laboratory, Wirral, UK

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Introduction

Many shelf seas are dominated by shelf-wide motions that vary from day to day. Oceanic tides contribute large coastal sea-level variations and (on broad shelves) large currents. Atmospheric pressure and (especially) winds generate storm surges; strong currents and large changes of sea level. Other phenomena on these scales are wind-forced upwelling, along-slope currents and poleward undercurrents common on the eastern sides of oceans, responses to oceanic eddies, and alongshore pressure gradients.

All these responses depend on natural waves that travel along or across the continental shelf and slope. These waves, which have scales of about one to several days and tens to hundreds of kilometers according to the width of the continental shelf and slope, are the subject of this article. Also included are 'Kelvin' waves, also coastally trapped, that travel cyclonically around ocean basins but with typical scales of thousands of kilometers both alongshore and for offshore decrease of properties.

The waves have been widely observed through their association with the above phenomena. In fact they have been identified along coastlines of various orientations and all continents in both the Northern and Southern Hemispheres. Typically, the identification involves separating forced motion from the accompanying free waves. The 'lowest' mode with simplest structure (see below) has been most often identified; its peak coastal elevation is relatively easily measured. More complex forms need additional offshore measurements (usually of currents) for identification. This has been done (for example) off Oregon, the Middle Atlantic Bight and New South Wales (Australia). Observations substantiate many of the features described in the following sections.

Formulation

Analysis is based on Boussinesq momentum and continuity equations for an incompressible sea of near-uniform density between a gently-sloping seafloor $z = -h(\mathbf{x})$ and a free surface $z = \eta(\mathbf{x}, t)$ where the surface elevation $\eta = 0$ for the sea at rest. Cartesian coordinates $\mathbf{x} \equiv (x, y), z$ (vertically up) rotate with a vertical component $f/2$. The motion, velocity components (u, v, w), is assumed to be nearly horizontal and in hydrostatic balance. (These assumptions are almost always made for analysis on these scales; they are probably not necessary but certainly simplify the analysis.) At the surface, pressure and stress match atmospheric forcing (for free waves). There is no component of flow into the seabed (generalizing to zero onshore transport uh at the coast); $u \to 0$ far from the coast (the trapping condition) or is specified by forcing.

Straight Unstratified Shelf

This is the simplest context. Taking x offshore (and y alongshore; **Figure 1**) the depth is $h(x)$. Uniformity along shelf suggests wave solutions $\{u(x), v(x), \eta(x)\}$ $\exp(iky + i\sigma t)$. For positive wave frequency, σ, $k > 0$ corresponds to propagation in $-y$, with the coast on the right ('forward' in the Northern Hemisphere). Then the momentum equations give u, v in terms of η satisfying

$$(h\eta')' + K\eta = 0 \qquad [1]$$

where $K(\alpha) \equiv kfh'/\sigma + (\sigma^2 - f^2)/g - k^2 h$ (uniform f); primes ($'$) denote cross-shelf differentiation $\partial/\partial x$. The boundary conditions become

$$h(\sigma\eta' + fk\eta) \to 0 \, (x \to 0); \quad \eta \to 0 \, (x \to \infty) \qquad [2]$$

Free wave modes are represented by eigensolutions of eqns. [1] and [2]. Successive modes with more offshore nodes correspond to large positive K and arise in two ways.

The term $(\sigma^2 - f^2)/g - k^2 h$ in K represents the gravity wave mechanism, modified by rotation; it increases with frequency σ. For $kf > 0$ it gives rise to the 'Kelvin wave' – the mode with simplest offshore form; decay but no zeros of elevation. Other forms depending on this term ($kf < 0$ and/or nodes of elevation offshore) are termed edge waves and discussed elsewhere. If there is no slope or boundary, K equals this term alone and plane inertiogravity waves are solutions of eqn. [1].

The term kfh'/σ in K increases with decreasing σ if everywhere h increases offshore and $kf > 0$. It represents the following potential vorticity (angular momentum) restoring mechanism. If fluid is displaced into shallower water, it spreads laterally to conserve volume and therefore spins more slowly in total; taking account of the earth's rotation, it acquires anticyclonic relative vorticity. 'Forwards' along the slope, the resulting up-slope velocity implies an up-slope displacement in time. Hence the up-slope displacement propagates 'forwards' along the slope. (Behind it, the anticyclonic relative vorticity implies down-slope flow restoring the fluid location from its previous up-slope displacement.) This sequence is depicted in **Figure 2**. These modes are referred to as continental shelf waves.

