Circulation, exchange and water masses at the ocean margin: the role of physical processes at the shelf edge

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Abstract — The coastal ocean meets the deep sea at the continental shelf edge. Questions of global change entail elucidation of the processes that determine the quantities, transformation and fate of materials transported between the shelf and ocean, the measurement and definition of exchange processes, and the development of prognostic models of exchanges.

Physical processes control the large-scale movement and irreversible small-scale mixing of water and its constituents. At the shelf edge, steep bathymetry may inhibit ocean-shelf exchange, but in combination with stratification gives rise to special processes and modelling challenges.

A preliminary assessment is made of coastal-trapped waves; along-slope currents, instability and meanders; eddies; upwelling, fronts and filaments; downwelling, cascading; tides, surges; internal tides and waves as potentially influential processes in ocean-shelf exchange, water-mass structure and general circulation, according to their scales and context. For this purpose, theory and previous measurements are interpreted.

Future studies needed to improve this assessment are discussed.

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<table>
<thead>
<tr>
<th>Symbol</th>
<th>Meaning</th>
<th>Value</th>
<th>Unit</th>
</tr>
</thead>
<tbody>
<tr>
<td>a</td>
<td>surface wave amplitude</td>
<td>1</td>
<td>m</td>
</tr>
<tr>
<td>ADCP</td>
<td>Acoustic Doppler Current Profiler</td>
<td></td>
<td></td>
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<tr>
<td>AVHRR</td>
<td>Advanced Very High Resolution Radiometer</td>
<td></td>
<td></td>
</tr>
<tr>
<td>A/hb</td>
<td>marginal sea area / entrance strait cross-section</td>
<td>$10^5$</td>
<td>—</td>
</tr>
<tr>
<td>c</td>
<td>characteristic slope for internal wave motion (§7)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>C</td>
<td>specific heat (for water)</td>
<td>4.2</td>
<td>J/(g°C)</td>
</tr>
<tr>
<td>C^2</td>
<td>constituent concentration</td>
<td>0.003</td>
<td>—</td>
</tr>
<tr>
<td>CD</td>
<td>quadratic bottom friction coefficient</td>
<td></td>
<td></td>
</tr>
<tr>
<td>CTD</td>
<td>conductivity-temperature-depth instrument</td>
<td></td>
<td></td>
</tr>
<tr>
<td>d</td>
<td>deep ocean depth</td>
<td>4000</td>
<td>m</td>
</tr>
<tr>
<td>div</td>
<td>divergence in western boundary current</td>
<td>$10^{-3}$</td>
<td>km$^{-1}$</td>
</tr>
<tr>
<td>EU</td>
<td>European Union</td>
<td></td>
<td></td>
</tr>
<tr>
<td>f</td>
<td>Coriolis parameter</td>
<td>$10^{-4}$</td>
<td>s$^{-1}$</td>
</tr>
<tr>
<td>g</td>
<td>gravitational acceleration</td>
<td>10</td>
<td>ms$^{-2}$</td>
</tr>
<tr>
<td>g'</td>
<td>reduced gravity: $g \times$ density change across thermocline / $\rho$</td>
<td>0.01</td>
<td>ms$^{-2}$</td>
</tr>
<tr>
<td>h</td>
<td>water depth (shelf and slope)</td>
<td>100</td>
<td>m</td>
</tr>
<tr>
<td>h_x</td>
<td>depth gradient across slope</td>
<td></td>
<td></td>
</tr>
<tr>
<td>h_o</td>
<td>depth of principal oceanic circulation or thermocline</td>
<td>1000</td>
<td>m</td>
</tr>
<tr>
<td>h_o/h</td>
<td>ocean current depth / (ocean current depth - shelf-sea depth)</td>
<td>1</td>
<td>—</td>
</tr>
<tr>
<td>h_st</td>
<td>steric slope</td>
<td>$10^{-3}$</td>
<td>—</td>
</tr>
<tr>
<td>h'</td>
<td>depth of seasonal thermocline</td>
<td>25</td>
<td>m</td>
</tr>
<tr>
<td>H</td>
<td>surface heat flux (winter cooling)</td>
<td>100</td>
<td>Wm$^{-2}$</td>
</tr>
<tr>
<td>JEBAR</td>
<td>Joint Effect of Baroclinicity and Relief</td>
<td></td>
<td></td>
</tr>
<tr>
<td>k</td>
<td>linear bottom friction coefficient</td>
<td>1</td>
<td>mm/s</td>
</tr>
<tr>
<td>k_f</td>
<td>linear friction coefficient below oceanic circulation</td>
<td>$\frac{1}{2}$</td>
<td>mm/s</td>
</tr>
<tr>
<td>K</td>
<td>coefficient of (lateral) diffusion</td>
<td></td>
<td></td>
</tr>
<tr>
<td>L_t</td>
<td>topographic length scale $h/h_z$ (over steep slope)</td>
<td>10</td>
<td>km</td>
</tr>
<tr>
<td>L_z/L_y</td>
<td>ratio of zonal to meridional scales, oceanic gyre</td>
<td>5</td>
<td></td>
</tr>
<tr>
<td>M_2</td>
<td>lunar semi-diurnal tidal constituent</td>
<td></td>
<td></td>
</tr>
<tr>
<td>N^2</td>
<td>squared Brunt-Väisälä frequency = $-g \rho^{-1} \partial \rho / \partial z$</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Ri</td>
<td>Richardson number = $N^2 \partial u / \partial z^2$</td>
<td></td>
<td></td>
</tr>
<tr>
<td>R_i</td>
<td>Internal Rossby deformation radius = $(g' h')^{1/2} / f$ or $Nh/f$</td>
<td></td>
<td></td>
</tr>
<tr>
<td>SPM</td>
<td>Suspended Particulate Matter</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Sv</td>
<td>1 Sverdrup volume transport</td>
<td>$10^6$</td>
<td>m$^3$s$^{-1}$</td>
</tr>
<tr>
<td>t</td>
<td>duration of wind forcing</td>
<td>$10^5$</td>
<td>s</td>
</tr>
<tr>
<td>t_n</td>
<td>time constant for tidal shear dispersion</td>
<td>$10^3$</td>
<td>s</td>
</tr>
<tr>
<td>u</td>
<td>cross-slope velocity</td>
<td>0.1</td>
<td>m/s</td>
</tr>
<tr>
<td>u</td>
<td>vector velocity</td>
<td></td>
<td></td>
</tr>
<tr>
<td>v</td>
<td>along-slope velocity</td>
<td>0.1</td>
<td>m/s</td>
</tr>
<tr>
<td>V_w</td>
<td>western boundary current speed</td>
<td>1</td>
<td>m/s</td>
</tr>
</tbody>
</table>
Shelf-sea and ocean meet at the continental shelf edge, where there is varied and increasing interest. Exchanges and the possibility of an active ocean-shelf boundary, with particular shelf-edge contributions to fluxes, are of topical interest for global fluxes, budgets and their response to climate change and human activities. For example, shelf-sea production may be sustained by nitrate fluxes from the ocean; WALSH (1991) estimates a sufficient global supply and WOLLAST (1993) demonstrates the importance and uncertainty in amount of this supply. (In one particular example, the North Sea, an estimated 80% of nitrogen and phosphorus comes from the North Atlantic; HYDES and EDMUNDS, 1989; BROCKMANN, LAANE and POSTMA, 1990). On the other hand (WOLLAST, 1991) there may be a net flux of organic carbon in the opposite sense, ie. an export from shelf seas. There are uncertainties of order 1 Gt/y in the global budget of organic carbon. A similar amount has been estimated as export from shelf seas to the continental slope below the seasonal thermocline (WOLLAST, 1993) where the issue becomes the proportions further exported to the deep ocean or deposited locally. The character of the slope as a “sink” in this sense depends on the vertical flux of particles and/or convergent near-bed transport. In the net shelf-sea sediment budget, outer-shelf sediment transport may be critical; an off-shelf component here may guarantee loss. In areas of upwelling, deep waters over-saturated with respect to atmospheric CO₂ release unknown quantities of the gas when brought to the sea surface. MACKENZIE (1991) tabulates (after KNAUER, 1987) physical processes and their importance to nutrient supply, production, export to the ocean and particle transports.

In addition, ARM (1978) raised interest in mixing at the continental slope as a possibly significant contributor to oceanic diapycnal mixing globally, emphasised by recent reduced estimates (1 to 2 x 10⁻⁵ m²s⁻¹) of diapycnal mixing in the deep ocean interior (TOOLE, POLZIN and SCHMITT, 1994) in contrast with increased values (> 10⁻⁴ m²s⁻¹) in the presence of increased internal
wave energy near steep slopes (cf. §7). GARRETT (1991) and GARRETT, MACCREADY and RHINES (1993) review this interest for ocean circulation, and describe recent progress.

There is a challenge to match ocean and shelf models with their distinct physical and biogeochemical regimes. Boundary conditions (for deep-ocean models or shelf-sea models) typically relate boundary values and fluxes to interior variables. However, it seems unlikely that both the ocean and shelf can have a realistic boundary condition without some influence of the other, as implied if treated separately in this way. This raises the question of where the "control" lies.

At the shelf edge itself, there are special processes and modelling challenges, because stratification combines with steep bathymetry.

There is a need to elucidate the basic processes that determine the quantities, transformation and fate of materials transported between the shelf and ocean, to measure and define exchange processes, and to develop prognostic models for exchanges. This paper is concerned with physical processes and emphasises ocean-shelf exchange. It builds on and updates some earlier reviews, eg. HUTHNANCE (1981), PIETRAFESA (1983, focusing on the US east and west coasts), WROBLEWSKI and HOFMANN (1989, proceeding in stages from the coast to the ocean and emphasising the eastern US shelf). NITZTROUER and WRIGHT (1994) provide a complementary review of particulate transport across continental shelves (the emphasis is inshore of the shelf-break).

1.1 Physical processes

These control the "bodily" (reversible) movement and the irreversible small-scale mixing of water and its constituents, and thus underlie shelf-water characteristics and many fluxes. (The respective "reversible" and "irreversible" character also tend to be associated with earlier and later stages of a process, eg. upwelling; in an ultimate steady state, mixing maintains a constituent flux negating further distortion of the constituent field by advection).

The importance of physical scales is emphasised by TETT and EDWARDS (1984) who relate biological growth (rate $\mu$) to light and especially vertical mixing. With horizontal mixing presented by a diffusivity $K$, only patches of scale exceeding $\pi(K/\mu)^{1/2}$ survive. (For typical $K = 10$ to $10^3$ m$^2$/s and $\mu \sim 1$/day this scale is 3 to 30 km). Otherwise, mixing smears out the patch in a time less than $1/\mu$.

At present, however, the spatial and temporal measurement scales required for flux estimates are not known. Currents varying on time-scales of days or longer tend to be constrained by geostrophy to flow along depth contours; the net flux of water across a closed depth contour and integrated through depth is a consequence of ageostrophy (Appendix A and Fig. 1). Thus geostrophy inhibits ocean-shelf exchange, but other factors may facilitate it.

(i) Processes enhanced or special to the shelf edge may tend to increase such exchange, as reviewed by HUTHNANCE (1981).

(ii) The geostrophic constraint is relaxed near the equator. For example, MULLER-KARGER, MCCLAIN and RICHARDSON (1988) find that the Amazon outflow plume turns offshore in June to January, according to the wind-related North Brazil Current, which retroreflects to join the North Equatorial Countercurrent flowing across the Atlantic (see also §3.2).

(iii) Friction relaxes the geostrophic constraint, notably in Ekman layers.

(iv) Contributions to a net flux $<uC>$ of constituent $C$ may occur without a net water flux, as water flows (with velocity $u$) onto the shelf with one value $C$ and leaves with another after some non-conservative process. (Examples are: nitrogen consumption by phytoplankton growth; mixing with cooler shelf water in winter; diffusion representing unresolved contributions to
However, the non-conservative element is essential and typically weak. Estimates of exchange across the Hebrides shelf edge, from observed cross-slope currents and salinity difference (Huthnance, 1986), suggest that on-shelf mixing between the on- and off-shelf excursions is only O(10%) effective.

(v) Local, non-linear and time-dependent flows may well be ageostrophic, even if a global view is sought that would lend credence to geostrophy.

Figure 1. Conceptual diagram for §1.1 and Appendix 1. Flow (→) near a shelf-edge depth contour (- - -) with relatively small net cross-slope flux as an integral around the basin (the flow nearly returns to its starting point). On smaller time and space scales, geostrophy is broken, allowing excursions L onto the shelf. Non-conservative processes (here production on the shelf) result in a net import of nitrogen N and export of carbon C (for example). Coastal-trapped waves (§2) propagate this pattern of on-offshelf excursions in the sense of the arrows in the northern hemisphere (i.e. cyclonically).

Indeed, (iv) and (v), together with the geostrophic inhibition of cross-slope flow, suggest that flux contributions <uC> from (spatially- and/or temporally-) fluctuating components of u and C are likely to be equally or more important than mean contributions <u><C>. A priori, one would expect the largest scales breaking the geostrophic constraint to make the largest contributions. This suggests time scales O(1 day) through tidal and meteorological forcing, for example, or length scales O(U/f ~ 2 to 10 km) where U is a typical flow speed (~ 0.2 to 1 m/s) and f is the Coriolis parameter (typically 10^-4 s^-1). However, friction may also break the geostrophic constraint, allowing wind stress to drive upwelling flow across the shelf edge on larger scales (for example).

Unfortunately, there are few known coherent estimates <uC> checked by budget closure. Lentz (1987b) and Rudnick and Davis (1988) (for the same CODE experiment) were aided by the fact that C represented temperature (easily recorded) and a dominant near-uniform upwelling signal provided coherence on the scales measured. Dever and Lentz (1994) also achieved heat and (well-correlated) salt balances on the rather uniform northern Californian shelf in winter and spring. Otherwise, the cross-slope component of velocity at the shelf edge often contains contributions from motion on small time and length scales which are typically aliased by measurements, poorly modelled (Brink, 1991), responsive to small-scale forcing which may be poorly known (Brink, Jacasce and Irish, 1994) and of small magnitude relative to the longshore component. Uncertainties of alignment are magnified in the estimation of cross-slope flux (Kosro, 1987). Another caveat pertaining to estimates <uC> is that the inventory of the constituent C is
only affected by the divergent part of $\langle \mathbf{u} \rangle$, viz. $U_S \langle \mathbf{C} \rangle$ if C is conservative in small-amplitude fluctuations (Loder and Horne, 1991). [Here $U_S$ is the Stokes drift (Lagrangian - Eulerian) velocity $\langle \mathrm{d} \mathbf{t} \cdot \mathbf{V}_u \rangle$.] On the other hand, whereas water moves with the Lagrangian velocity $\langle \mathbf{u} \rangle + U_S$, less mobile near-bed sediment moves in response to the local bed stress and may correspond more closely with the Eulerian velocity $\langle \mathbf{u} \rangle$ (Butman, 1988).

If exchange takes place as a result of many uncorrelated small steps (excursions of constituent-bearing water), then it may be effectively diffusive. By analogy with molecular gas dynamics (Batchelor, 1967), the effective diffusivity scales as $T \langle u^2 \rangle$ or $L \langle u^2 \rangle^\gamma$ where $T$ is the duration and $L$ is the cross-shelf magnitude of an individual excursion.

The uncertain experience with naive estimates $\langle \mathbf{u} \mathbf{C} \rangle$ suggests a more informed approach through process understanding. Such an approach is encouraged by satellite images for the Hebrides shelf edge contrasting (i) bathymetric constraint [coastal zone colour scanner on 17 May 1980, see Pingree and MardeLL (1981)] in a sharp shelf-edge boundary to a coccolithophore bloom, with apparently small cross-slope mixing, and (ii) apparently large cross-slope exchange [infra-red on 13 April 1981, see Booth and Ellett (1983)] with no sharp indicator of the shelf edge.

It is necessary to identify the important processes and to determine their occurrence, magnitude, time- and length-scales, energy and energy available for mixing. Processes need to be related to the controlling context:

- bathymetry: depth profile, shelf width, shelf depth (irregularities deserve special consideration, notably canyons juxtaposing ocean and shelf waters);
- density field: freshwater, alongshore density gradient, stratification (depends on season);
- slope-, tidal- and storm-driven currents, associated friction and form drag, and meteorology (seasonal).

1.2 Previous studies

Process studies will be discussed below within the section treating that process.

There have also been previous studies addressing shelf-edge exchange *per se*, notably at the edge of the Scotian shelf (Smith, 1978), SEEP-I and SEEP-II in different parts of the Middle Atlantic Bight (Walsh, Biscaye and Csanady, 1988; Biscaye, Flagg and Falkowski, 1994), the Coastal Transition Zone Program off northern California (Brink and Cwles, 1991) and ECOMARGE in the Gulf of Lions (Monaco, Biscaye, Soyer, Pocklington and Heussner, 1990). The Coastal Ocean Dynamics Experiment (CODE) off northern California (Beardsley and Lentz, 1987) also provided good data for several exchange phenomena.

At the edge of the Scotian shelf, fluctuating transports of salt, heat and nutrients were found to be concentrated in the motions of longest period, and to be important to shelf-sea budgets (Smith, 1978).