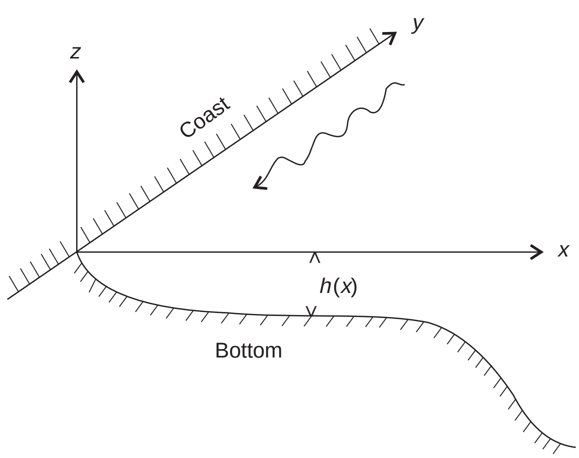

Figure 1 The wavy arrow denotes the sense of wave propagation for $k > 0$. This is 'forwards' if $f > 0$ (Northern Hemisphere).

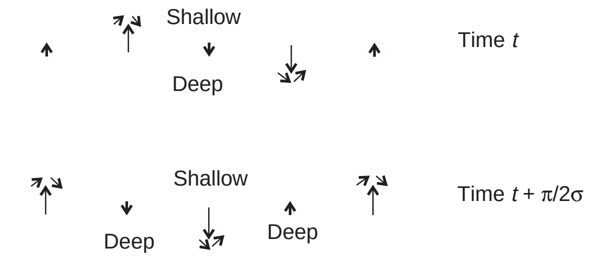

Figure 2 Topographic wave mechanism. ↑ displacement, ↑ velocity, ↗ ↘ relative vorticity (Northern Hemisphere).

The mode forms and frequencies are known for several analytic models, e.g., level, uniformly sloping, exponential concave and convex shelves bordering an ocean of uniform depth. Numerical solutions for the waveforms and dispersion relations $\sigma(k)$ are easily found for any depth profile $h(x)$.

For any monotonic profile $h(x)$ the following have been proved. Phase propagates 'forwards' for all modes with $\sigma < |f|$. Waves forms with 1, 2,... nodes offshore have frequencies $|f| > \sigma_1 > \sigma_2 > \ldots$ defined for all k (subject to $kf > 0$). The Kelvin wave frequency $\sigma_0(k) > \sigma_1(k)$ is likewise defined for all k ($kf > 0$) and passes smoothly through $|f|$ to $\sigma_0 > |f|$ for large enough k. Edge waves with 0 (if $kf < 0$), 1, 2,... nodes in the offshore form have increasing frequencies $\sigma > |f|$; however, low wavenumbers (and frequencies) are excluded where the dispersion curves break the trapping criterion $0 < K(\infty) \equiv (\sigma^2 - f^2)/g - k^2 h(\infty)$ (**Figure 3**).

Besides these properties, the following features are typical. Bounded h'/h ensures a maximum σ_M in $\sigma(k)$; near σ_{1M}, mode 1 velocity tends to be maximal near the shelf edge, and polarized anticyclonically; the nearest approach to inertial motion in this topographic context; here the group velocity $\partial\sigma/\partial k$ of energy propagation (in $-y$) reverses through zero. If $k \to \infty$, then $\sigma_n \to f/(2n+1)$ for the n-node continental shelf wave which becomes concentrated over the 'beach' at the coast. As σ, $k \to 0$, the shelf wave and Kelvin wave speeds σ/k approach constant (maximum) values and $u/\sigma, v, \eta$ approach constant

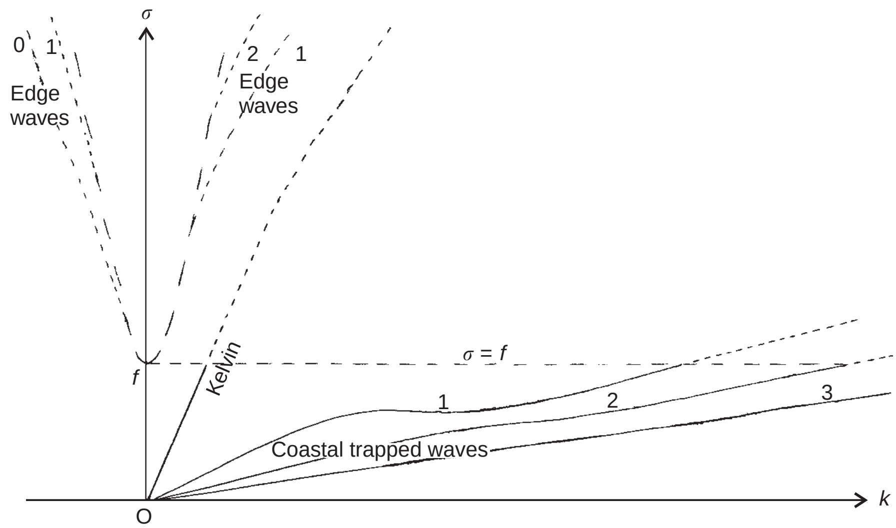

Figure 3 Qualitative dispersion diagram. —— trapped waves, - - - - nearly trapped waves, – – – $\sigma^2 = f^2 + k^2 gh(\infty)$.

forms so that cross-slope velocities tend to zero. Variables v, η and η' are in phase or antiphase, v and η' being near the geostrophic balance $fv = g\eta'$; u is 90° out of phase. Typically, Kelvin-wave currents in shallow shelf waters are polarized cyclonically but first-mode continental shelf wave currents are anticyclonic.

Continental shelf wave-forms depend on the shape rather than the horizontal scale L of the depth profile. Phase speeds scale as fL and (u, v, η) scale as $(\sigma U/f, U, fUL/g)$ where the velocity scale U may be typically $0.1\,\mathrm{m\,s^{-1}}$. Kelvin wave forms depend more on the depth; usually the phase speed is just less than $[gh(\infty)]^{1/2}$ and (u, v, η) scale as $(\sigma ZL/h, (g/h)^{1/2}Z, Z)$ where Z is typically 0.1 to 1 m. Quantitative results depend on the strength of forcing and accurate profile modeling; numerical calculations should be used for real shelves.

The typical maximum in $\sigma(k)$ and associated reversal of group velocity $\partial\sigma/\partial k$ appears to be of practical significance. Shelf waves with frequency near the maximum (for some mode) appear in several observations, e.g., North Carolina sea levels, Scottish and Vancouver Island diurnal tides, wind-driven flow north of Scotland and over Rockall Bank. There may be a bias in seeking motion correlated with local forcing, i.e., responses with non-propagating energy. **Figure 4** shows modeled rotary currents over the shelf edge, continental shelf waves near the maximum frequency with slow energy propagation, as a response to impulsive wind forcing.

Other Geometry

Continental shelf waves exist in more general contexts than a straight shelf, as identification in nature testifies. Analyses also verify their possibility in rectangular and circular basins. Perfect trapping around islands is only possible if $\sigma < |f|$; then results are qualitatively as for a straight shelf except that wavelength around the island (and hence frequency) is quantized.

For a broad shelf (distant coast), with the continental slope regarded as a scarp, again only waves in $\sigma < |f|$ are trapped and results are qualitatively as for a straight shelf. An exception is the lowest mode, a 'double' Kelvin wave decaying to both sides. A seamount is again similar but introduces the same quantization as an island.

A ridge comprises two scarps back-to-back. Each has its set of waves propagating 'forwards' (relative to the local slope) in $\sigma < |f|$; a double Kelvin wave is associated with any net depth difference. Edge waves also propagate in both senses for $\sigma/|f|$ large enough to make $K > 0$ (see eqn. [1]). Similarly, a trench has sets of waves appropriate to each side.

All cases $h(x)$ and radial geometry $h(r)$ can be treated numerically in the same way as a straight monotone profile.

Stratification

We consider the simplest context, a straight shelf with rest state density $\rho_0(z)$; let $N^2 \equiv -g/\rho_0 \mathrm{d}\rho_0/\mathrm{d}z$. A wave form $\{u(x,z), v(x,z), w(x,z), \rho(x,z), p(x,z)\}$ $\exp(iky + i\sigma t)$ is posed. Hydrostatic balance, density, and momentum equations give ρ, w, u and v, respectively, in terms of p; then

$$\partial^2 p/\partial x^2 + (f^2 - \sigma^2)\partial/\partial z(N^{-2}\partial p \partial z) - k^2 p = 0 \quad [3]$$