In SEEP-I, fluctuating cross-shelf velocities were $O(0.1 \text{ m/s})$ on the shelf and near the surface, and $O(0.03 \text{ m/s})$ in deeper waters (1250 m) over the slope; the mean offshore flow just exceeded 0.01 m/s at the shelfbreak (Aikman, Ou and Houghton, 1988). The idea of a mid-slope *minimum* in current speeds (Csanady, Churchill and Butman, 1988) was supported by a maximum in the accumulation of detritus (Hecker, 1990). Butman (1988) found a downslope bias $\langle u \rangle = 0.01$ to 0.05 m/s in near-bed currents (down-slope fluctuations occurred more frequently); as an Eulerian estimate this does not imply a net (Lagrangian) movement of water but would encourage downslope sediment transport. In this context, Churchill, Biscaye and Aikman (1988) found that
the amount of suspended particulate matter (SPM) was affected by trawling; SPM was transported seaward on density surfaces from the shelf into the slope region, in the sense of decreasing fluid stress, when the shelf/slope front moved seaward. Correspondingly, Biscaye, Anderson and Deek (1988) concluded that SPM fluxes over the slope entailed lateral transport from the water and sediments of the shelf and upper slope. One storm made a noticeable contribution. However, it was also concluded from SEEP that there was little export of primary production from the shelf to the slope: less than 10% of the annual figure (Biscaye, Anderson and Deek, 1988) because of local grazing, sedimentation, oxidation, decay and along-shelf transport on the shelf (Falkowski, Flagg, Rowe, Smith, Whitleedge and Wirick, 1988). [A larger export fraction found by Walsh, Wirick, Pietrafesa, Whitleedge, Hoge and Swift (1988) relates only to outer-shelf production, and is also rather uncertain, deriving from a difference between off-shelf fluxes at two adjacent moorings]. SEEP-II, further south towards Cape Hatteras (Biscaye, Flagg and Falkowski, 1994) generally supported these findings: mean offshore flows typically less than 0.02 m/s, more offshore at the bottom than near the surface and much smaller than alongshore flows; cross-shelf (mostly tidal) flows O(0.1 m/s); a minimum in total kinetic energy at ~1000 m water depth where there was evidence of maximal depositional flux. An increase of this flux downward in the water column, and the sediment composition corresponding to that resuspended in "events" at the shelf break, suggested a shelf source, buffered by canyons to even out the flux in time (Biscaye and Anderson, 1994). However, the estimated export of biogenic particulates from the shelf is even less, << 5%.

In ECOMARGE, similar conclusions were reached. Episodic currents up to 0.5 m/s occurred. Output from the Rhône river was found to be transported in a benthic nepheloid layer, mostly alongshore to the south-west in the sense of the Liguro-Provençal gyre. SPM was observed to increase in this sense (Monaco, Biscaye, Soyer, Pocklington and Heussner, 1990). Then in moving offshore, nephels tended to follow isopycnals, but the offshore extent was bounded by Levantine Intermediate Water (Durrieu de Madron, Nyffeler and Godet, 1990). Export was preferentially down canyons, notably the Lacaze-Duthiers Canyon at the south-west end of the broader shelf sector, evidence being provided by sediment provenance (Monaco, Courp, Heussner, Carbonne, Fowler and Deniaux, 1990) and a correlation <u>.

Reimers, Jahnke and McCorkle (1992) make a similar inference from deposition rates on the lower slope and rise off central California: upper-slope and shelf organic carbon is exported by local sub-surface episodic advection (via nepheloid layers, debris flows and/or turbidity currents). Canyons are known to be active conduits there. In the Southern California Bight, on the slope adjacent to the San Pedro Basin, Huh, Small, Niemni, Finney, Hickey, Kachel, Gorsline and Williams (1990) correlated mass fluxes to sediment traps over monthly periods with the number of days during which waves were estimated to be sufficient to move sediments at 60 m depth on the adjacent shelf. These mass fluxes decreased offshore, and increased towards the bottom, also suggesting an origin in sediments of the shelf or upper slope. Thunell, Pilskaln, Tappa and Sautter (1994) and Thunell, Moore, Dumond and Pilskaln (1994) found a widely-varying lithogenic component in San Pedro Basin sediment fluxes, which they correlated with variable runoff and storms on the shelf. Huh, Small, Niemni, Finney, Hickey, Kachel, Gorsline and Williams (1990) also associated local turbidites with severe storms in 1969 and the 1979-1980 winter; location at the ends of the Basin axis suggest alongshore transport from the canyons or fans of origin.

These observations illustrate the importance of developing a complete picture of the circulation and interacting processes, in order to estimate cross-slope SPM fluxes. Tracer measurements and subsequent water-mass analysis have provided some integral
estimates of exchange over particular sectors of the shelf edge. For example, using oxygen-18, CHAPMAN, BARTH, BEARDSLEY and FAIRBANKS (1986) find that Scotian Shelf Water continues to comprise about 70% of the shelf water at Chesapeake Bay, 1000 km downstream, putting a limit on slope-water exchange. Using radio-caesium (discharged from the nuclear plant at Sellafield), BRADLEY, SCOTT, BAXTER and ELLETT (1991) construct a box model of elements of the Scottish Coastal Current to infer an Atlantic-water inflow of about 0.2 Sverdrup onto the Main-Hebrides shelf (Ireland to Lewis, about 350 km). (One Sverdrup - Sv - is 10^6 m^3 s^-1).

From these estimates and various process-based figures to follow, 1 Sv / 1000 km, ie. 1 m^2 s^-1 appears to be a convenient unit for cross-slope exchange. Global sediment fluxes appear to have 1 Gt/year as a convenient unit; distributed around the shelf edge of length 3 x 10^5 km, this is very approximately 10 tonnes/day per km.

1.3 Processes in combination

The previous studies show that a broad view should be taken. Physical processes may have different significance for different constituents, as biogeochemical processes and consequent constituent concentrations C have varied distributions. In particular, sediments tend to sink and therefore have greater concentration at depth, unless very fine; bedload transport, erosion and deposition are related to near-bed currents; in turn, the depth-dependence of particles and fluxes has been shown to be affected by cross-slope transport to deeper waters as resuspended sediments [eg. nepheloid layers: DICKSON and MCCAVE (1986); MONACO, BISCAYTS, SOYER, POCKLINGTON and HEUSSNER (1990)]. Primary production takes place in illuminated near-surface waters, possibly exhausting nutrients supplied by local mixing upwards across a seasonal thermocline. At the Celtic Sea shelf edge, a thickened seasonal thermocline has been measured in conjunction with a thickened distribution of increased chlorophyll (PINGREE, MARDELL, HOLLIGAN, GRIFFITHS and SMITHERS, 1982). Additionally, patchy constituent distributions are affected by uniform alongshore currents, even if there is no cross-slope flow.

The relative importance of the "ongoing" versus extreme "events" is an issue, especially with regard to sediment transport, (which is sensitive to extremes of current). MILLIMAN (1991) cites the Santa Clara river (USA) where an 18-year record showed 60% of the sediment transport occurring in just six days. (The significance of such a statement is unclear with regard to transports in tonne-kilometres or "downstream" in the system, but it does highlight the sampling problem for measurements at fixed locations). The Storegga slide was even more extreme, displacing 5700 km^3 down the continental slope (KENYON, 1987), an amount comparable with ~1000 years' fluvial sediment flux at 15 x 10^9 tonnes per year globally. Even cross-slope heat flux may need many years' data to average over long-period fluctuations (eg. BOOTH, 1985). Extremes of current may be associated with extreme forcing (eg. storms) but also with particular conditions of stratification (for example, q.v. §7).

In a similar vein, contributions in different seasons need to be balanced. Forcing of cross-slope flow may be stronger in winter, whereas near-surface stratification in summer may assist cross-slope exchange, by reducing the bathymetric constraint, cf. §3.1 (GAWARKIEWICZ, CHURCH, LUTHER, FRIEDMAN and CARUSO, 1992) and §3.3 (KELLY and CHAPMAN, 1988). [There is a natural scale L_N/N for the vertical extent of topographic influence up through the water column; this scale decreases with stratification N. In fact the direction of geostrophic flow is steered by topography (HIDE, 1971) up to a height at which the flow becomes zero; this will depend also on lateral gradients of density].

Processes interact. The description of this may be simplified if two interacting processes have
temporal or spatial scales which greatly differ, so that (i) the smaller-scale process and its effects can be parametrised (represented by summary statistics which may vary on the larger scale), and (ii) the larger-scale process provides a slowly-varying (quasi-stationary) context in which to set the smaller scale. An example is turbulence (small scale, causing bottom drag) in a tidal current (large scale, but whose shear is the source of the turbulence). Thus the interaction can be estimated although the processes are considered in turn. However, comparable scales imply full two-way interaction and simultaneous consideration of both processes (a special example being self-interaction of a non-linear oscillatory flow to generate a mean current). In neither case are flux contributions from the separate processes simply added to estimate the total.

Some of these issues (and processes) are discussed in BLANTON (1991).

1.4 Outline

The purpose of this paper is to estimate the relative contribution of physical processes to shelf-edge circulation, exchange and mixing, bearing the above considerations in mind and taking the context into account.

In the following sections, we consider the character, previous studies and occurrence of: (§2) coastal-trapped waves; (§3) along-slope currents, western boundary currents and their relation to ocean circulation (including instability, meanders, eddies, Gulf Stream rings; secondary circulation); (§4) Ekman transport, upwelling (including jets, squirts, filaments), downwelling, cascading; (§5) tides, surges, inertial currents; (§6) fronts; (§7) internal tides, internal waves; (§8) surface waves (including Stokes drift).

These processes are discussed with regard to circulation, exchange and water mass formation, according to context. The ordering of sections is intended to correspond approximately with this sequence of consequences. An alternative view with similar outcome is that as length- and time-scales decrease, we discuss advection, stirring and mixing in turn. The boundary between the explicit (of larger scale: advection and stirring) and the smaller-scale parametrised “mixing” processes will be taken as 1 km, 1 hour, very approximately. [There is no natural boundary; internal motions occur with small length scales and correspondingly high frequency and non-linearity, forming a chain leading to turbulence and ultimately molecular diffusion. Moreover, the chosen boundary may depend on the application; for global calculations, averages may need to be taken over very much larger scales than 1 km and 1 hour. For example, WALSH (1991) aggregates all eastern “upwelling”-type boundaries. HOFMANN (1991) discusses problems associated with scaling up from shelf-model detail, with site-specific aspects, to global (eg. ocean general circulation) models. The chosen boundary here may be regarded as a compromise between (i) minimal averaging, because the interaction between averaged quantities is not equal to the average of the interaction calculated from their exact values, and (ii) practicalities of calculation and measurement, which only allow limited resolution].

Section (§9) discusses effects of capes and canyons. Concluding sections inter-compare processes for the separate consequences (§10) and discuss future research needs (§11).

2. COASTAL TRAPPED WAVES

These propagate along the shelf and slope, cyclonically around the ocean at sub-inertial frequencies (HUTINANCE, MYSAK and WANG, 1986). For a straight shelf and given wave-number (corresponding to sinusoidal along-shelf form) there is a sequence of modes with decreasing frequencies corresponding to increasingly complex cross-shelf structure (ie. an increasing number
of nodes on the sloping sea floor). Typically, the lowest-mode "Kelvin wave" has the deep-ocean gravity wave speed $(gd)^{1/2} \sim 200$ m/s and simple offshore decay on the corresponding scale $(gd)^{1/2}/f \sim 2000$ km (for notation and values see §0). The next mode "1" typically has in-phase currents over the shelf, rapid off-shelf decay, and speed $O(W_s f)$ or 1 to 10 m/s for shelf widths $W_s = 10$ to 100 km. The speeds of higher modes scale similarly or as fractions of a "reduced gravity" wave speed $O(1$ m/s) associated with the stratification; all these modes decay off-shelf in the deep ocean, although some may be most energetic over the slope. (At super-inertial frequencies, some wave energy radiates away; the waves cannot be perfectly trapped.)

The waves may be generated by the full variety of mechanisms forcing currents over the continental shelf and slope, as reviewed in (eg.) HUTHNANCE, MYSAK and WANG (1986) and HUTHNANCE (1992). This includes scattering by irregularities in the shelf, and generation by oceanic eddies offshore, thereby representing eddy-shelf interaction. The in-phase currents of mode 1 on the shelf make it especially responsive to wind forcing. Unless time-scales are long compared with a frictional decay time, the waves respond to off-shelf forcing according to the closeness of match between the length and time scales and the dispersion curve of a free wave (CHAPMAN and BRINK, 1987). If near-resonant in this sense, increased motion on the shelf is expected in response to wind or oceanic pressure gradients (POWER, GRIMSHAW and MIDDLETON, 1990); alongshore gradients in either wind or pressure imply cross-slope flow.

Theory and calculation are quite well established for waves below the inertial frequency in idealised contexts (assumed uniform along-shelf with a corresponding wavenumber). Realistic stratification and depth profiles entail numerical calculation of wave forms in the cross-shelf section. Programs have been made widely available by BRINK and CHAPMAN (1985, 1987). Calculations in the presence of a mean current have been made by LUTHER and BANE (1985). NARAYANAN and WEBSTER (1987) retained primitive variables with only low-order derivatives, allowing the inclusion of mean currents in the context for the waves; this approach might facilitate the inclusion of friction.

A review of observations to 1986 (HUTHNANCE, MYSAK and WANG, 1986), supplemented in HUTHNANCE (1992), shows that coastal trapped waves occur along shelves of various orientations and all continents in both hemispheres. Mode 1 in barotropic form is most often identified. In particular, topographic waves were associated with much of the near-bottom variability in SEEP-I (CSANADY and HAMILTON, 1988; CSANADY, CHURCHILL and BUTMAN, 1988) and SEEP-II (SHAW, PIETRAFESA, FLAGG, HOUGHTON and SU, 1994).

The waves may be significant for:

- rotary near-diurnal currents near the shelf break (HUTHNANCE, MYSAK and WANG, 1986; HUTHNANCE, 1992), see §5;
- slope current stability (eg. BANE, 1980; PEDLOSKY, 1980), §3;
- their longshore propagation of a wide range of phenomena (eddies, upwelling and wind-driven flow generally, slope current, tides and surges, §§ 3 to 5);
- along-shelf evolution or adjustment of the same phenomena, notably the slope current (HUTHNANCE, 1987a), §3;
- rectified flow through non-linear effects (DENBO and ALLEN, 1983); large-amplitude cross-slope currents $u$ and small topographic length scales $L \sim h/|\nabla h|$ are necessary if the rectified flow scaling as $u|u/\nabla L|$ is to be comparable with other flows $u$ (HUTHNANCE, 1984). Example scales (§0) give $u/\nabla L = 0.1$;
- biased form drag, resisting along-shelf flow which opposes their sense of propagation and thereby favouring cyclonic flow around the ocean margin (BRINK, 1986; HAIDVOGEL and BRINK, 1986).
Thereby, coastal trapped waves underlie and control phenomena important to ocean-margin circulation, exchange and mixing, rather than making an independent contribution. Hence their magnitude is generally determined by the forcing of the "primary" phenomenon. They have been discussed here to emphasise this role, and especially their propagation of effects of forcing from one location to another along the shelf.

The waves may make a distinctive contribution to the magnitude of circulation and exchange, via (i) a near-resonant response or (ii) propagation to a shelf/slope sector of different character (e.g. narrower). (i) has been observed in the form of anti-cyclonic diurnal tidal currents O(0.1 m/s), see §5.1. Changes of shelf character (ii) result in "scattering" of the motion in various ways (Huthnance, Mysak and Wang, 1986). For motion at low frequencies (<<f) the flow is predominantly along-slope and approximately proportional to |V|hl, i.e. it speeds up as depth contours converge. (Cross-slope flow, induced by canyons and other shelf irregularities, is discussed more specifically in §9).

3. ALONG-SLOPE CURRENTS

Currents along the continental slope can and do occur with non-zero time-average and velocities exceeding or opposing those to either side in the adjacent shelf and ocean. The precise cross-shelf distribution of the flow depends upon the character of the forcing, balanced by alongshore pressure gradient, friction and acceleration (including the Coriolis force on cross-shelf flow). There are several potential forcing terms (see also Huthnance, 1992).

Freshwater runoff may cause cross-shelf flow and Coriolis forces thereon. The result tends to be a baroclinic coastal current, but for sufficient runoff and a narrow shelf the coastal current may be broad enough to include the slope; the Norwegian Coastal Current is an example. A balance between along-shelf bottom friction and the Coriolis force on the cross-shelf transport consequent on runoff suggests that bottom currents are very small (mm/s; Hepas, 1980), i.e. the flow is almost entirely baroclinic. There is no inherent cause for flows on the slope to exceed those on the shelf; a tendency for a shelf-edge frontal boundary to fresher shelf water is discussed in §6.