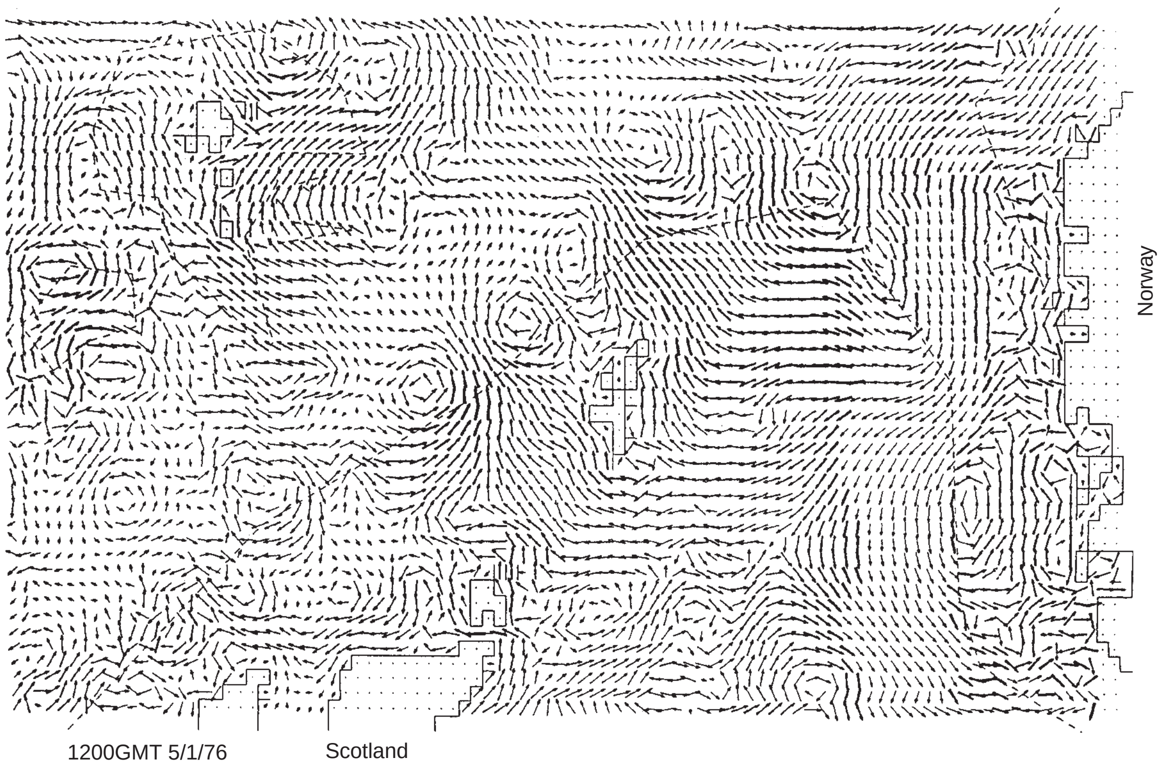

Figure 4 Modeled continental shelf waves around Scotland at 1200Z, 5 January 1976 after impulsive wind forcing near Shetland on 3 January 1976. Dashed line shows 200 m depth contour. (Reproduced with permission from Huthnance JM (1995) Circulation, exchange and water masses at the ocean margin: the role of physical processes at the shelf edge. *Progress in Oceanography* 35: 353–431.)

by continuity. The boundary conditions for no flow through the bottom, zero pressure at the surface and trapping become

$$\partial h/\partial x(\partial p/\partial x + fkp/\sigma) + (f^2 - \sigma^2)N^{-2}\partial p/\partial z = 0 \quad (z = -h) \qquad [4]$$

$$\partial p/\partial z + N^2 p/g = 0 \quad (z = 0) \quad p \to 0\,(x \to \infty) \qquad [5]$$

A flat bottom (uniform h) with uniform $N^2 \ll g/h$ admits the simplest Kelvin wave and (for $n \geq 0$) internal Kelvin wave solutions

$$\cos[N(z + h)/c_n] \exp[i\sigma t + i\sigma y/c_n - fx/c_n] \qquad [6]$$

where $c_n = (gh)^{1/2}(n = 0), c_n = Nh/n\pi\,(n = 1, 2, \ldots)$. Similar solutions, with vertical structure distributed roughly as N, exist for any $N(z) > 0$. Propagation is 'forwards' (cyclonic around the deep sea; anticyclonic around a cylindrical island) but depends only on density stratification.

For a sloping bottom $h(x)$ with offshore scale L (shelf width) and depth scale H, the parameter $S \equiv N^2H^2/f^2L^2$ indicates the importance of stratification. For small S, the solutions in $\sigma < |f|$ are depth-independent Kelvin and continental shelf waves. As S increases, wave speeds σ/k increase and nodes of u, v, p in the (x, z) cross-section tilt outwards from the vertical towards horizontal. Correspondingly, seasonal changes have been observed in the vertical structure and offshore scale of currents on the Oregon shelf (for example) and off Vancouver Island, where stratification increases the offshore decay scale of upper-level currents. Quite moderate S may imply σ monotonic increasing in k, contrasting with the maximum in $\sigma(k)$ common among unstratified modes. For large S, the modes in $\sigma < |f|$ become internal Kelvin-like waves with x replaced by $x - h^{-1}(z)$; $S \to \infty$ corresponds to a shelf width L much less than the internal deformation scale NH/f; the slope is 'seen' only as a coastal wall. Internal Kelvin waves have been observed in the Great Lakes and around Bermuda, where the bottom slope is steep. Similarly, records off Peru show an offshore scale ~ 70 km, greater than the shelf width, because c_n/f is large near the equator.