Alongshore pressure gradients are typically imposed by the ocean and may incorporate density forcing. In the west European context, forcing by the "Joint Effect of Baroclinicity And Relief" (JEBAR) is suggested (Huthnance, 1984). An oceanic south-north density gradient - corresponding to O(10^-7) surface slope - combines with the shelf bathymetry to drive the current. The flow speed scales as

$$\frac{1}{2} h (h_o-h) g |p| \nabla p / k$$

ie. typically 0.1 m/s (c.f. §6). Thus JEBAR provides forcing for barotropic flow \( \propto h(h_o-h) \) which is greatest over the slope in water depths \( h \) intermediate between \( h_o \) and 0. In detail, this cross-slope distribution depends on the relative influence of bottom friction and lateral diffusion, which strongly affect the associated cross-slope density field (Huthnance, 1984); see also §3.3. As well as the simple closed rectangular ocean in Huthnance (1984), a two-layer analytic model has been used to show that the Indian Ocean density gradient drives the Leeuwin Current poleward off western Australia against all but the strongest equatorward winds (Weaver and Middleton, 1990). A numerical barotropic model of the west European shelf and adjacent ocean has been integrated to simulate the slope current driven by JEBAR (Pingree and Le Cann, 1990).
Steady along-shelf wind stress forcing $\tau$ (considered alone) eventually gives a steady flow $\rho^{-1}\tau/k$ along the shelf (Huthnance, 1984) in a simple balance against bottom friction (linearised coefficient $k$). However, it should be noted that of all the distinct forms of forcing, wind stress is perhaps the one for which the slope-current response is likely to be mediated by a reactive pressure gradient. On the eastern side of the ocean, this is the upwelling or downwelling scenario (§4) as the wind and flow are equatorward or poleward. The estimated flow $\rho^{-1}\tau/k$ is typically 0.1 m/s (c.f. §6), independent of water depth.

An offshore increase in along-shelf wind stress (i.e. wind-stress curl) thus leads to stronger flow offshore. It also implies Ekman suction and vertical velocities in the interior. In turn, offshore variation in this Ekman suction implies tilting of isopycnals, and vertical shear (López-Mariscal and Clarke, 1993).

The time scale for accelerating the flow (up to the speed $\rho^{-1}\tau/k$) is $h/k$, i.e. O(1 day) for each 100m, and so depends on water depth $h$. On shorter time-scales $t$, the flow is accelerated to a speed $\rho^{-1}\tau t/h$ inversely proportional to the water depth: the shallow seas over the shelf are more responsive.

The slower response to wind stress over the slope lends importance to steady forcing but also to long-term bias in the response. Such bias occurs as “form drag” or “topographic stress” (Brink, 1986; Haidvogel and Brink, 1986; Samelson and Allen, 1987). Flow in an anticyclonic sense around the ocean and over shelf irregularities sets up stationary lee waves (coastal trapped waves propagating cyclonically relative to the flow); the associated pressure field acts as a drag on the flow. Thus poleward flow along the eastern side of the ocean is favoured (or equatorward flow along the western side). (The associated bottom Ekman layer will then be a downwelling). If such form drag is extremely effective, then a model equation for the long-shelf flow is

$$\partial v / \partial t = \rho^{-1}\tau/h - kv/h \quad (v > 0 \text{ only})$$

For $\tau = \tau_{\text{max}} \sin \sigma$, (2) is easily solved to give approximate estimates

$$\langle v \rangle = \rho^{-1}\tau_{\text{max}}/(2\pi k) \quad (\sigma << k/h)$$

$$\left(1 + (2\pi)^{-1}\rho^{-1}\tau_{\text{max}}/(\sigma h)\right) \quad (\sigma >> k/h)$$

Hence (in this model) the mean flow is greater at lower frequencies, but only up to a fraction $(2\pi)^{-1}$ of the value under a steady wind. In numerical calculations, Haidvogel and Brink (1986) find effective form drag by small-scale shelf irregularities, and mean flows of a few cm/s, principally in shallower shelf waters.

Non-linear waves may generate rectified currents (Denbo and Allen, 1983). This is more effective if the waves result from fluctuating wind-stress curl over the ocean. Typical currents are $O(0.01 \text{ m/s})$ only (for oscillatory currents $\sim 0.1 \text{ m/s}$ and topographic scale $h|\nabla h|^{-1} \sim 10 \text{ km}$); they are concentrated close to the shelf break. Rectification of tidal currents (in a context where the cross-slope transport $Q = uh$ is locally uniform) gives an along-slope current $O(h^{-3}|\nabla h|^{-2}\sigma^{-2})$ similarly concentrated close to the shelf break. [The magnitude is $O(0.01 \text{ m/s})$ for scales as in §6, i.e. small except on broad shallow shelves of high tidal range]. In similar vein, topographic Rossby waves incident from the ocean yield up their mean onshore $(x-)$ flux of longshore momentum (Garrett, 1979); the mean flow location corresponds to the extent of wave propagation inshore before substantial frictional decay. An estimate of the current is $\partial/\partial x<huv>/k$, e.g. 0.1 m/s for $h=100 \text{m}$, $k=10^{-3} \text{ m/s}$, and quite large $<uv>=10^{-2} \text{ m}^2/\text{s}^2$ if the momentum is lost over a short distance,
Moreover, random oceanic eddies or waves may cause a mean streamfunction correlated with bathymetry, i.e. cyclonic flow around the sloping ocean margin (Holloway, 1987; Zhikharev, 1994).

Other causes of an along-slope current or its (re)distribution over the cross-slope section may be: geostrophic adjustment following mixing; transfer and deposition of momentum by internal lee waves.

In general, temporally and spatially varying wind-stress acts in combination with JEBAR, waves and/or eddies with some degree of nonlinearity, and a reactive barotropic pressure gradient. The latter acts equally throughout the water depth, and is therefore most effective in deep water where it tends to give a "return flow" balancing the "forward flow" driven by the other agencies over the shallower shelf and slope.

The different acting forces elicit responses with distinctive cross-shelf distributions as outlined above. These distinctions illustrate the near-geostrophic character of the flow, allowing as a first approximation arbitrary cross-slope structure, sensitive to the form of ageostrophic forcing at the next approximation. Such sensitivity allows complexity of cross-slope structure in the presence of more than one forcing agent, and changing forms if the balance of forcing changes. Unfortunately, diagnosis through the ageostrophic along-shelf momentum balance is rarely possible for steady flows from measurements; in particular, the steady pressure gradient cannot be measured directly. Exceptionally, Atkinson, Lee, Blanton and Chandler (1983) find a balance for the southern Atlantic Bight between alongshore wind-stress, bottom-frictional drag on the current and a pressure gradient given by Sturges' (1974) steric slope; it is inferred that the latter transfers from the ocean to the shelf.

The varied response to along-shelf variations in the forcing may be propagated by coastal-trapped waves, cyclonically around the ocean by up to 90° in phase if friction is relatively weak (eg. Huthnance, 1984). The pressure gradient associated with along-shore wind of finite extent drives a "return" undercurrent; this exists also beyond the finite wind region in the cyclonic sense (Wang, 1982a).

Observations off California (Tisch, Ramp and Collins, 1992) illustrate some of these points. In repeated surveys over 18 months, the California Undercurrent was found six out of seven times, at varied depths and distances offshore. It was weaker during strong easterly winds. The current and geostrophic alongshore flow showed strong correlation with local and remote wind events, and much cross-slope structure.

Other observations of currents along continental margins bordering the eastern sides of ocean basins (Neshyba, Mooers, Smith and Barber, 1989) include those off British Columbia, the US and Mexican west coasts and Chile, the Leeuwin current off Western Australia (with average currents of 0.3 m/s and 5 Sv transport; Smith, Huyer, Godfrey and Church, 1991) and the Benguela system in the South Atlantic. Off western India, a poleward current from 10°N becomes restricted to 100 km width over the slope at 22°N with a transport estimated at 7 Sv; Shetye, Gouveia, Shenoi, Michael, Sundar, Almeida and Santanam (1991) relate it to annual cycles of alongshore pressure gradient (dominant during the north-west monsoon) and wind-stress (dominant during the south-west monsoon) and find consistency with the model of Mccreary, Shetye and Kundu (1986); Muraleedharan, Ramesh Kumar and Gangadhara Rao (1995) find the poleward undercurrent to vary with the wind-forced surface flow.

Around the eastern margin of the North Atlantic, the northward slope current is possibly continuous (in a rather undefined sense) from the Gulf of Guinea to Biscay (Barton, 1989). Measurements are numerous but the majority are of short duration. In particular, a general northward winter flow west of Iberia extends from the surface down to the level of the
Mediterranean intermediate water, and appears to be entirely continuous from the Gulf of Cadiz to Biscay. Transport appears to increase northwards off Portugal (FRONI, FUZA, AMBAR and BOYD, 1990). Then PINGREE and LE CANN (1990) note that the mean slope current near NW Spain (~ 44°N) is markedly seasonal, with warm near-surface water flow, typically 0.15 m/s, in winter according to the seasonal wind stress along the north African and Portuguese coasts. In summer, the winds cause an equatorward flow near the surface, O(0.1 m/s) off Portugal. Then northward flow along the slope appears as an undercurrent. Further observations continue around the western border of Britain (PINGREE and LE CANN, 1989; BOOTH and ELLETT, 1983; HUTHNANCE, 1986) as far as the Wyville-Thomson Ridge near 60°N. The evidence includes many current measurements, observations of a saline “core” over the upper slope at 56°-59°N west of Scotland (HILL and MITCHELSON-JACOB, 1993) and the nature of sediments on the slope there, down to ~ 600 m (KENYON, 1986). An augmented branch extends further north east to the Norwegian slope (HUTHNANCE and GOULD, 1989), as emphasised by drogue tracks (BOOTH and MELDRUM, 1987) and the sediments down to ~ 800 m (KENYON, 1986).

Typical speeds of the above are O(0.1 m/s), but the slope current around Scotland increases northwards to O(0.3 m/s) at the Wyville-Thomson Ridge and 0.5 m/s beyond. Speeds up to 1.8 m/s have been measured in the Leeuwin Current where it narrows to the south. In the ECOMARGE context, the Liguro-Provençal current in the Gulf of Lions reaches speeds up to 0.5 m/s (MILLOT, 1990).

Western boundary currents are typically O(1 m/s) and are discussed separately (§3.1).

Slope currents as described are a significant contribution to shelf edge circulation, particularly where speeds exceed about 0.1 m/s. Near the equator, there is the possibility of a direct contribution to shelf-ocean exchange, exemplified by the Amazon outflow plume turning offshore seasonally, eventually to join a current flowing across the Atlantic (MULLER-KARGER, MCCLAIN and RICHARDSON, 1988). Elsewhere, associated pressure gradients can drive flows in adjacent (marginal) seas via connecting straits; MINATO and KIMURA (1980) analysed the dependence on friction, marginal sea/ocean depth ratio and strait width. (Their analysis was actually for a western boundary current). Less directly, slope currents can be influential in controlling the water available for exchange (eg. the poleward current off western Europe is warm and saline).

Exchange is facilitated because ageostrophy allows cross-slope flow (HUTHNANCE, 1984; see also Appendix A). In particular, oceanic baroclinic flow may come over the slope (see §3.3). Equivalently, a “downwelling” bottom Ekman transport $\rho \nu \tau f$ is expected below a poleward homogeneous slope current on an eastern boundary (eg. 1 m$^2$/s, c.f. §8). VENNELL and MALANOTTE-РИZOLI (1990) found that inflows from the ocean (or coast) spread slowly across a vertically-mixed shelf sea by virtue of friction on the sloping bottom. HILL (1995) analyses the “leakage” of a slope current onto the shelf; it is enhanced by friction and by changes to a steeper slope. Some empirical estimates of overall exchange in the slope-current context are available. Apparently adjustment of baroclinic oceanic flow increases the winter slope current off Portugal from ~0.3 Sv near 38°N to ~0.6 Sv near 42°N (FRONI, FUZA, AMBAR and BOYD, 1990). For steady flow, an exchange rate can in principle be calculated from

\[
\text{slope current transport} \times \text{salinity decrease} = \frac{\text{salinity excess in current} \times \text{longshore distance}}{4}
\]

(A corollary is that salination of the shelf water occurs owing to the presence of the slope current). Figures in DANIAULT, MAZ and ARHAN (1994) for Mediterranean water along the Iberian slope suggest a salinity excess ~0.4 and a decrease ~0.4 in 700 km so that the exchange is (slope current...
transport) / 700 km. HILL and MITCHELSON-JACOB (1993) showed that the “core” salinity decreased northwards along the Hebridean shelf, consistent with mixing with the surrounding fresher water. However (personal communication) they found too much temporal and along-slope variability in their 200 km Scottish sector, for the estimate (4) to be reliable. From increasing transports along the upper slope (above 1000 m) from Biscay to Scotland, viz. 0.6 Sv at 47°N, 1 Sv at 52°N and 2 Sv at 58°N (PINGREE and LE CANN, 1989) the entrainment of oceanic water may be inferred; over 1000 km it is roughly comparable with the slope current transport. A similar estimate comes from drogued-buoy tracks over the Scottish slopes (BOOTH, 1988a); after about 1500 km, four out of ten remained over the slope (four had come on to the shelf and two went off-shelf). In an alternative approach, BOOTH (1988b) estimated across-slope diffusivity K from the buoys’ dispersion: about 700m²s⁻¹ along the eastern side of the Faroe-Shetland Channel (on scales ~50 km) and about 500m²s⁻¹ in the Norwegian Trench (on scales ~40 km). If we equate alternative expressions for the rate of exchange of constituent C in the x-direction across the slope,

\[ u \Delta C = K \Delta C / \Delta x \]  

(5)

the volume rate of exchange uh in depth h (500 m) is thus Kh/Δx or 6 to 7 m²s⁻¹ in this case. As the along-slope transport has been measured at about 7 Sv along the Faroe-Shetland Channel (HUTHNANCE and GOULD, 1989) this estimate of exchange is again consistent at about

(slope current transport) per 1000 km

(6)

With respect to the exchange of bottom sediment rather than constituents within the water column, KENYON (1987) points out that a strong along-slope current prevents sedimentation, except in deeper areas which are preferentially filled to inhibit slope failure. As well as potentially introducing a different water mass from “upstream”, the bottom boundary Ekman layer associated with a slope current can also affect mixing over the slope. WHITE (1994) reports changes of stratification in the bottom 100 m, associated with changes in the mean slope current and consistent with the effect of Ekman flow up/down the slope. TROWBRIDGE and LENTZ (1991) analysed and solved numerically for the bottom boundary layer, which is thicker (thinner) for slope currents favouring downwelling (upwelling).

3.1 Western boundary currents

Apart from the Antarctic Circumpolar Current, these form the largest confined transports in the ocean, typically tens of Sverdrups, and are therefore a principal contributor to circulation wherever they occur. They are well-known as part of the ocean circulation, whether wind-driven, thermohaline or inertial. It is beyond our scope here to provide a thorough review, but the oceanic origin appears to make a clear distinction with slope currents as discussed hitherto. Estimates of shelf-ocean exchange adjacent to these currents (mostly discussed in §3.4) are not in the range of tens of Sverdrups per 1000 km. The slope and shelf are not implicated in the generation of these currents, which appear to be subject to strong geostrophic constraint against exchange across the slope.

Nevertheless, consequent exchanges which are small relative to such strong currents may still be significant. HOFMANN, PIETRAFESA and ATKINSON (1981) point out that an offshore displacement of the Gulf Stream off North Carolina draws cold water to the shelf break (by continuity in the cross-shelf section) facilitating upwelling onto the shelf.
GAWARKIEWICZ, CHURCH, LUTHER, FRIEDMAN and CARUSO (1992) document an intrusion, estimated at 12 km³, over the shelf north of Cape Hatteras in the summer of 1990, possibly aided by seasonal stratification reducing topographic influence and lateral density gradients.

Several papers have discussed the interaction between the Kuroshio and the rather complex marginal seas from Taiwan to Japan. A bend in the margin results in the Kuroshio approaching the shallower marginal sea. In winter, Kuroshio surface waters intrude on to the shelf: in summer, Kuroshio waters are slightly more dense than shelf surface waters; principally subsurface intrusion occurs (SU and PAN, 1987; CHEN, RUO, PAI, LIU and WONG, 1994). A simple step-model by HSUEH, WANG and CHERN (1992) suggests that a fraction \( \frac{h}{h_0} > \frac{2}{3} \) of the Kuroshio transport enters the marginal sea of depth \( h < h_0 \) (the Kuroshio depth above a bottom layer assumed passive). Other conclusions from barotropic models (QIU and IMASATO, 1990; SU and PAN, 1987; SU, PAN and LIANG, 1994) are: the intrusion to shallower water is anticyclonic (by conservation of potential vorticity) and affected by both planetary and topographic “beta” (gradients of rest-state vorticity). Topographic beta concentrates some onto-shelf flow against Taiwan and planetary beta facilitates anticyclonic flow on the shelf as part of the intrusion. With regard to the reduced fraction of Kuroshio transport entering the shelf sea in summer, HSUEH, CHERN and WANG (1993) provide an interpretation regarding the Kuroshio as an approaching bottom current blocked by a step up to the shelf; additionally, much of the shelf is occupied by a “pool” of subsurface Kuroshio water remaining from the winter incursion. CHAO (1991) also found with a numerical model that stratification over the shelf reduced the Kuroshio intrusion.