For $\sigma > |f|$ and nonzero S, trapping is imperfect. However, there are frequencies $\sigma(k)$ at which waves are almost trapped, radiate energy only slowly or respond with maximal amplitude to sustained forcing. These $\sigma(k)$ appear to correspond to dispersion curves in $\sigma < |f|$ but there is still some uncertainty about the role of these waves; they may need some oceanic forcing.

Bottom-trapped waves are an idealized form with motion everywhere parallel to a plane sloping seafloor (in uniform N^2); they decay away from the seafloor. They may propagate for $\sigma < |f|$ or $\sigma > |f|$ and up or down the slope, but always with a component 'forwards' along the slope. If this phase propagation direction is ϕ relative to the alongslope direction, then the velocity being transverse is at angle ϕ to the slope and the frequency is $\sigma = N\,\partial h/\partial x \cos\phi$. In $\sigma < |f|$ the general coastal trapped wave form tends for large k to bottom-trapped waves confined near the seafloor maximum of $N\partial h/\partial x < |f|$; then $\sigma = N\partial h/\partial x$. If the maximum $N\partial h/\partial x > |f|$, then σ increases to $|f|$ as k increases, a qualitative difference from unstratified behavior; formally there is a smooth transition at $\sigma = |f|$. Bottom-trapped wave identification may be difficult (despite the apparent prevalence of near-bottom currents), requiring knowledge of the local slope and stratification. There is evidence from continental slopes off the eastern USA and NW Africa.

Friction

Friction causes cross-shelf phase shifts and significant damping of coastal trapped waves. The depth-integrated alongshore momentum balance for idealized uniform conditions ($\partial p/\partial y = 0, \int udz = 0$) is $\partial v/\partial t + rv/h = \tau/\rho h$ suggesting that the flow v lags the forcing stress τ less for low frequency, shallow water, and large friction r. For example, nearshore currents lag the wind less than currents in deeper water offshore. Damping rates may be estimated as $r/h = O(0.003\,U\ h^{-1})$, i.e., a decay time less than 4 days for a typical current $U = 0.1\,\mathrm{m\,s^{-1}}$ and depth $h = 100\,\mathrm{m}$. In this estimate of friction, U should represent all currents present, e.g., tidal currents can provide strong damping. This decay time converts in to a decay distance $c_g h/r$ for a wave with energy propagation speed c_g. Such decay distances are largest (hundreds to a thousand kilometers or more) for long waves with 'forward' energy propagation; much less for (short) waves with 'backward' energy propagation.

Mean Flows

Mean currents are significant in many places where continental shelf waves have been observed, e.g., adjacent to the Florida current. Waves with phase speeds of a few meters per second or less may be significantly affected by boundary currents of comparable speed, or by vorticity of order f. By linear theory, advection in a uniform mean current V is essentially trivial, but could reverse the propagation of slower waves (higher modes or short wavelengths). Shear $V' \equiv \mathrm{d}V/\mathrm{d}x$ modifies the background potential vorticity to $P(x) \equiv (f + V')/h$; then the gradient of P (rather than f/h) underlies continental shelf wave propagation.

'Barotropic' instability is possible if V is strong enough; necessary conditions are $P'(x_s) = 0$ (some x_s) and $P'[V(x_s) - V] > 0$ for some x; the growth rate is bounded by max $|V'/2|$. Gulf Stream meanders have been interpreted as barotropically unstable shelf waves from Blake Plateau. In a stratified context, these effects of mean flow V may still apply. Additionally, vertical shear $\partial V/\partial z$ is associated with horizontal density gradients:

$$f\partial V/\partial z = -g\rho_0^{-1}\partial\rho/\partial x \qquad [7]$$

and associated 'baroclinic' instability extracting gravitational potential energy. Two-layer models represent $\partial\rho/\partial x$ by a sloping interface; the varying layer depths are another source of gradients $\partial P/\partial x$ in each layer. Thus baroclinic instability may occur even if the current and total depth are uniform. More generally, slow flows over a gently sloping bottom are unstable only if $\partial P/\partial x$ has both signs in the system. Such a two-layer channel model predicts instability at peak-energy frequencies in Shelikov Strait, Alaska (for example) and a corresponding wavelength roughly matching that observed. However, we caution that two-layer models may unduly segregate internal Kelvin and continental shelf wave types, say, exaggerating the multiplicity of wave forms and scope for instability.