3.2 Relation to ocean circulation; boundary current separation

SPITZ and NOF (1991) discuss a coastal boundary current encountering a step-down scarp inclined to the coast. The current separates from the coast to flow along the scarp, accompanied by water drawn from coastal waters on the deeper side, i.e. there is coastal convergence and offshore flow along both sides of the scarp. However, in the case of a scarp normal to the coast and supporting double-Kelvin waves propagating onshore, GILL, DAVEY, JOHNSON and LINDEN (1986) and JOHNSON and DAVEY (1990) find that barotropic Rossby adjustment results in an alongshore current which becomes confined to a boundary layer where the scarp meets the coast. It appears that the SPITZ and NOF (1991) solution must be reached from a different initial condition, effectively controlled from far downstream along the scarp. If the scarp supports double-Kelvin waves propagating offshore, then GILL, DAVEY, JOHNSON and LINDEN (1986) find that an alongshore current is diverted offshore along the scarp to a propagating double-Kelvin wave front and back; this is in agreement with SPITZ and NOF (1991). [JOHNSON and DAVEY (1990) also find fluid interchange across the scarp as eddies propagate along it].

Convergence of opposing currents along a boundary causes separation in the form of offshore flow. AGRA and NOF (1993) use 1½- and 2½-layer models to discuss the speed, width and angle of this merged offshore flow constrained by the balance of momentum etc. MATANO (1993) finds by numerical experiment that the southward Brazil Current separates further north than it would otherwise do because of the Malvinas Current coming from the south. This finding is reinforced by the observed correlation of separation latitude with winds, both seasonally and for anomalous winds south of the confluence (GARZOLI and GIULIVI, 1994)

Sufficient coastal convexity may cause a boundary current to separate, for example: at a corner about which the ocean occupies a sector \( > \frac{5\pi}{4} \) (STERN and WHITEHEAD, 1990) somewhat dependent on the distribution of vorticity across the current; at a convex cape of such small radius of curvature that centrifugal force shoals upper layer depth to zero at the coast (KLINGER, 1994,
considering a model with just one - upper - active layer): see also §9. After separation, a “slope sea gyre” circulation is expected between the jet and the coast (OU, 1994) with entrainment just downstream from separation if the coast is concave. Indeed, in the region of Cape Hatteras, the Gulf Stream is observed to separate from the shelf, accompanied on its inshore side by 0.1 to 0.5 Sv of “Ford water” from the shelf (LILIBRIDGE, HITCHCOCK, ROSSBY, LESSARD, MORK and GOLMEN, 1990).

At about 18°S off Africa, the Angola-Benguela front separates a cyclonic gyre in the tropical south Atlantic from the Benguela upwelling system (GORDON and BOSLEY, 1991). Surface Ekman transport and geostrophic flow from the latter appear to leave the African coast in the region of the front, amounting locally to a substantial offshore near-surface transport. However, the Ekman currents dominate, and their contribution to offshore transport is estimated as τ/ρf in accord with §4. Any divergence and offshore component of along-shelf geostrophic flow is accompanied by alongshore pressure gradients and “adjustment” as discussed in §3.3.

As mentioned in the introduction, the wind-related North Brazil Current retroreflection to join the North Equatorial Countercurrent flowing across the Atlantic, doubtless facilitated by the low latitude (≈7°N) and associated relative weakness of geostrophy. Oscillations of period 40-60 days in this area have an eddy character with currents ½ to 1 m/s over the upper slope and a substantial cross-slope component at times (JOHNS, LEE, SCHOTT, ZANTOPP and EVANS, 1990). Surface drifters have shown some onshore displacements during January to June and some offshore displacements in July to December (RICHARDSON, HUFFORD and LEMBURNER, 1994) which correlates with the period June to January when the Amazon plume turns offshore (MULLER-KARGER, MCCLAIN and RICHARDSON, 1988).

3.3 Relation to ocean circulation; secondary circulation

Any poleward shoaling of isopycnals in the ocean implies an eastward geostrophic shear flow, right up to the shelf edge or until ageostrophy provides some accommodation of the flow. For example, at a wall with friction and diffusivity, baroclinic shear associated with a poleward increase in density may be closed by downwelling accompanied by a poleward surface current and equatorward undercurrent (IKEDA, 1986; McCREARY, SHETYE and KUNDU, 1986). With no friction or diffusivity, the baroclinic shear would have to be turned inertially to an intensifying coastal jet. Over a slope, a slope current can intensify without invoking inertia. Indeed, a slope current can accommodate depth-integrated transport from the ocean interior (Fig. 2) while adjusting along-slope (eg. from zero at a closed zonal boundary) to a non-zero “equilibrium” value.

The adjustment is effected by coastal-trapped waves as discussed in HUTHNANCE (1992). The length-scale of adjustment is the waves’ propagation distance before decay. This decay scale, and cross-slope flow within it, can be seen in several model calculations, for example: WILLMOTT and COLLINGS’ (1994) “shadow zone” of finite length after a sudden broadening of the shelf; GREATBATCH, PAL and REN (1994) with redistribution of shelf input to the slope of an idealised eastern Canada shelf; WANG (1982b) concerning “insulation” of the shelf from oceanic pressure gradients. Stratification and surface-intensified oceanic pressure gradients may reduce the insulating effect (KELLY and CHAPMAN, 1988) and facilitate cross-slope flow. HUTHNANCE (1992) suggested that “insulation” (inhibiting cross-slope flow) is provided by depth contrast between the shelf and the ocean flow.

The “equilibrium” slope current is limited by friction which balances the forcing. In equilibrium, eastward oceanic baroclinic shear flow can be accommodated, eg. in JEBAR-driven flow (HUTHNANCE, 1984) via the Ekman transport in the frictional boundary layer below the slope.
Figure 2. Schematic cross-slope section. Streamlines of cross-slope exchange (→) match the baroclinic shear (→) associated with the oceanic density field and form the Ekman transport under an along-slope current ($\vec{V}$; velocity contours ---). However, solutions allowing for buoyancy forces in this section are the subject of continuing research ($\S$ 3.3).

current. There is observational evidence for this in the Leeuwin Current (Smith, Huyer, Godfrey and Church, 1991). However, this solution for “secondary” cross-slope flow lacks full account of stratification, and the meridional variation of Coriolis parameter, which may affect its stability. More detailed local solutions for the bottom boundary layer over the slope have been found by Thorpe (1987b), Garrett (1990, 1991), MacCready and Rhines (1993), Garrett, MacCready and Rhines (1993) and Imberger and Ivey (1993). These solutions typically scale for transport as $K_v h_q$ where $K_v$ is the vertical eddy diffusivity (Salmon and Phillips, 1992); balance boundary mixing against the tendency of buoyancy to restore stratification; determine up/down-slope flow in the boundary layer (zero integral through the boundary layer if external diffusivity is zero); tend to give a downwelling-favourable along-slope current outside the boundary layer. Trowbridge and Lentz (1991) also discuss the latter asymmetry: the downslope Ekman transport is thicker and takes longer to be “arrested” by buoyancy. The inhibition of bottom Ekman transport by buoyancy (MacCready and Rhines, 1993) has yet to be reconciled with external factors that may also determine secondary flow (notably alongshore pressure gradient representing sustained forcing, and alongshore flow divergence and time-dependence which may relax constraints on the cross-slope flow). On the broader scale of the cross-slope section outside the boundary layer, the structure of the slope current and the resulting density fields are sensitive to the boundary-layer character: friction, density diffusivities and their ratio (Huthnance, 1984). Holloway (1987) suggests that topographic irregularities of scale $\Delta h$ favour a secondary near-bed upwelling transport $O(\nu \Delta h)$ in a slope current of strength $\nu$ (eg. $1 \text{ m}^2\text{s}^{-1}$, c.f. §0).

Accommodation of oceanic interior baroclinic shear flow by the equilibrium slope current requires zero eastward depth-integral transport. Thus the shelf and slope may determine the normal depth-integral transport.

Appropriate eastern boundary conditions remain an open question for modelled ocean circulation adjacent to a shelf. In an ocean model (Pedlosky, 1983), an eastern boundary condition of zero normal flux in each layer (as opposed to zero sum over all layers) is a severe restriction which qualitatively affects the computed ocean circulation, especially the ventilated layers. For continuous stratification, Huang (1989) considered the effect of imposing merely the depth-
integrated constraint, allowing meridional density gradients and a large area of ventilated thermocline in the southeast but implicitly assuming vertical exchange at the eastern boundary. The problem is not well-posed for the ocean region separately from the coastal region; the vertical structure of the zonal flow becomes part of the whole oceanic solution, of which many are possible. These calculations, together with Hodnett and Ansari (1994) and the review of Huang (1991), suggest that a determined depth-integral normal transport may also be a more appropriate boundary condition for an ocean model than specifying the normal flow at each depth. In turn, the strong influence of the boundary condition on ocean model results affects calculated flows at the ocean margin.

In the eastern North Atlantic (e.g. Arhan, Colin de Verdière and Méney, 1994), poleward shoaling of isopycnals implies an eastward geostrophic shear flow (of order 10 Sv in 20° to 60°N) as also suggested by drogued buoy tracks west of northern Europe (Pingree, 1993). The known presence of a slope current and associated transport along the continental slope of NW Europe (for example) also suggests a non-zero normal transport.

The oceanic baroclinic shear is \( \frac{g s}{h_0 f} \) integrating to an estimated transport

\[ gh_0 s/8f \]

(eastward in depth \( \frac{1}{2}h_0 \), above an equal westward transport), eg. 1.25 m²s⁻¹ for values as in §0 and steric slope \( s = h_0 \rho^2 \nabla \rho \perp = 10^3 \) (eg. Storch, 1974) from the density gradient \( \nabla \rho \).

3.4 Instability, meanders, eddies, Gulf Stream rings

Instability is identified (for example) by Blanton (1991) as a possible contributor to features (eddies, swirls, filaments) on the Norwegian Coastal Current, California Current system, Yangtze plume and the inshore side of the Gulf Stream; by Sur, Özsoy and Ünlüata (1994) for the Black sea “rim” current along the North Anatolia coast triggered by shelf irregularities, notably Sakarya Canyon. Booth (1988a) also invokes slope-current instability as a possible source of the “Porcupine” eddy off north-west Ireland [citing baroclinic instability from theory without bathymetry, and Mysak and Schott (1977) for upper-layer along-slope flow]. Killworth and Stern (1982) find instability of a current of less dense water along a coastal “wall”; Griffths, Killworth and Stern (1982) find instability of an along-slope current of more dense water, which is expected to result in a “street” (double row) of eddies (Nof, 1990). In both cases, instability occurs under quite a wide range of circumstances; an extremum of potential vorticity is not needed, as confirmed by a rather general analysis for two layers over small slopes by Swaters (1993a). Swaters (1993b) found some sufficient conditions for stability of a dense along-slope current under finite depth, and also for near-surface currents (Swaters, 1993a) for which the slope is generally a stabilising factor. Models numerical in the cross-slope section, but harmonic in time and with a complex alongshelf wavenumber, have been used to investigate stability of the Gulf Stream off the South Atlantic Bight. Luther and Bane (1985) found an 8-day mode corresponding to observed meanders downstream of the “Charleston bump”. Oey and Chen (1991) and Oey, Ezé, Mellor and Chen (1992) found that such a “bump” increased variability (ie. cross-slope displacements of the Gulf Stream); downstream as small-amplitude propagating frontal waves and developing meanders; upstream through quasi-stationary meanders manifesting topographic waves. Xue and Mellor (1993) found that the slope slows the growth and speeds propagation of the unstable wave, to reduce growth in the Bight.

Typically, instabilities will be manifested as developing meanders in a current, which may further
evolve into distinct "filaments" of characterised water, locally strengthened currents or circulatory "eddies", possibly becoming detached from the source current. Such eddies on the north-west side of the Gulf Stream, and enclosing warm water from the south-east side, are specially termed "warm-core rings". Such successive regimes are exemplified in a single (lower) active layer quasi-geostrophic model of VIERA and GRIMSHAW (1994); the coastal current of different uniform potential vorticity is forced by topography of increasing amplitude to give in turn: small-amplitude wave-like disturbances, breaking of the primary trapped disturbance to give filaments, breaking of a secondary moving disturbance to give filaments, and winding of a primary filament around the topographic feature to give an eddy.

Both meanders (CSANADY, 1988) and eddies (LOUIS and SMITH, 1982; CHAPMAN and BRINK, 1987; GRIMSHAW, BROUTMAN, HE and SUN, 1994) have been suggested as sources of topographic waves on the continental slope. The strong slope was supposed to linearise the impinging eddy, but SHAW and DIVAKAR (1991) have applied a numerical model with non-linearity via density advection to find that density displacements over the sloping bottom were more effective than the eddy vorticity in generating topographic waves. CHAPMAN and BRINK (1987) find that the eddy is squashed on the ocean side of the shelf break, where a counter jet may be set up. WANG (1992) estimated the volume exchange associated with barotropic eddy-slope interaction; the result appears to scale roughly as

$$h_{CL}^2 \div (f\Delta h) \quad (8)$$

eg. $10^3 \text{ km}^3$ or 1 Sv for ~12 days (c.f. §0) for an eddy circulation $@ = 10^3 \text{ m}^2 \text{s}^{-1}$ corresponding to azimuthal velocity 0.5 m/s at radius 32 km approximately. Calculations were also carried out by BIDLOT and STERN (1994) for cross-scarp exchange when a sheared boundary current is rendered unstable by the continental slope and forms a sequence of eddies; the result appears to scale roughly as

$$\text{shelf depth} \times \text{breadth of shear over the shelf} \quad (9)$$

for each occurrence of instability, eg. 1 Sv for 1000 km if the boundary current is over the outer 10 km of shelf and (at 1 m/s) develops the instability in 1000 km or 12 days. These barotropic results should be regarded with caution; as already mentioned, stratification and eddy depth affect the influence of the shelf on the eddy and cross-slope transports.

Eddies have been associated with cross-slope flow in several locations. Drogued buoys provided evidence of the anticyclonic Porcupine eddy off north-west Ireland, somewhat offshore from the slope current (BOOTH, 1988a) and satellite infra-red images suggested an eddy pair at the end of a filament originating near the slope. PINGREE and LE CANN (1992a,b) describe anticyclonic warm slope-water eddies (SWODDIES) in the Bay of Biscay, formed in winter if the warm flow along northern Spain is strong (PINGREE, 1994). F90a was shed, apparently through instability of the slope current, with an estimated volume 400 km$^3$, and persisted for a year. Retention of slope-water characteristics for so long suggests that (for example) phytoplankton populations could evolve within the eddy differently from the surrounding ocean. Retention of a drogue within the rotating core for 200 days, while the eddy roamed 500 km (PINGREE, 1994), reinforces this view. However, such eddy shedding is not thought to occur every year; more often, an eddy is stationary in the south-east corner of the Bay of Biscay. Off Portugal, PINGREE and LE CANN (1993) present infra-red imagery with evidence of eddies shed from the slope at Cape St. Vincent, Setubal canyon and Lisbon canyon; additionally temperature, salinity and transmission characteristics of a shallow
off-shelf eddy at 12°-12½°W suggested an origin on the slope in the Gulf of Cadiz. West of Australia, satellite images have shown meanders in the Leeuwin current diverting its transport far offshore from the shelf break (PEARCE and GRIFFITHS, 1991). CRESSWELL and PETERSON (1993) record an offshoot from the Leeuwin Current south of Western Australia, of estimated volume 1000 km³ equivalent to about 1 Sv transport for 12 days. An anticyclonic eddy (about 90 km alongshore by 50 km cross-shelf scale) against the North California slope is described in WASHBURN, SWENSON, LARGIER, KOSRO and RAMP (1993). Currents up to 0.5 m/s, but more typically transporting suspended sediments in a plume through 80 km in 9 days, give an estimated sediment transport 10³ t/day. (This could remove most of the riverine sediments reaching the outer shelf here). The offshelf transport of water, assumed 200 m deep, is about 0.2 Sv.

[Oceanic eddies are ubiquitous, and exchanges as above will occur, whether or not there is a boundary current locally. For example, there has been well-documented evidence in the Bay of Biscay bordering Iberia and the Celtic Sea: satellite images in DICKSON, GURBUTT and PILLAI (1980); the Turbillon experiment described in HARVEY and GLYNN (1985)].

Much attention has focused on meanders and eddies associated with the Gulf Stream off the South Atlantic Bight. BROOKS and BANE (1983) found alongslope currents off North Carolina to vary between 0.5 m/s southward and 1 m/s northward owing to meanders. BRINK (1987), LEE, YODER and ATKINSON (1991) and XUE and MELLOR (1993) give brief reviews. Scales are roughly 10-25 km cross-shelf, 130 km along-shelf, with northward propagation, acceleration past Cape Hatteras and occurrence at intervals of roughly four to six days (GLEEN and EBBESMEYER, 1994a,b). LEE and ATKINSON (1983) and LEE and PIETRAFESA (1987) show the Gulf Stream influence on currents at the shelf break (75 m depth), with reverse flow only when the Gulf Stream influence is weak (ie. offshelf). Cold cyclonic perturbations (upwelled North Atlantic Central Water) propagate to the north and are not apparently correlated with the wind or surface elevation, in contrast with wind-related flow on the shelf (40 m water depth). Such Gulf-Stream influence on the shelf appears to increase when the Stream is displaced off-shelf (SAVIDGE, BLANTON, LEE and EVANS, 1992) and there is also increased scope for the cold-core eddies and warm filaments extending cyclonically around them onto the shelf. LEE and ATKINSON (1983) estimate an upwelling rate w = 10⁻⁴ m/s associated with the eddies; then

\[\text{exchange of water} = w \times \text{cross-shelf scale} \times 20 \text{ km} \times \text{half-time influence} = 1 \text{ m}^2\text{s}^{-1}\]

However, much of this exchange is confined to the outer shelf; ATKINSON, LEE, BLANTON and CHANDLER (1983) estimate that only 20% of the (inner) shelf water is exchanged each month, ie. 2 km² cross-section x 0.2/m = 0.15 m²s⁻¹. The associated nitrogen input to the shelf dominates the input from rivers and (at an estimated 1.3 Mt/year for the whole shelf sector) exceeds the requirements for observed primary production on the shelf (LEE, YODER and ATKINSON, 1991). ATKINSON, OKA, WU, BERGER, BLANTON and LEE (1989) show that winter variability of heat content on the shelf is caused by intrusion of Gulf Stream water, in meanders or via Ekman transport ($\S$4).

Further north, warm core rings (WCRs) are larger features but the slope appears to limit their impact on the shelf, in accord with laboratory experiments; CARNEVALE, KLOOSTERZIEL and VAN HEIJST (1991) also reviewed literature showing that anti-cyclones tend to propagate to the “south-west” in the context of beta or its large-scale topographic equivalent; the slope thus imparts an offshelf component to the propagation of WCRs. Thus CHAPMAN, BARTH, BEARDSLEY and FAIRBANKS (1986) find that very little offshore water, only 30% of the along-shelf flow, comes onto the Middle Atlantic Bight shelf over a distance of 1000 km. BEARDSLEY, CHAPMAN, BRINK, RAMP and SCHLITZ (1985) found that WCR currents did not show in records from a mooring in 104 m
off Nantucket Shoals; high salinity appeared at a mooring in 200 m for one of three rings in summer, although currents exceeded 0.4 m/s there for all three rings. In winter, with more wind forcing, the influence of rings was less clear.

The contribution of WCRs to shelf/slope water exchange has been reviewed by Joyce (1991) referring to some of the following studies. There is evidence that WCRs cause "streamers" of water to leave the shelf. For WCR82B, Joyce, Bishop and Brown (1992) found such a flow with off-shelf speed 0.1-0.5 m/s and shelf water down to 100 m, and estimated a transport of 0.8-0.9 Sv (for water of salinity S < 35 psu.) or 0.38 Sv (for water with S < 34 psu.). The streamer was not wrapped around within the ring circulation, but advected to the inshore edge of the Gulf Stream. For a rather weaker anticyclonic eddy between Cape Hatteras and Hudson Canyon, with diameter ~100 km and currents ≤ 0.3 m/s, Churchill, Levine, Connors and Cornillon (1993) estimated the off-shelf transport as 0.13 Sv (within a factor 2). In 1979, there were 12 WCRs off the NE USA (Beardsley, Chapman, Brink, Ramp and Schlitz, 1985) but there is large interannual variability in their occurrence. For the sector 60°-70°W (about 800 km) Smith (1978) estimates an average exchange 10^4 km^3 per year ~0.3 Sv on the basis of six events per year with typical entrainment of an area 50,000 km^2 with a depth of 30-50 m. Garfield and Evans (1987) estimate a similar average off-shelf streamer transport, 0.18 Sv from Cape Cod to Cape Sable (about 400 km) on the basis of 0.25 Sv average per streamer and 70% occurrence (from satellite infra-red imagery). However, this average implies a comparable overall loss when a water budget is attempted for the Gulf of Maine, emphasising the uncertainty in transport estimates. The oxygen-18 evidence already referred to (Chapman, Barth, Beardsley and Fairbanks, 1986) also suggests that cross-shelf transports are much reduced from these cross-slope figures. Further south, Churchill and Cornillon (1991) describe the occurrence of water originating from the Gulf Stream, sometimes over the upper slope and occasionally on the shelf north of Cape Hatteras. An associated along-slope density gradient (less-dense Gulf Stream water in the south) leads to a rough geostrophic estimate 0.1 Sv for export from the 250 km of shelf just north of Cape Hatteras. Biscaye, Flagg and Falkowski (1994) estimate about 0.2 Sv outflow across 400 km north of Cape Hatteras, and Mountain (1991) estimates about 0.3 Sv outflow across ~800 km shelf edge from the Middle Atlantic Bight (Nantucket Shoal to Cape Hatteras) on the basis of an overall budget of shelf water.

In the case of WCR82B, Bishop and Joyce (1986) found that the concentration of suspended particulate matter in the "entrainment zone" exceeded that of slope water or within the ring, and estimated a net offshelf flux O(10^4/day).

Vidal, Vidal and Pérez-Molero (1992) note an unquantified amount of ocean-shelf exchange when a Loop Current anticyclonic ring collided against the shelf in the Gulf of Mexico. In deeper water O(100 km) from the shelf break, on/offshore transports between anticyclonic and cyclonic eddies are very large in this context; 18 Sv (Vidal, Vidal, Hernández, Meza and Pérez-Molero, 1994) or even 30 Sv (Biggs and Müller-Karger, 1994), but only a small fraction is thought to be exchanged with the shelf. In the latter case, one might estimate an upper bound 2.5 Sv from

0.5 m/s (speed) x 50 km (breadth) x 100 m (maximum likely depth).

Sturges (1994) reports 22 such rings separated from the Loop Current in 20 years.

"Coastal" water upwelled in the Benguela system has been found "wrapped around" an off-shelf Agulhas ring (Duncombe Rae, Shillington, Agenbag, Taunton-Clark and Grundlingh, 1992). They estimate an aggregate volume 5000 km^3. With about nine rings per year (Lutjeharms...
and VAN BALLEGOOYEN, 1988) about 60% are needed to entrain the estimated 0.91 Sv accumulated upwelled water (GORDON, BOSLEY and AIKMAN, 1995); water mass analysis suggests that such rings may be the primary mechanism for removal. LUTJEHARMS, SHILLINGTON and DUNCOMBE RAE (1991) made in situ measurements in one such filament, extending more than 1000 km offshore, estimated about 1.5 Sv transport and that satellite images showed at least 11 such filaments (extending to more than 500 km offshore) in 1978-1984.

In general we lack parametrisations for the exchange contribution of all these features as they develop to non-linear forms. Entrainment into eddies’ persistent rotary circulation may be particularly effective in lengthening the shelf-ocean water interface; then diffusive processes have increasing scope to cause eventual mixing. The stronger currents associated with them also encourage mixing and boundary turbulence.

4. EKMAN TRANSPORT AND UPWELLING

Following offshore transport of upper waters, upwelling of deeper ocean waters onshore to the shelf is expected, to maintain continuity in the cross-shelf section. For example, northerly winds along western European shelves drive a surface (Ekman) transport offshore. The magnitude is $1.5 \times 10^{-3}$ m²s⁻¹ (c.f. §9). Such upwelling-favourable winds are equatorward on an eastern oceanic boundary, and poleward at the western side of the ocean. They may also be expected to accelerate an alongshelf current; if this reaches an equilibrium balanced by bottom friction, then upwelling might be expected in the form of onshore Ekman transport $p \times 10^{-3}$ in the bottom boundary layer below the current. More generally, however, this equilibrium is not reached; acceleration, stratification and along-shore pressure gradients also distribute the stress and upwelling through the water column. Other features associated with forcing of finite extent (briefly reviewed in HUTHNANCE, 1981) include an opposing undercurrent (poleward on the eastern side of the ocean) and propagation of upwelling by coastal trapped waves, cyclonically around the ocean. Many upwelling regions in fact have a poleward current along the slope (for whatever reason, §3) and so a bottom Ekman layer with more offshore transport; then the interior should carry all onshore transport.

Evidently, the vertical distribution of onshore upwelling flow is controlled by the shelf depth profile. If the shelf is shallow, or near the equator so that the Ekman layer is relatively thick, then the surface flow offshore and any bottom Ekman layer transport may occupy most of the water column; the flow is frictionally determined. Supposing that the Ekman layers are relatively thin, the additional influence of stratification is considered in Appendix B; here the upwelling replaces offshore surface Ekman transport in a stratified sea against a vertical coastal wall. If $f$ is the Coriolis frequency and $N$ is the buoyancy frequency, then there is an inherent ratio $N/f$ of horizontal and vertical scales. On a flat shelf of depth $H$, sufficiently far (> $HN/f$) from the coast, then the inflow is uniform through depth without a vertical component. An upper-layer concentration of stratification ($N_1H_1 > N_2H_2$ for layer 1 above layer 2 and depths $H_1$, $H_2$) gives a majority of upwelling inflow in the upper layer within $O(H_1N_1/f)$ of the coast. $HN/f$ is also the scale for distance from the coast within which vertical velocity decays significantly over a flat bottom.

In time, these distributions will change as diffusive effects become important. DALU and PIELKE (1990) studied time-dependent upwelling, particularly in relation to the time-dependence of the forcing. Upwelling and mixing themselves reduce the coastal layer depth and density stratification ($H_1N_1$) and hence the offshore scale (DE SZÖKE and RICHMAN, 1984); upwelling causes the pycnocline to surface as a correspondingly narrow front which is advected offshore. In the presence of wind mixing and surface heating, DE SZÖKE and RICHMAN (1981) found a tendency to a steady
state in which cool upwelling was balanced by the input and mixing down of heat.

Modelling the CODE 2 (northern California) upwelling, CHEN and WANG (1990) find deepening at the upwelling front, hence offshore convergence and then divergence in the near-surface Ekman transport. A two-cell pattern results. A two-cell pattern also occurs in the Benguela system (BARANGE and PILLAR, 1992).

A shelf break distinct from the coast also allows more complex patterns. Divergence in the upper-layer Ekman transport may induce shelf-break upwelling (HUTHNANCE, 1981). JOHNSON and ROCKLIFF (1986) review several upwelling models where a change of slope at the shelf break induces specific “secondary” upwelling there (Fig. 3), one reason being enhancement by lateral viscosity of up-slope transport. There is also a tendency for offshore migration of the “front” of upwelled water at the coast, as upwelling continues, to be limited to the shelf, stopping at any shelf break.

Figure 3. Cross section showing diagrammatically the circulation streamlines over the shelf break. The shear layer thicknesses are in terms of the Ekman number $E$. From JOHNSON & ROCKLIFF (1986).

BAKUN (1990) infers a climate trend increasing the wind-forcing of such upwelling. On the other hand, ROEMMICH and McGOWAN (1995) attribute a decline of zooplankton in the California Current to climatic warming (hence increased density contrast across the thermocline so that upwelling is from shallower water bringing less inorganic nutrient for new production).

Cyclonic winds (giving a wind-stress curl) drive a surface Ekman layer divergence and hence upwelling of the water below. The intensity of upwelling is inversely proportional to the length scale over which divergence occurs; as the scale of the wind field is generally greater than the shelf width, topographically-related upwelling tends to be the more intense, but upwelling driven by wind-stress curl may be correspondingly more extensive (DALU and PIETKE, 1990). Locally strong wind-stress curl may be caused by land (e.g. coastal mountains) distorting the wind.

Upwelling may also occur beneath an along-slope current, e.g. a western boundary current. The onshore Ekman transport $p^{-1}t/f$ in a bottom boundary layer referred to above, where $t$ is the bottom stress on homogeneous flow, is irrespective of the original driving mechanism for the flow. Furthermore, on the cyclonic side of fast flow with low Rossby number (e.g. the west side of the Gulf Stream), JANOWITZ and PIETRAFESA (1982) found theoretically that isobath divergence
should induce upwelling and associated onshore flow. The order of magnitude of the transport is
$2h_0^2V_w(\nabla h)$, x divergence, eg. 20 $m^2s^{-1}$ for a cross-stream (x-) scale 100 km determining the
vorticity $\nabla V_w$ and depth gradient $h_x$ divergence 0.1 radians across 100 km and other values as
in §0. This is a large figure (2 Sv/100 km) but should be compared with the along-stream transport
0(100 Sv) with this scaling, and the 5% cross-axis flow (of order 5 Sv per 100 km) implied by the
assumed divergence. Li, WIMBUS, WATTS and LEE (1986) show also that flow through a
narrowing strait (eg. the Gulf Stream through the Florida Strait) leads to increased tilt and shoaling
of the flow’s base (another form of upwelling on the cyclonic side). Similarly, GILL and SCHUMANN
(1979) considered an upper-layer inertial current (poleward at the western oceanic boundary)
adjacent to a sloping continental shelf of varying width; at a minimum of shelf width, the current
could pass smoothly from sub-critical to super-critical, involving a substantial fraction of the
transport crossing the shelf edge and an outcrop of cold lower-layer water downstream of the
constriction. SUR, ÖZSOY and ÜNLÜGÜ (1994) invoke a widening shelf for upwelling near Cape
Kerempe (north Anatolia, Black Sea) where the “rim” current separates from the coast. In the
context of two-layer stratification against the continental shelf, CAI and LENNON (1993) find that
upwelling is inhibited by an along-slope bottom-layer current in the sense of an upwelling-forcing
wind, and by an opposing upper-layer current. In this case, the upwelling at the slope is inversely
related to distance from the coast.

Upwelling may also be an adjunct of ocean circulation as described in §3.3.

Upwelling occurs at many locations world-wide where winds drive a surface Ekman transport
offshore, as reviewed by HUTHNANCE (1981; briefly), JOHNSON and ROCKLIFF (1986) and others.
The eastern margins of oceans in the tropics and sub-tropics are particularly susceptible, owing to
prevailing winds with equatorward and westward components. NW Africa is a particularly well-
known example (MITTELSTAEDT, 1991) as are the eastern margins of the Pacific. In the north, the
latter has summer upwelling from Vancouver Island - where the regime includes a shelf-break
current, reviewed by CRAWFORD and THOMSON (1991) - to southern California (ATKINSON,
BRINK, DAVIS, JONES, PALUSZKIEWICZ and STUART, 1986, describe upwelling around Point
Conception) and Baja California (eg. WANG and WALSH, 1976). HICKEY (1992) reviews
circulation around the Santa Monica-San Pedro Basin off Southern California, including upwelling
events in winter. Upwelling off the west coast of India from 9° to 15°-20°N is described by SHETYE,
GOUVEIA, SHENOI, SUNDAR, MICHAEL, ALMEIDA and SANTANAM (1990) including equatorward
surface flow increasing southwards to ~ 4 Sv, and a poleward undercurrent with associated
downwelling at the bottom. SHANNON (1985) reviews the Benguela system extending from about
15°S to Agulhas Bank, and describes an associated shelf-break front, jets and a poleward
undercurrent against the slope. 0.5 Sv is quoted for upwelling in the sector from Cape Columbine
to the Cape peninsula, about 300 km. At the European ocean margin, upwelling is most prominent
west of Iberia, where there is a very strong seasonal signal; a complete reversal of the slope and
shelf circulation between summer upwelling and a northward flow along the slope in winter. There
is also evidence of occasional upwelling west of Scotland and Ireland (BOOTH and ELLETT, 1983;
DICKSON and McCAVE, 1986).

Possible preconditioning is shown by one example off NW Scotland near 59°N (DICKSON and
MCCAUE, 1986); the downwelling-favourable slope current stopped about two days before the
wind blew strongly to force the upwelling (which was another two days later). Similarly, upwelling
in the Gulf of Guinea may depend on a prior raising of the thermocline level in summer (PHILANDER,
1979). Indeed, VERSTRAETE (1992) attributes seasonal upwelling there primarily to remote forcing
to the west, and the equatorial currents and thermocline displacements propagating from the west,
rather than to local or equatorial divergence in the wind-driven Ekman transport.

An example on the western oceanic margin is off Oman during the summer monsoon (ELLIOTT
and SAVIDGE, 1990) with associated along-shore flow $O(10\ Sv)$.