In continuous stratification, the equivalent potential vorticity gradient $\partial f/\partial x + \partial^2 V/\partial x^2 + f^2\partial(N^{-2}\partial V/\partial z)/\partial z$ may support waves in the interior. Density contours rising coastward in association with a shelf edge surface jet modify the fastest continental shelf wave to an inshore 'frontal-trapped' form. The bottom slope also supports waves. Over a uniform bottom slope, additional bottom features can couple and destabilize the interior and bottom modes, even if $V(x, z)$ is otherwise stable. However, the bottom slope stabilizes bottom-intensified waves under an intermediate uniformly stratified layer.

Non-linear Effects

These mirror typical nonlinear effects for waves. For example, each part of the nonlinear Kelvin wave form moves with the local speed $v + [g(h+\eta)]^{1/2}$; crests (η and v positive) gain on troughs (η and v negative) and wave fronts steepen. Similarly, for internal Kelvin waves between an upper layer, depth h, and a deep lower layer (density difference $\Delta\rho$) the local speed is $(gh\Delta\rho/\rho)^{1/2}$ everywhere; troughs gain on crests. Dispersion limits this nonlinear steepening; the associated shorter wavelengths propagate more slowly; and the steepening is left behind by the wave. For small amplitudes and long waves, steepening and dispersion are small and can balance in permanent-form $\mathrm{sech}^2(ky+\sigma t)$ solutions. Steepening may be important for internal Kelvin waves, but for typical continental shelf wave amplitudes (near linear) it takes weeks or months, longer than likely frictional decay times.

Mean currents may be forced via frictional contributions to the time-averaged nonlinear convective derivatives in the momentum equations. The mean flows, scale $h^{-1}h'f^{-1}\hat{u}^2$, are typically confined close to the coast or the shelf break ($\hat{u}$ denotes on–offshore excursion in the waves). Another mechanism is wave-induced form drag over an irregular bottom, giving a biased response to variable forcing; a 'forward' flow along the shelf. For example, low-frequency sinusoidal wind forcing may give a mean current up to a maximum fraction $(2\pi)^{-1}$ of the value under a steady wind, i.e., some centimeters per second, principally in shallower shelf waters.

Three-way interactions between coastal-trapped waves can occur if $\sigma_3 = \sigma_1 \pm \sigma_2$, $k_3 = k_1 \pm k_2$, possible for particular combinations according to the shape of the dispersion curve. Typical timescales for energy exchange are many days; effects may be masked by frictional decay. Near a group velocity of zero, there may be more response to a range of energy inputs.

Alongshore Variations

If changes in the stratification and continental shelf form are small in one wavelength, then individual wave modes conserve a longshore energy flux; local wave forms are as for a uniform shelf. Thus Kelvin wave amplitudes increase as $f^{1/2}$ and are confined closer to the coast at higher latitudes. Energy flux conservation implies a large amplitude increase if waves of frequency σ approach a shelf region where the maximum (σ_M) for their particular mode is near σ, as for Scottish and Vancouver Island diurnal tides; the energy tends to 'pile up.' If shelf variations cause σ_M to fall well below σ, then the waves are totally reflected, with large amplitudes near where $\sigma_M = \sigma$. Poleward-propagating waves experience changing conditions. Near the equator, f is small, S (effective stratification) is large and internal Kelvin-like waves are expected, as off Peru; as f increases poleward, waves evolve to less stratified forms, more like continental shelf waves. (Variations of f are special in supporting offshore energy leakage to Rossby waves in the ocean.)

Small irregularities in the shelf (lateral, vertical scales $\varepsilon(gH)^{1/2}/f, \varepsilon H$) generally cause $O(\varepsilon^2)$ effects, but $O(\varepsilon)$ nearby and in phase shifts after depth changes. Scattering occurs, preferentially to adjacent wave modes and (if unstratified) the highest mode at the incident frequency (having near-zero group velocity). However, long waves, L_W, on long topographic variations (as above; L_T) adopt the appropriate local form; scattering is slow unless $L_W \sim L_T$.

If the depth profile has a self-similar form $h[(x-c(y))/L(y)]$ then long ($\gg L$) continental shelf waves propagate with changes of amplitude but no scattering or change of form, provided that c and L also vary slowly ($c/c', L/L' \gg L$). Likewise, there is no scattering if the depth is $h(\xi)$ where $\nabla^2\xi(x,y) = 0$, representing approximately uniform topographic convexity. In these cases, stronger currents and shorter wavelengths are implied on narrow sections of shelf.