Boundary currents tend to have more influence on upwelling at western oceanic margins. A “standing” cold eddy is associated with the Kuroshio NE of Taiwan (CHERN, WANG and WANG, 1990), being related to the intrusion of the Kuroshio onto the shelf in warm seasons, as simulated by the 2-layer model of SU, PAN and LIANG (1994). The eddy is not evident in winter, but the upwelling is extended and intensified by the NE monsoon in autumn (LIU, GONG, SHYU, PAI, WEI and CHAO, 1992). ANDREWS and GENTIEN (1982) found that intensifications of the East Australia Current were associated with upward movement of cold, saline, nutrient-rich water to the shelf break off the Great Barrier Reef (with subsequent movement across the shelf in a bottom Ekman layer associated with wind-driven flow). Intrusions to the Reef were further related to changing winds and corresponding along-shelf flow by ANDREWS and FURNAS (1986); onshore propagation in the bottom Ekman layer could be quite fast ~0.6 m/s. McCLEAN-PADMAN and PADMAN (1991) found that half the summer occurrences of upwelling near Sydney were associated with inshore effects of East Australia Current eddies. SAHL, MERRELL and BIGGS (1993) show evidence of shelf-edge upwelling in an Ekman layer under a north-eastward current along the Texas-Louisiana shelf; their sections show an aggregate cross-slope transport ~40 m x 40 km ~1 m$^2$s$^{-1}$ for 20 days. Gulf Stream influence is felt along the North Atlantic margin bordering the USA. Thus AIKMAN, OU and HOUGHTON (1988) found an Ekman response only at the shallowest SEEP-I moorings off New England, with oceanic influence in 500 m waters and deeper. CSANADY (1987) estimates 6 Sv upwelling under the Gulf Stream between Cape Hatteras and the Grand Banks, or 4 m$^2$s$^{-1}$, although this quantity is to the slope water; it is not necessarily exchanged with the shelf (indeed, the evidence in §§1.2, 3.4 suggests otherwise). The large figure relates to the fast flow of the Gulf Stream, and potentially supplies the slope water between the Gulf Stream and shelf, including nutrients sufficient for productivity to be limited by diffusion upwards across the seasonal thermocline. Following CSANADY and HAMILTON (1988), WALSH, BISCAYE and CSANADY (1988) estimate 1 Sv for upwelling driven by wind-stress curl over the slope sea west of the Gulf Stream in this sector, but that only 20% of this crosses on to the shelf, some of which in turn is re-exchanged off the shelf before the off-shelf flow near Cape Hatteras. Nevertheless, on-offshelf flows $O(0.01\ m/s)$ were seen in top and bottom Ekman layers in SEEP-I and SEEP-II. In the South Atlantic Bight, the outer shelf is affected by upwelling from the adjacent Gulf Stream water, bottom waters being cooled from May to September (ATKINSON, LEE, BLANTON and CHANDLER, 1983). A calculation may be made from the June-July increase in the fraction of water with salinity < 36 psu on the shelf under upwelling-favourable winds:

$$\frac{0.52 (July) - 0.29 (May)}{5x10^9 s} \times \{2360 \ km^3 shelf \ sea\} \times \{28^\circ-35^\circN (-1000 \ km)\}$$

or about 0.1 m$^2$s$^{-1}$. This is more than reversed by September with downwelling-favourable winds. HOFMANN, PIETRAFESA and ATKINSON (1981) quote a replacement of 20% of Onslow Bay volume by bottom water intrusions in 14-40 days; with a cross-sectional area of about 30 m depth x 100 km width, this is 0.2-0.5 m$^2$s$^{-1}$ locally for the 140 km extent of the Bay. LEE and PIETRAFESA (1987) describe two cold intrusions in circumstances when the Gulf Stream had migrated inshore with a frontal eddy and northward winds; summer stratification also implies less of a barrier effect from horizontal gradients of density. ATKINSON, LEE, BLANTON and PAFFENHOFER (1987) provide more evidence of deep Gulf Stream water appearing in bottom waters on the shelf; although stratification remained strong, the nitrate content was used for primary production owing to the shallow depths on this shelf.
Upwelling budgets were attempted for CODE off northern California. For the spring transition, LENTZ (1987a) found on-offshelf volume transport $O(1\text{ m}^2\text{s}^{-1})$ replacing about half the shelf water in five days. Upwelling-associated heat fluxes were also budgeted by LENTZ (1987b) and RUDNICK and DAVIS (1988).

4.1 Jets, squirts and filaments

“Squirts” were characterised by DAVIS (1985a) as strong localised offshore flows. They are typically associated with “filaments” - elongated bodies of water with characteristics distinct from their surroundings.

STERN (1986) modelled the development of squirts from a convergence in a coastal current, against a coastal “wall” without bathymetry. From the large transport in such jets, KOSRO and HUYER (1986) infer that impinging eddies or a diverted coastal current must be the origin. The 2½-layer model of MCCREARY, FUKAMACHI and KUNDU (1991) reproduces jets and eddies at an eastern boundary, without bathymetry. They review previous stability models and experiment with simpler models to infer two types of instability, as also found by BARTH (1994) with continuous forms of stratification and coastal current. These two instabilities are: a shorter-scale frontal instability depending on gradients of sea-surface temperature, trapped to the front and upper water and extracting potential energy; a larger-scale “traditional” baroclinic instability manifested later in filament development. From satellite imagery and laboratory experiment, NARMOUSA and MAXWORTHY (1989) find that only the frontal instability occurs over a regular shelf under NE Pacific conditions (as north of Cape Blanco). Their experiments associated plumes and upwelling maxima with offshore ridges (notably the Mendocino escarpment), upwelling centres with the downstream side of capes and ridges, and eddy growth with moderate-to-weak stratification (as a factor in the upwelling dynamics). Other models invoke bathymetry; GRIMSHAW and Yi (1991) find steepening of waves in a potential-vorticity front to form thin filaments; MITSUDERA and GRIMSHAW (1991) find a “resonance” between topographic forcing and baroclinic instability if a long-wave phase speed $-c$ in a two-layer model. HAIDVOGEL, BECKMANN and HEDSTRÖM (1991) find dependence on capes (as suggested by observations) and on the southward sense of the (summer) California Current, in their 3-D model. A reconciliation regarding the role of topographic features is that there are basic instabilities of the flow; these set an approximate spacing (corresponding to maximum growth rate) between the disturbances that grow to be filaments; topographic features control the alongshelf location of the filaments (undefined by the instability) and their exact spacing (to which the maximal growth rate is not sensitive).

Filaments are associated with Gulf Stream eddies in the South Atlantic Bight (eg. LEE, ATKINSON and LEEGKEKS, 1981). North-east of Taiwan, CHERN, WANG and WANG (1990) report a filament of mixed shelf and Kuroshio water from both sides of the island. In conjunction with upwelling off Oman, ELLIOTT and SAVIDGE (1990) found evidence in currents and temperature of a cool filament with offshore transport 0.3 Sv. In the Mediterranean (Balearic Sea) off NE Spain, WANG, VIEIRA, SALAT, TINTORÉ and LA VIOLETTE (1988) observed drifters to move faster than the advance of a filament of low-salinity shelf water, inferring convergence and subduction thereof. Off the Atlantic coast of the Iberian peninsula, upwelling filaments first develop in June; they occur principally from late July to extend 200-250 km offshore in September, subsequently diminishing through October which ends the upwelling season (some may remain until December). They typically number five to six and provide a seasonally intense contribution to cross-slope flux (HAYNES and BARTON, 1991; HAYNES, BARTON and PILLING, 1993). Some are related to capes, but two intermediate filaments are suggested to result from instability of the southward upwelling-
related flow. Further south off NW Africa, KOSTIANOY (1992) found the occurrence of filaments to be correlated with upwelling, and concentrated at particular locations along the shelf (18°, 19°, 20½-21°, 23°N). Average width was 30 km and length 130 km (some exceeded 200 km); they formed in the upwelling front and the surface manifestation of development in satellite images appeared to progress seaward at nearly 1 m/s on average (it is not clear if this represents a current or propagation speed). From satellite images, VAN CAMP, NYKJÆR, MITTELSTAEDT and SCHLITZENHARDT (1991) found a filament nearly all year between 20° and 24°N; it varied seasonally and interannually, but reached a length of up to 450 km.

Filaments off the western USA have received much attention. They occur from Point Conception to Cape Blanco (~1000 km) and are described by BRINK (1987), NARIMOUS A and MAXWORTHY (1989), BRINK and COWLES (1991) et seq. cited in BRINK and COWLES (Fig. 4). The many measurements off northern California include the Coastal Zone Transition Program in 1987 and 1988 (BRINK and COWLES, 1991), OPTOMA and CODE. They show alongshore spacing 100-200 km between cold filaments with high nitrate and chlorophyll values, suggesting upwelled water, possibly aided by wind-stress curl. The flow is offshore with evidence of sinking at 10 to 25 m per day, and may form the offshore leg of a meander in the southward-flowing California Current. Sometimes there is an onshore (return) jet along the southern side. DEWEY, MOUM, PAULSON, CALDWELL and PIERCE (1991) describe one filament's structure, momentum balance, vorticity and estimated vertical velocities. Summer 1982 OPTOMA cruises (MOOERS and ROBINSON, 1984) showed cool water associated with offshore flow off Point Arena, a common feature. AVHRR and in situ measurements in 1986 showed three filaments in this area (RIECKER and MOOERS, 1989). CODE measurements included drifters (DAVIS, 1985a,b) and shipborne ADCP (KOSRO, 1987) showing squirts with offshore currents of 0.4-0.5 m/s, especially off Point Arena. The combination of ADCP and CTD (KOSRO and HYER, 1986) showed jets with 1.5 Sv transport. AVHRR and a “to-yo” (FLAMENT, ARM and WASHBURN, 1985) showed a filament off Point Arena as a meander (with its own growing eddies) out to 300 km with transport exceeding 1 Sv. In the Coastal Zone Transition Program in 1987, RAMP, JESSEN, BRINK, NIILER, DAGGETT and BIEST (1991) measured a cool saline filament extending more than 200 km off Point Arena, with some signature down to 500 m and persisting for more than three weeks; currents were 0.6-0.9 m/s offshore on the northern edge and 0.7-1.5 m/s onshore on the southern edge; the estimated offshore transport of 3 Sv greatly exceeds any concentrated Ekman transport. Drifter releases, combined with satellite infra-red images (SWENSON, NIILER, BRINK and ABBOTT, 1992) showed currents up to 1 m/s in a jet broader than a filament off Point Arena, and convergence corresponding to downwelling of as much as 60 m per day over an area of 60 km². In a filament off Point Reyes (“next” south from Point Arena) PADUAN and NIILER (1990) found a strongest jet speed of about 0.7 m/s, moving location in times less than a day; speeds were only 0.2-0.3 m/s a few kilometres away, but total transport was about 2 Sv and again there was some downwelling as water moved offshore through the filament.

Such transport estimates, associated with individual filaments, vary with context. The above for northern California are consistent with BRINK (1987); 0.5 m/s offshore flow in a mature filament 20-50 km wide by 100-200 m deep gives typically 2 Sv. These values hold also for the detailed example of DEWEY, MOUM, PAULSON, CALDWELL and PIERCE (1991). STRUB, KOSRO and HYER (1991) estimate 2 to 4 Sv offshore transport as a meander grows and redirects the California Current. It should be remarked that these transports off California, although originating in an upwelling system, are over deep water and hardly constrained by bathymetry.

A value of 2 Sv was estimated during 1-3 days filament development off NW Africa (KOSTIANOY, 1992) if the surface extension represents a current speed. LUTJEHARMS, SHILLINGTON and DUNCOMBE RAE (1991) estimate 1.5 Sv for an extended filament off Lüderitz in the Benguela.
Figure 4. Upwelling and filaments off California from Cape Mendocino (MD), -41°F, southward to Point Conception (PC) and ~33°N. From Naramou and Maxworthy (1989).
system (see also §3.4). NE of Taiwan (CHERN, WANG and WANG, 1990) an estimated 50 km wide by 100 m deep and 1 m/s gives about 1 Sv after reduction factors \(\frac{1}{2}\) for shelf-water content and for offshore (mis-)alignment. At the other extreme, off NE Spain (WANG, VIEIRA, SALAT, TINTORÉ and LA VIOLETTE, 1988) speeds 0.2 m/s and scales 10 m deep by 10 km broad give only 0.02 Sv.

Overall transport estimates require statistics of filament occurrence in combination with their individual properties. Their regularity in the South Atlantic Bight facilitates the average estimate 0.7 m²s⁻¹ (\(\frac{1}{2}\) time, 20 m thick bottom layer, 0.07 m/s) and 55 x 10⁴ t/year nitrogen import to the shelf (LEE, ATKINSON and LEEGCIKIS, 1981). Otherwise, statistics of occurrence are more uncertain. However, there have been estimates of overall effective cross-shelf diffusivity \(K\) on scales \(\Delta x\), which in depth \(h\) is an effective transport \(Kh/\Delta x\) (§3 after (5); these estimates \(K\) do not distinguish the shelf and slope so that the equivalence is sensitive to arbitrary choices in the following: \(h = 200\) m and \(\Delta x = \) distance of 200 m depth from the coast). HAYNES and BARTON (1991) estimate \(K = 340\) m²s⁻¹ off Iberia, ie. transport 1.7 m²s⁻¹ for \(\Delta x = 40\) km. Off northern California, the CODE drifters mostly remained over the shelf for just a few days (DAVIS, 1985b), suggesting an exchange of the order of

\[
\text{cross-section}\left(\frac{1}{2} \times \text{shelf-edge depth 200 m} \times \text{breadth 20 km}\right) / 10^6s = 2 \text{ m}^2\text{s}^{-1}.
\]

DAVIS (1985b) estimates a coastally-inhibited value \(K = 0.008\) m/s \(\times \Delta x\) (mks units), ie. an equivalent transport \(Kh/\Delta x = 1.6\) m²s⁻¹. [BRINK, BEARDSLEY, NILER, ABBOTT, HUYER, RAMP, STANTON and STUART (1991) estimate a large value 8000 m²s⁻¹ for \(K\) at larger offshore distances (100-500 km) apparently beyond the gross coastal bathymetric inhibition of on-offshore exchange].

These effective diffusivities still carry uncertainties regarding persistence and the extent of shelf to which they apply. Moreover, JONES, MOOERS, RIENECKER, STANTON and WASHBURN (1991) estimate from 1986 OPTOMA 21 data that return shoreward flow from a filament may convey 90% of the volume and 85-90% of the nitrate, ie. the offshore excursion did not result in much true exchange.

HAIDVOGEL, BECKMANN and HEDSTRÖM (1991) estimate an offshore heat flux 0.06 PW per 1000 km in their 3-D model.

The squirts and filaments are distinguished by rapid and confined offshore translation, reversible only if the confined flow turns onshore again. Although the northern California examples are in an upwelling area, the local transports (1 Sv or more; FLAMENT, ARMI and WASHBURN, 1985; KOSRO and HUYER, 1986) appear to be much larger than the upwelling transport \(p^3\) (typically 1 m²s⁻¹) integrated over the length of coastline (typically 200 km) between filaments. In this, there may be a contrast with the Iberian example where these transport estimates are believed to be comparable (A. FRIZA, personal communication) so that the filaments could be "fed" by the upwelling without recourse to return flow.

4.2 Downwelling and cascading

In the initial stages, downwelling is the reverse of upwelling or a relaxation after upwelling. The discussion of causes, scales and forms, based on linear theory, continues to apply (as §4 above and Appendix B). Downwelling is well-documented along the coast of Asia Minor, and modelled by FELIKS (1991). Satellite images of sea-surface temperature show a warm band along this coast, as cyclones over the eastern Mediterranean give downwelling-favourable (easterly) winds of 15 to
30 m/s, typically several times each winter. Relaxation or downwelling intervals occur between upwelling periods off California; e.g. the period April 17-20, 1983 (ATKINSON, BRINK, DAVIS, JONES, PALUSZKIEWICZ and STUART, 1986) when a corresponding westward flow developed past Point Conception, southern California. Downwelling-favourable winds occur for the majority of the year over the Louisiana shelf in the Gulf of Mexico, with apparent effect in convergent flow (ETTER, ULM and COCHRANE, 1985). Downwelling was also observed by LI, WIMBUSH, WATTS and LEE (1986) on the Georgia shelf (eastern USA near 32°N). This was sometimes related to downwelling-favourable winds, usually occurred when the Gulf Stream was close to the shelf, and was probably assisted by winter cooling during the measurement period. There was no evidence of the downwelled or cascaded water going off-shelf beyond the upper slope.

Cascading may be regarded as a special form of downwelling. It occurs when shallow waters on the shelf are cooled in winter to lower temperatures and greater density than adjacent waters over the slope. (The density is enhanced if the cooling forms ice, expelling salt into the water below; then there is a tendency for the dense water to form a distinct bottom layer). The dense shelf waters flow down the slope under gravity, aided by friction and any topographic channelling counteracting the geostrophic tendency for along-slope flow: there is a general understanding of gravity flows in various environmental contexts (SIMPSON, 1987). [If friction is weak, CONDIE and RHINES (1994) find from theory and laboratory experiment that convective "topographic Hadley cells" may form. These would align along the slope and scale as the internal deformation $\varepsilon_i$; several might form on a wide sloping shelf. However, it is not known under what conditions the friction would be weak enough to allow such cells - probably not under winter wind-mixing and cooling.] The down-slope flow entrains ambient water, reaches a depth where the density is the same, and spreads off-slope, as modelled and extensively discussed by PRICE and BARINGER (1994).

Cascading is an irreversible exchange of oceanic and shelf waters. It is sensitive to the hydrographic conditions and degree of winter cooling in any one year. COOPER and VAUX (1949) also suggest that it will be favoured in marginal seas facing the equator (citing the Aegean and Adriatic) because there will tend to be more cooling further from the equator, reinforcing the effect of shallower depth. Proximity to a cold continent in winter might act similarly. However, most shelves are relatively narrow in this sense; the effect of different depths is likely to greatly outweigh any differential cooling.