Abrupt features are apt to give the strongest scattering, substantial local changes or eddies on the flow. Scattering is the means of slope–current adjustment to a changed depth profile. However, all energy must remain trapped in $\sigma < |f|$; even in $\sigma > |f|$ special interior angles $\pi/(2n+1)$ can give perfect Kelvin wave energy transmission (for example). Successive reflections in a finite shelf (embayment) may synthesize near-resonant waves with small energy leakage. A complete barrier across the shelf implies reflection into (short, slow) waves of opposite group velocity. There is a considerable literature of particular calculations. However, it is difficult to generalize, because of the several non-scattering cases interspersed among those with strong scattering.

Generation and Role of Coastal-trapped Waves

Oceanic motion may impinge on the continental shelf. Notably at the equator, waves travel eastward to the coast and divide to travel north and south. In general, oceanic motions accommodate to the presence of the coast and shelf; at a wall, by internal Kelvin waves (vertical structure modes); for more realistic shelf profiles, by coastal trapped waves. Oceanic signals tend to be seen at the coast if along-shelf scale > wave-decay distance or if the feature is shallower than the shelf-water depth.

Natural modes of the ocean are significantly affected by continental shelves. Modes depending on f/h gradients have increased frequencies and forms concentrated over topography. Numerical models have shown 13 modes with periods between 30 and 80 hours, each mode being localized over one shelf area.

Atmospheric pressure forcing the sea surface can be effective in driving Kelvin and edge waves, especially if there is some match of speed and scale, more likely in shallower (shelf) seas.

Longshore wind stress τ is believed to be the most effective means of generating coastal trapped waves. Within the forcing region, the flow tends to match the wind field; when or where the forcing ceases, the wave travels onwards ('forwards') and is then most recognizable. A simple view of this forcing is that τ accelerates the alongshore transport $h\mathbf{u}$. A more sophisticated view is that τ induces a cross-shelf surface transport $|\tau|/\rho f$; coastal blocking induces a compensating return flow beneath, which is acted upon by the Coriolis force to give the same accelerating alongshore transport. A typical stress $0.1\,\mathrm{N\,m^{-2}}$ for $10^5\,\mathrm{s}$ (~ 1 day) accelerates 100 m water to $0.1\,\mathrm{m\,s^{-1}}$. Winds blowing across depth contours may be comparably effective if the coast is distant.

Other generation mechanisms include scattering (of alongshore flow, especially) by shelf irregularities as above, variable river runoff, and a co-oscillating sea.

Shelf-sea motion is often dominated by tides and responses to wind forcing. On a narrow shelf, oceanic elevation signals penetrate more readily to the coast as the higher-mode decay distances are short; the tide is represented primarily by a Kelvin wave spanning the ocean and shelf. Model fits to semidiurnal measurements, showing a dominant Kelvin wave, have been made off California, Scotland and north-west Africa, for example. Several areas at higher latitudes show dominant continental shelf wave contributions to diurnal tidal currents, e.g., west of Scotland, Vancouver Island, Yermak Plateau. On a wide shelf, there is correspondingly greater scope for wind-driven elevations and currents. The extensive forcing scale, typically greater than the shelf width, induces flow with minimal structure (typically no reversals across the shelf) corresponding to the lowest-mode continental shelf wave with maximum elevation signal at the coast. In stratified conditions, the upwelling or downwelling response also corresponds to a wave (or waves).

As a wave travels, its amplitude is continually incremented by local forcing. A model based on this approach was used to make the hindcast of measured currents on the south-east Australian shelf shown in **Figure 5**. At any fixed position, the motion

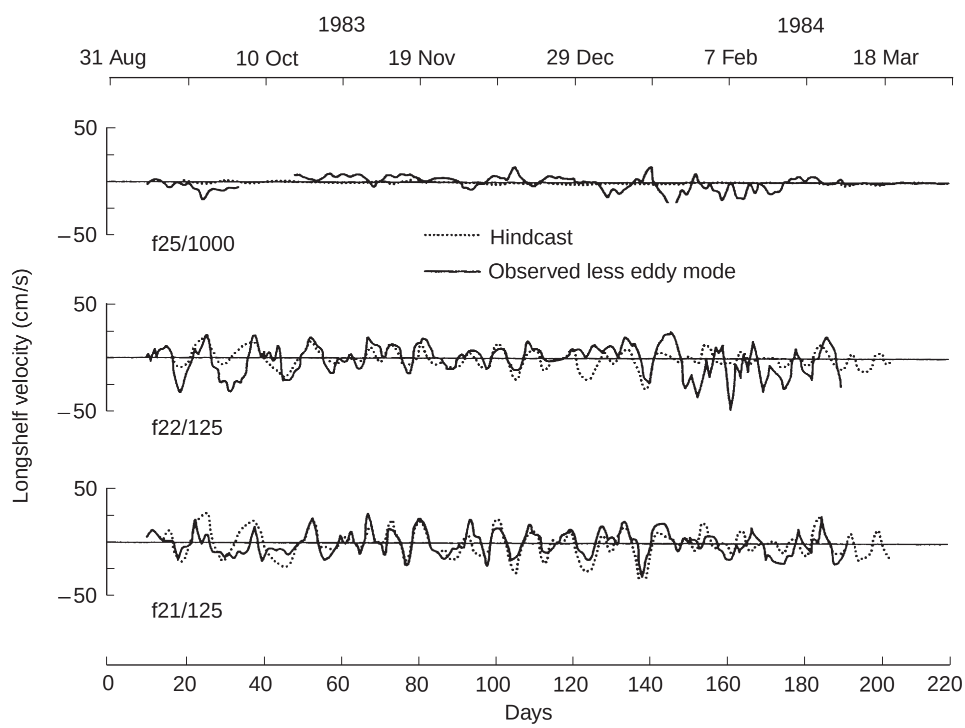

Figure 5 Comparison of hindcast along-shelf currents using three coastal-trapped wave modes with measured currents in a cross-shelf section after band-pass filtering and removing an eddy mode. (Reproduced with permission from Church JA, White NJ, Clarke AJ, Freeland HJ and Smith RL (1986). Coastal-trapped waves on the East Australian continental shelf. Part II: model verification. *Journal of Physical Oceanography* 16: 1945–1957.)

results from local forcing and from arriving waves, bringing the influence of forcing (e.g., upwelling) 'forwards' from the 'backward' direction. In the Peruvian upwelling regime, for example, variable currents are not well correlated with local winds but include internal Kelvin-like features coming from nearer the equator. This is hardly compatible with (common) simplifications of a zero alongshore pressure gradient. Moreover, the waves carry the influence of assumed 'backward' boundary conditions far into a model. The same applies for steady flow. Friction introduces a 'forward' decay distance for a coastal-trapped wave; this distance has a definite low-frequency limit. Currents decay over these distances according to their structure as a wave combination. Thus alongshore evolution or adjustment of flow (however forced) is affected by coastal-trapped waves whose properties should guide model design.

Summary

This article considers waves extending across the continental shelf and/or slope and having periods of the order of one day or longer. Their phase propagation is generally cyclonic, with the coast to the right in the Northern Hemisphere, a sense denoted 'forward'; cross-slope displacements change water-column depth and relative vorticity, causing cross-slope movement of adjacent water columns. At short-scales, energy propagation can be in the opposite 'backward' sense. Strict trapping occurs only for periods longer than half a pendulum day; shorter-period waves leak energy to the deep ocean, albeit only slowly for some forms. The waves travel faster in stratified seas and on broad shelf-slope profiles; speeds can be affected, even reversed, by along-shelf flows and reverses of bottom slope. Large amplitudes and abrupt alongshore changes in topography cause distortion and transfers between wave modes. The waves form a basis for the behavior (response to forcing, propagation) of shelf and slope motion on scales of days and the shelf width. Hence, they are important in shelf and slope–sea responses to forcing by tides, winds (e.g., upwelling), density gradients, and oceanic features. Their propagation (distance before decay) implies nonlocal response (over a comparable distance), especially in the 'forward' direction.

See also

Coastal Circulation Models. Internal Tides. Internal Waves. Regional and Shelf Sea Models. Rossby Waves. Storm Surges. Tides. Upper Ocean Responses to Strong Forcing Events. Vortical Modes. Wind Driven Circulation.

Further Reading

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COASTAL ZONE MANAGEMENT

Seoung-Yong Hong, Ministry of Maritime Affairs and Fisheries (MOMAF), Seoul, South Korea
Chan-Wen Lee, Kyungnam University, Masan, South Korea

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Introduction

Developed and developing nations alike depend significantly on the resources and transportation opportunities in the marine environment. In most countries with a marine coast, development-oriented national policies have led historically to the concentration of populations and industrial activities in areas adjacent to the ocean known as the coastal zone. For many developing nations, shipping, fishing, aquaculture, and coastal tourism are vitally important to their economies.

Notwithstanding this importance, coastal resources are often developed with a land-oriented perspective that fails to consider the unique physical