In laboratory experiments, WHITEHEAD (1993) related the resulting shelf-ocean temperature difference $\Delta T$ to cooling rate $H$ (heat flow per unit area):

$$\Delta T = (g\alpha h)^{-1} \left( 7.4 \, Q_{29} + 0.6 \, f \, W_s \, Q^{13} \right)$$

where $Q = g \alpha h W_\perp / \rho c_p$. Maintenance of a steady $\Delta T$ under differential cooling of the shelf at a rate $H_W$ per unit length implies an exchange rate $H_W / (\rho c_p \Delta T)$ with $\Delta T$ as given above. With typical values (c.f. §0) the Coriolis term is dominant in the expression for $\Delta T \sim 9^{\circ}C$; the exchange rate $\sim 0.25 \text{m}^2\text{s}^{-1}$. Dominance of the Coriolis term gives a dependence of the exchange on environmental factors of the form $h^4 (17 W_s)^{13}$. SYMONDS and GARDNER-GARDEN (1994) describe convective cooling, resulting in greater density in shallower water, followed by geostrophic adjustment to a sloping front. The cross-front transport volume per unit length of shelf, which becomes an exchange after subsequent convective cooling again, is approximately $0.23 h^4 (g \Delta \rho / \rho)^{12}$, i.e. $0.23 h^4 (g \alpha H / \rho c_p)^{12}$ if the density difference $\Delta \rho$ is the result of cooling through the depth $h$ for time $t$. For the typical values above, the exchange at the end of winter cooling ($t = 10^7$) is $10^5 \text{m}^2$, a low rate overall. Increased rates depend on successive cooling events (or some other relaxation of geostrophic adjustment).
In the Gulf of Lions (western Mediterranean), deep winter mixing occurs with dense cooled water sinking in chimneys (MILLOT, 1990) and observed bottom currents up to 0.5 m/s. The deep mixing may take surface material down to large depths (MONACO, BISCAYE, SOYER, POCKLINGTON and HEUSSNER, 1990). Cooling over the shelf is believed to give relatively slow cascading. Clear evidence of cascading off eastern Spain (the Ebro) was found by ANDRÉ and MERLIVAT (1988); tritium increased downwards in depths 900-1500 m near this shelf, in conjunction with decreasing salinity (and temperature) and high oxygen concentrations. Winter cooling and evaporation over the northern Adriatic provides a source of cold, dense (but relatively fresh) water that has been seen following isobaths southwards and sinking down canyons (BIGNAMI, MATTIETTI, ROTUNDI and SALUSTI, 1990). Off north-west Europe, evidence of individual cascading events has been found off the shelf of the Celtic Sea (COOPER and VAUX, 1949), the Hebrides (A.E. HILL, personal communication) and Rockall Bank (D.J. ELLETT, personal communication; Fig. 5). The Bass Strait cascade (SE Australia) may be the best documented (TOMCZAK, 1985, 1987), comprising water of higher temperature and salinity than the adjacent Tasman Sea and manifesting two winter sources at the northern side of the Strait. Measurements by LUICK, KASE and TOMCZAK (1994) show a front and along-slope flow across the eastern entrance of Bass Strait; however, winter-cooled water from the shallow Strait is liable to transgress down the slope in various places, especially to the north. The transport, estimated geostrophically from hydrography for a flow of depth z, is \( g'(\Delta z)^2/f \sim 0.2 \text{ Sv} \) from the Strait, breadth \( \sim 250 \text{ km} \). Much of it forms a northward undercurrent along the slope at 300-400 m depth in winter, extending to 500 km from the sources. The East Australian Current was observed to include packets of cascaded water.

Figure 5. Evidence of late-winter cascading, 1989, off the eastern side of Rockall Bank near 57\(^\circ\)N, 13\(^{1/2}\)\(^\circ\)W (D.J. ELLETT, personal communication).
John M. Huthnance

Yoder and Ishimaru (1989) present evidence of cascaded water in the South Atlantic Bight, consistent with previously upwelled water being cooled by a cold-air outbreak. The (seasonally-limited) transport estimate is large:

\[ 4 \text{ times per month} \times 13 \text{ km wide} \times 80 \text{ m deep} = 1.6 \text{ m}^2\text{s}^{-1} \]

and suggests very strong cooling or alongshelf concentration of the cascading.

Cascading off Spencer Gulf, South Australia, across the shelf to \( \sim 250 \text{ m depth} \) on the slope (Lennon, Bowers, Nunes, Scott, Ali, Boyle, Wenyi, Herzfeld, Johansson, Niell, Petrusevics, Stephenson, Suskin and Wuffels, 1987) is controlled by the excess salinity generated by summer evaporation in the Gulf. Thus it occurs early, during autumn cooling. Similarly, off north-west Africa, the Banc d’Arguin is very shallow and evaporation causes high salinity, so that water sometimes sinks off the Banc to \( \sim 400 \text{ m depth} \) (Mittelstaedt, 1991). Mediterranean water entering the Black Sea has about twice the salinity of the ambient shelf waters, and hence forms a thin (2 to 3 m) bottom layer which cools to the ambient temperature, subsequently sinking along the continental slope (Özsoy, Ünlüata and Top, 1993) until forming a neutrally buoyant plume off-shelf (Oguz and Rozman, 1991).

One documented case of a plume of cold saline water after ice formation is the outflow from Storfjorden, Svalbard. On the basis that the dense water below 100 m in the fjord is renewed each year, the annual outflow exceeds 385 km\(^3\), reasonably corresponding to less than 2 m ice formation each winter (Anderson, Jones, Lindgren, Rudels and Sehestedt, 1988). The outflow has been found in the depression Storfjordenna south of Svalbard, and on the slope at depths between 1000 m and 2100 m (a width between 40 and 60 km) to the west of Svalbard (Quadfasel, Rudels and Kurz, 1988). Temperature and salinity characteristics indicate a 500% volume increase, i.e. a transport \( \sim 0.1 \text{ Sv} \). From the wider Barents Sea, Blindheim (1989) estimates 0.8 Sv outflow of bottom water along the north side of the Bear Island channel. Across the Mackenzie shelf (SE Beaufort Sea) Melling’s (1993) values for buoyancy flux and deficit in the dense plume imply a water flux \( \sim 0.5 \text{ m}^2\text{s}^{-1} \) during winter. Melling and Moore (1995) show other evidence of input from the Beaufort Sea to the deeper Arctic halocline (in some years). From Arctic shelves as a whole, Aagaard, Coachman and Carmack (1981) estimate 2.5 Sv flux of dense water into the polar basin to form the water-mass characteristics of the Arctic halocline with a 10-year residence time. The shelf area \( \sim 3 \times 10^6 \text{ km}^2 \), unevenly distributed along \( \sim 10^4 \text{ km shelf edge} \), is not expected to be uniformly productive of dense water; saline areas (eg. near the Atlantic inflow) and shallow areas are favourable for the production of dense saline water after freezing.

The above estimates form an uncertain basis from which to predict exchanges by cascading. Indeed, initial causes may vary: instability of the flow in geostrophic balance with the cross-slope gradient of density; the bottom Ekman layer under this flow but also affected by the slope and density field. Then any depression in the shelf edge is liable to facilitate and concentrate the process; there, hydraulic control may operate. The balance of these effects is likely to depend on the size and number of any shelf-edge depressions. However, the rate of cascading may ultimately be determined by the rate of production of dense water on the shelf, and frictional processes allowing transfer to the slope in the first instance, as in Melling (1993). Such determination “at source” seems especially applicable if salinity adds distinctly greater density.

The related but distinct topic of deep-water outflows from marginal seas is reviewed by Price and Baringer (1994) for the four principal cases (Mediterranean, Denmark Strait and Faroe Bank Channel outflows from the Norwegian-Greenland Sea, Weddell Sea). For the Mediterranean (and some other marginal seas) evaporation is important in forming dense water (cf. Spencer Gulf).
Cascading (and downwelling) may be distinctive in moving sediment down-slope off the shelf.

4.3 Summary

Upwelling, downwelling and cascading juxtapose ocean and shelf waters. Hence they enable mixing to change water masses and bring about ocean-shelf exchange, especially if the interface between water types is lengthened (Huthnance, 1981). Upwelling may be the largest source of shelf-sea nutrients and production (for example). If upwelling and downwelling alternate, the net effect may still be a supply of nutrients to the shelf; the water downwelled and transported off-shelf may be depleted in nutrients (relative to oceanic upwelled water) having fuelled production while near the surface. By contrast, the results of production may tend to be transported off-shelf. Up/downwelling may also be an important contributor to shelf-sea heat flux (Booth, 1985; Etter, Ulm and Cochrane, 1985).

5. TIDES, SURGES AND INERTIAL CURRENTS

5.1 Tides

Tidal motion occurs primarily on the global scale, generated by the gravitational forces of the sun and moon. The surface elevations would be \( O(0.2 \text{ m}) \) and currents \( O(0.01 \text{ m/s}) \) except for effects of resonance, bathymetry/diffraction and friction. These effects are greatest in marginal seas.

Enhanced response to tidal forcing at frequency \( \sigma \) occurs in a basin of depth \( d \) and radius \( r \) if \( r \sim (gd)^{1/2}/\sigma \) (eg. 1400 km for semi-diurnal \( \sigma \sim 1.4 \times 10^4 \text{s}^{-1} \)). This is comparable with some ocean basin scales; in conjunction with the typical form of cyclonic propagation approximating a Kelvin wave with a coastal maximum (Huthnance, 1981) typical ocean-boundary tides are \( O(1 \text{ m}) \) in elevation. These may be further enhanced over shelves of appropriate width \( W_s \) and depth \( h \), by a \( \frac{1}{4} \)-wave resonance if \( W_s \sim (gh)^{1/2}/\pi/2\sigma \) (eg. 350 km if \( h = 100 \text{ m} \)). A few shelves are this broad: the north-west European shelf, the Argentine shelf, the Gulf of Maine/Bay of Fundy and shelf areas of east Asia and the Arctic. Then increased cross-slope tidal currents \( u \) supply the elevated shelf-sea tide \( \zeta \) if this is all in phase: \( u \sim \sigma W_s \zeta^2 h \) (eg. 0.5 m/s for \( W_s = 350 \text{ km}, \zeta = 1 \text{ m} \) and semi-diurnal \( \sigma \)). The along-slope component of current is typically at least as large, and \( O(\zeta (g/h)^{1/2}) \) if this is greater (eg. 0.3 m/s for \( \zeta = 1 \text{ m} \)).

Marginal seas that are almost enclosed may have reduced tides. For example, the Mediterranean and Baltic Seas are too small to respond to local tidal forcing, and the adjacent oceanic tide is diffracted and thereby reduced through the restricted connecting strait. Then large gradients of tidal elevation may drive strong currents \( u \) through the strait; in principle there is only a hydraulic limit \( u < (gh)^{1/2} \) imposed by a sill depth \( h \) in the strait. In practice, a closer (and much smaller) estimate is again obtained from the transport needed to supply the elevated tide \( \zeta \) in the marginal sea (area \( A \)) if this is all in phase: \( u \sim \sigma A \zeta^2/h b \) where the strait has depth \( h \) and breadth \( b \).

Frictional damping may noticeably affect tidal propagation in extensive marginal seas. An estimate of frictional decay time is \( h/k \sim 10^5 \text{s} > 1 \text{ day} \) (cf. §6). Only shelf seas already having significant tidal structure by virtue of their size (>semi-diurnal tidal wavelength) should be strongly affected.

Tidal elevations are relatively well-known world-wide as a result of coastal tide-gauge measurements; predictions based on these are to be found in national publications. Knowledge also
comes from sea-bed pressure measurements (SMITHSON, 1992) and satellite altimetry. The latter has been used to construct global tidal charts (CARTWRIGHT and RAY, 1991). Global tides have also been modelled, by SCHWIDERSKI (1981) with $1^\circ \times 1^\circ$ resolution using available measurements as a strong constraint, and recently by LE PROVOST, GENCO, LYARD, VINCENT and CANESE (1994) with more detail on some shelves and constrained by measurements only between ocean basins. FLATHER (1981) modelled tides in the north-east Atlantic, including adjacent shelves, with $1/3^\circ \times 1/2^\circ$ resolution and constrained by measurements from the model boundary. RAY (1993) reviews the altimetric and modelled (SCHWIDERSKI, 1981) charts: both have root-mean-square differences O(0.05 m) from "ground truth". Altimetry is slightly better for lunar constituents and the model for solar constituents (which tend to be aliased by satellite orbits). LE PROVOST, BENNETT and CARTWRIGHT (1995) look forward to root-mean-square accuracy 0.02 to 0.03 m in the deep ocean, from continuing Topex/Poseidon altimetry and refined models in combination.

For tidal-current distributions near the continental shelf edge, numerical models are needed in general, owing to complex topography, and have been run successfully, eg. FLATHER (1981) for the north-east Atlantic and PINGREE, MARDELL, HOLLIGAN, GRIFFITHS and SMITHERS (1982) for the Celtic Sea shelf in more detail. In particular, shelf-slope topography may induce anticyclonic coastal-trapped wave motion, maximal near the shelf edge, as part of the diurnal tide at mid-to-high latitudes (HUTHNANCE, 1986). Such motion typically has a length scale comparable with the shelf width (ie. an along-shelf scale less than $(gh)^{1/2}/\sigma$ which is typical of tides generally). PINGREE and GRIFFITHS (1984) have modelled this feature of tides over Porcupine Bank, Rockall Bank and the Hebrides shelf. It is also observed on the Vancouver Island shelf (CRAWFORD and THOMSON, 1982) and elsewhere. Generation is thought to be via some irregularity in the shelf, distorting the regular diurnal tide; resonance builds up the wave amplitude to an extent limited by propagation and frictional loss of the wave energy. The observed cases are believed to exemplify small energy propagation so that frictional loss is limiting. We may roughly estimate the amplitude as

"incident" tidal current $x$ (frictional decay time $\sim 1$ day) / (tidal time-scale $\sim 1/(2\pi)$ day).

In the above observational contexts, the deep-ocean diurnal current is O(0.01 m/s) giving rise to the shelf values O(0.1 m/s).

Cross-slope transports induced by the tide may be large: as above, $hu \sim \sigma W_S \xi$ (14 m$^2$s$^{-1}$ for semi-diurnal $\sigma$, $\xi = 1$ m and $W_S = 100$ km). However, little of this transport may result in genuine exchange, as most of the water returns each tidal period and gives little time for any constituent to suffer non-conservative processes. Vertical mixing between waters suffering different tidal displacements give rise to "shear dispersion" which tends to be the main process of lateral exchange in these circumstances. PRANDLE (1984) found an effective dispersion coefficient to be $t_D \tau$ where $t_D$ was $10^5$s to match a tidally averaged model to observed dispersion on the northwest European shelf. The equivalent local volume exchange across the topographic scale $L_T$ is then $t_D \tau L_T$ (cf. §3), eg. 0.2 m$^2$s$^{-1}$ for $u$ as above and other values as in §8.

Mixing may be especially important in the strong currents through straits, especially if stratification introduces internal hydraulic effects; SIMPSON, SOUZA and LAVIN (1994) show evidence that Pacific intermediate water is entrained into the "estuarine" circulation of the northern Gulf of California. Tidal currents may also be effective in moving suspended particulate matter; the ebb flow may carry a larger load from the more energetic shelf, with subsequent settling in quieter slope waters. YANAGI, SHIMIZU, SAINO and ISHIMARU (1992) describe such "tidal pumping" off Tokyo Bay. TEE, SMITH and LEFAIVRE (1993) model a time-averaged current induced by the tides over Browns Bank off Cape Sable (Nova Scotia), with associated up/
downwelling. However, such tidal rectification is usually very limited and confined close to the shelf break (cf. §3). (Otherwise, as an oscillatory motion, tides do not contribute to the circulation).

5.2 Storm surges

Weather systems, especially winds, of spatial scale 100-1000 km, drive currents $v$ which are more responsive in shallow waters. Initially $v \sim \tau \phi$ for wind-stress $\tau$ of duration $t$ (eg. 0.1 m/s, cf. §0). (Eventually, pressure gradients are set up and friction brings the motion to a steady state in response to a steady wind. Then the along-shelf velocity may be $\rho^1 \tau k$, independent of water depth, eg. 0.1 m/s. Responsiveness in shallow water is matched by the increased effect of friction). As for tides, marginal seas may induce stronger currents through connecting straits, according to the surface elevation difference along the strait.

The forcing scale is generally larger than the shelf width, with duration typically 1 day or more. The greatest response being over the shelf, these forms couple well with the lowest-mode coastal-trapped wave (HUTHNANCE, 1986) having one-signed currents across the shelf and a rapid decrease off-shelf. Thus characteristic responses are anticyclonic polarised oscillations (for shorter scale impulsive winds; Fig. 6) and “quasi-steady” along-shelf flow following slowly-varying winds (GORDON and HUTHNANCE, 1987).

In general, topography and meteorological forcing are complex. For example (BEZARDSLEY, CHAPMAN, BRINK, RAMP and SCHLITZ, 1985) Nantucket Shoals show a greater coherent response to winds than do the shelf seas further east; possible reasons may be a closer coastline, less friction or greater generation distance from the North East Channel (Gulf of Maine). Hence surge-current distributions over the continental shelf and slope usually entail numerical models. Such models are run operationally, eg. for the north-west European shelf (FLATHER, PROCTOR and WOLF, 1991). These 2-D models have been tested for the ability to span the shelf and slope (GORDON, FLATHER, WOLF, KANTHA and HERRING, 1987) with reasonable success for surges not affected by stratification.

As noted above, longer-period wind forcing tends to be associated with along-slope flow, and may result in substantial displacements contributing to the circulation: 0.10 m/s is approximately 8.6 km/day. Shorter periods break the geostrophic constraint to allow cross-slope flow, estimated as $\tau \phi$ (as above, but the duration $t$ is limited to $f^1$). Values 0.012 or 0.08 m/s correspond to $\tau = 0.12$ or 0.8 Nm$^{-2}$ (ie. winds of about 10 or 20 m/s; $h$, $f$ as §0).

5.3 Inertial currents

“Pure” inertial currents are rotary, anticyclonic with constant amplitude and period $2\pi f$ where $f$ is the Coriolis frequency. This idealised form supposes horizontal uniformity. There have been several deep-ocean studies (far from boundaries); for example, RUDNICK and WELLER (1993) review simple theory and carefully isolate from measurements a rotating time-dependent Ekman spiral in response to rotating wind stress. A plane-wave form should theoretically raise the frequency above inertial; background vorticity should also shift the frequency, as invoked (for example) by SALAT, TINTORÉ, FONT, WANG and VIEIRA (1992) for slightly sub-inertial motion off north-east Spain. TREGUER and KLEIN (1994) find analytically that waves propagating against a wind stress may grow, as varying upper-layer depth allows wind-energy input. For barotropic oscillations over the shelf and slope, the “pure” form is modified by the bathymetry to a continental shelf wave of maximum frequency; the currents are sub-inertial but still rotary, anticyclonic with greatest strength near the shelf break (HUTHNANCE, 1981). However, during seasonal stratifica-
Figure 6. Modelled response around Scotland at 1200Z, 5th January 1976, to impulsive wind forcing near Shetland on 3rd January 1976. Rotary currents over the shelf edge are associated with near-diurnal shelf waves having slow energy propagation. Dashed line shows 200m depth contour. (R.A. Flather, personal communication).
tion with an internal deformation scale typically < 10 km on the shelf, they may occur to within ~10 km of the coast (Fox, 1992). Such currents are reviewed for the shelf-sea context by Baines (1986).

The estimate $\tau/\rho h f$ applies for the current (as at the end of §5.2). The response is greater if seasonal stratification reduces the effective depth $h$ to (typically) 30 to 50 m. Then strong shear can be set up across the seasonal thermocline: Largier (1990) emphasises the contribution that this shear (from wind-driven inertial flow and/or internal tides, §7.1) can make to deepening of the surface mixed layer. Indeed, the possible role of inertial waves in downward wind-mixing has long been recognised, following the model of Pollard, Rhines and Thompson (1973).

5.4 Summary

Tidal and wind-driven currents are important in many shelf and slope seas, notably the north-west European shelf where they commonly exceed 0.5 m/s and are generally the largest contributor to turbulence, friction and mixing. Inertial currents may be particularly important to vertical mixing in the interior via their associated shear. Although transports are large (and long-period winds may drive substantial along-shelf displacements of water) the oscillatory character of cross-slope flow tends to limit exchange to the results of shear dispersion.

6. Fronts

Water masses of different characteristics - notably temperature and salinity - may co-exist on the continental shelf and slope as a result of buoyancy inputs (heat from above, freshwater from inshore or advection along-shore) versus mixing (wind and waves from above, tidal-current friction on the bed). The importance of the two buoyancy inputs and both mixing contributions varies with water depth. Typically, sharp changes of water-mass characteristics occur (i) at tidal mixing fronts (Simpson and Hunter, 1974) between summer-stratified waters and shallower seas where tidal currents suffice to mix throughout the water depth, and (ii) between fresher waters inshore and more saline waters offshore (Fig. 7). Such fronts, especially (ii), may occur between shelf and slope waters; there is a theoretical tendency (Condie, 1993) for “anchorage” at the shelf break owing to geostrophy and potential-vorticity conservation. Upwelling fronts occur when a (seasonal) thermocline is brought to the surface.

Typically, along-front flow is in approximate geostrophic balance with the baroclinic pressure field resulting from across-front density gradients. These are generally the strongest flows contributing to circulation in association with fronts.

Friction inhibiting the along-front flow allows the residual pressure field to drive cross-front Ekman transports that tend to restore gravitational equilibrium (Welch, 1991). Over a slope, down-slope flow of the heavier fluid is favoured (Hsueh and Cushman-Roisin, 1983). Convergences occur, eg. at the surface as the less dense side flows towards the front.

A shelf/slope front in a passive tracer distribution may result from the convergence of barotropic flow at the shelf break, because the depth and slope are greater off-shelf (Chapman, 1986). In the interior of 3-D unstratified flow (Gawarkiewicz and Chapman, 1991), the same result holds via cross-slope shear in along-slope flow, but cross-slope transport occurs in the bottom frictional Ekman layer (c.f. §3). In a stratified context, a downwelling Ekman layer leads to convective overturning, vertical mixing on the shelf, and a front with the stratified slope water at the shelf break; here the bottom boundary layer flow detaches to follow isopycnals, which might rise to the euphotic zone (Gawarkiewicz and Chapman, 1992). Given a (less dense) freshwater inflow at the coast, Chapman and Lentz (1994) find that the bottom boundary layer transports the plume and front.
offshore; this process is limited at depth such that the vertical shear in alongshore flow of the plume gives reversed bottom boundary layer flow; then advection can balance diffusion in the cross-slope section.

Fronts (ii) are affected asymmetrically by winds (OU, 1984). If less dense surface water is blown offshore, then the surface layer may shoal and parcels of shelf water detach to traverse the slope; winds blowing surface waters onshore steepen and maintain the front. Off the south-east USA, this was modelled by OEY (1986); winds blowing surface Gulf Stream waters onto the shelf maintained the front with shelf waters; convective mixing by cold-air outbreaks then returned the Gulf-Stream water offshore at depth with a shelf-water component (20%).

Frontal stability has been considered for a range of configurations of sloping fronts and associated shear in along-shelf current, over a shelf and slope (FLAGG and BEARDSLEY, 1978; BARTH, 1989a; CHAO, 1990; HUANG and SU, 1991; BARTH, 1994). In general, the conclusion is baroclinic instability; maximum growth rates are decreased by the bottom slope (usually, depending on the configuration) and by bottom friction (BARTH, 1989b; however, friction can destabilise some wavenumbers); growth rates are enhanced by an overlying seasonal pycnocline (GAWARKIEWICZ, 1991) or horizontal shear (from Warm Core Ring observations; RAMP, BEARDSLEY and LEGECKIS, 1983).

There are various examples of fronts in upwelling regions. Examples of shelf-edge fronts include the Celtic Sea, where PINGREE (1984) presents satellite infra-red and Coastal Zone Colour Scanner evidence of filaments. At the north side of shallow Georges Bank, the shelf-edge front combines with the tidal mixing front (LODER, BRICKMAN and HORNE, 1992). A notable example of a shelf-break front is in the Middle Atlantic Bight (e.g. BEARDSLEY, CHAPMAN, BRINK, RAMP and SCHLITZ, 1985), separating shelf water (mainly responsive to wind forcing) from slope waters affected by Gulf Stream Rings. The front has a large measure of compensation in temperature and salinity effects on density below the summer thermocline. Another example of a front (ii) occurs off north-east Spain (TINTORÉ, WANG and LA VIOLETTE, 1990) where the shelf water is fresher, and cooler in winter, so that salinity is the principal contrasted variable, but evidence has been seen
of saline waters intruding onto the shelf at the base of an eddy.

Cross-frontal transport and exchange depend critically on frontal instability and relaxation of geostrophic constraints to allow cross-frontal flow. The very existence of a front suggests locally reduced diffusivity across the front. However, fronts caused by flow convergence or shear may facilitate exchange by juxtaposing different water masses. Thus CSANADY (1990) concludes that while storms or Warm Core Rings (for example, off the eastern USA) may distort the shelf/slope front, cross-front exchange is the result of instability. Indeed, plumes enclose cyclonic eddies on the Middle Atlantic Bight shelf-break front (perhaps only in summer) and grow on a time-scale ~1 day [GARVINE, WONG, GAWARKIEWICZ, McCARTHY, HOUGHTON and AIKMAN (1988); much faster than the theoretical growth rate in FLAGG and BEARDSLEY (1978)]. GARVINE, WONG, GAWARKIEWICZ, McCARTHY, HOUGHTON and AIKMAN (1988) estimate about 0.1 Sv in a plume (perhaps spaced 100 km apart?) but GARVINE, WONG and GAWARKIEWICZ (1989) find a lack of evidence for frontal eddy contributions to cross-slope flux, and near-zero resulting effective diffusivity.

Displacement of the Middle Atlantic Bight shelf-break front by a severe wind event in April 1988 was recorded by HOUGHTON, FLAGG and PIETRAFESA (1994). It amounted to 50% of the shelf water in nine days or O(2 m²s⁻¹), but mostly returned in the next two weeks.

A seasonal pycnocline above a shelf-break front may allow increased exchange: more near-surface fresh water offshore as above the Kuroshio from the East China Sea (CHEN, BEARDSLEY, LIMEBURNER and KIM, 1994); more saline water penetrating the shelf from offshore, as suggested by GAWARKIEWICZ, McCARTHY, BARTON, MASSE and CHURCH (1990) for the Middle Atlantic Bight, and detailed further south in the Middle Atlantic Bight by FLAGG, HOUGHTON and PIETRAFESA (1994). HOUGHTON, AIKMAN and OU (1988) also suggested an increased exchange of water along isopycnals in summer and autumn, contributing to a heat flux ~0.01°Cms⁻¹ through depth in these seasons (eg. 4.2 x 10⁶ W m⁻¹ in 100 m depth). In winter, they found large variability but mean eddy heat flux ~ 0 and insufficient small-scale mixing for significant net exchange during one “event”. Their overall mean heat flux ~0.008°Cms⁻¹ corresponds to a quoted water exchange rate of 0.2 Sv per 1000 km on the basis of oxygen-18 in the shelf water. This corresponds with the CSANADY and HAMILTON (1988) estimate of 0.3 Sv per 1000 km on the basis of a shelf-water balance of salinity, nutrients and oxygen-18. HOUGHTON, FLAGG and PIETRAFESA (1994) find an average onshore eddy heat flux ~ 4 x 10⁶ W m⁻¹ in SEEP-II (Middle Atlantic Bight) and compare this with similar values from a few other locations. Local eddies of scale 20-45 km are associated with the cool band and different water masses at the Celtic Sea shelf edge. PINGREE (1979) estimates eddy heat fluxes ~0.006°Cm s⁻¹, over the vertical extent of Celtic Sea fronts in summer. The corresponding equivalent exchange of water is α(h'g'h')¹/² (cf. §6), eg. 0.2 m²s⁻¹ if the mixing factor α takes an atmospheric value 0.0055, g' = 0.0065 ms⁻² and h' = 20 m as in PINGREE (1979). Taking the eddy length scale as 10(g'h')¹/²/h' = 30 km, the equivalent lateral diffusivity is αg'h'²/h, eg. 60 m²s⁻¹. Similar across-front values have been found in the North Sea (HILL, JAMES, LINDEN, MATTHEWS, PRANDLE, SIMPSON, GITMROWICZ, SMEED, LWIZA, DURAZO, FOXX and BOWERS, 1993).

In the NW African upwelling region, there are meanders in the off-shelf front between North Atlantic Central Water (in the Canary current to the south-west) and South Atlantic Central Water (in the poleward current against the slope). Inshore, there is mixing on the shelf and another front during strong upwelling. In between near the shelf edge, KOSTIANOY and STEPANOV (1990) found lens-like eddies of typical scale 15 km diameter by 150 m maximum depth. They estimated that up to 50 such eddies per month might be formed in the region of Banc d’Arguin (18°-23°N); if eventually an exchange, this is about 0.2 Sv.

Increased suspended particulate matter was found at the foot of the New England shelf-edge
front (and where a halocline capping a slope-water intrusion abutted the shelf break). PALANQUES and BISCAYE (1992) suggest that internal waves at near-critical tidal frequency (§7.1) are amplified as they propagate shelfwards into the shoaling "wedge" under the front.

7. INTERNAL TIDES AND WAVES

Internal motions, with depth-varying currents, vertical displacements of isopycnals, and sub-inertial periods, are ubiquitous in the ocean. The continental shelf edge is a principal generator of internal tides and consequent solitons contributing to the internal wave field. BAINES (1986) and HUTHNANCE (1989) give reviews of the waves and some consequences near the shelf edge.

7.1 Internal tides

Internal tides are formed when vertical displacements and associated pressure fields are induced in tidal flow across the steep continental slope. Generation appears to be strongest where the bottom slope $h_x$ exceeds the waves' characteristic slope $c$ somewhere on the shelf-ocean depth profile. \[ \tan c = (\sigma^2 - f^2)^{1/2}(N^2 - \sigma^2)^{1/2} \] where $\sigma$ is the tidal frequency in question; $c$ is the slope of particle motion in plane internal waves of frequency $\sigma$. Typically, $h_x > c$ on the steep upper slope; then the source term for the internal tide is concentrated where $h_x \sim c$ near any shelf break. Intensified motion is expected near the sea bed where $h_x \sim c$, and the motion is initially coherent and concentrated along characteristics (paths with local slope $\pm c$) leaving the source (Fig. 8). An extended source serves to spread the motion through depth, as does diffusion with increasing distance from the source. Then a description in terms of a few of the simplest vertical structure modes tends to be more appropriate. Thus mode 1 often describes internal tides propagating onto the shelf; just one flow reversal and one-signed isopycnal displacement through depth. Mode 2 has been identified, as a travelling thickening in the thermocline; mode 3 has been identified propagating oceanwards from the Celtic Sea shelf edge (PINGREE, MARDELL and NEW, 1986; NEW, 1988).

The precise form of the waves and possible bottom-intensified currents are very sensitive to the distribution of bathymetry and density at the shelf edge (BAINES, 1986) on account of the

![Figure 8. Contours (at values of 10, 1, 1/4) of the minimum Richardson numbers achieved during the tidal cycle, for spring tides and the summer stratification in a model for the Celtic Sea shelf edge. The broken lines represent characteristics from the shelf-break source of internal tides. Values of the exponent for instability growth are indicated along these paths. From NEW (1988).](image-url)
concentrated source where $h_c \sim c$. ROSENFELD (1990) found more internal tide in CODE during periods of relaxation as opposed to wind-forced upwelling. OU and MAAS (1988) modelling generation near the New England shelf-break found that the front alone ($\S 6$) was insufficient; the shelf-edge bathymetry was needed to generate the baroclinic tide.

Internal tides propagate slowly, at speeds $\sim (g'h)^{1/2}$, eg. 0.5 m/s. Hence they are affected by strong advective tidal (or other) currents, eg. over Georges Bank (LODER, BRICKMAN and HORNE, 1992) and the Celtic Sea (PINGREE, MARDELL, HOLLIGAN, GRIFFITHS and SMITHERS, 1982).

Strong cross-slope barotropic tidal currents generate large-amplitude internal tides. If vertical displacements are a substantial fraction of water depth, eg. because of the extent of up/down-slope displacements, then the motion is non-linear and higher frequencies are generated, eg. $M_4$ tides from $M_2$. Internal tides with sufficient vertical displacements steepen and form one or more solitons (ie. combinations of higher-frequency internal waves) as seen from satellites by Synthetic Aperture Radar (eg. NEW, 1988) and in thermistor-chain time-series (eg. PINGREE, MARDELL, HOLLIGAN, GRIFFITHS and SMITHERS, 1982; PINGREE, 1984). Typically, a lee depression of isopycnals forms in the off-shelf phase of tidal flow; as the flow slackens, the depression or evolving soliton(s) can propagate on-shelf. More solitons are formed to conserve integral properties of the flow when propagating onto a relatively shallow shelf (JOHNSON, 1973).

Solitons have been formed in laboratory experiments (eg. RENOUARD and ZHANG, 1989) and their generation has been simulated in a 2-D non-linear, non-hydrostatic prognostic numerical model (LAMB, 1994). Analysis of long waves of permanent form is quite well developed, including progression over a shelf (RENOUARD, SEABRA-SANTOS and ZHANG, 1987; GERMAIN and RENOUARD, 1991). Propagation speed increases with amplitude. BOGUCKI and GARRETT (1993) model soliton decay (perhaps 20% in one tidal period) due to shear in the seasonal thermocline.

Internal tidal shear of sufficient amplitude can cause density inversions (ie. static instability) which may lead to mixing for part of the tidal cycle (THORPE, 1987b).

The variety of internal tide models was reviewed by HUTHNANCE (1989). There have been some comparisons with measured internal tides, eg. NEW (1988) for the Celtic Sea; SHERWIN and TAYLOR (1990) for the Malin shelf using a quasi-numerical 2-D model (SHERWIN and TAYLOR, 1989). These models only represent 2-D (cross-slope) generation. The 3-D nature of internal tides has been recognised by WANG (1989; a 3-D time-stepping model of the upper layers in the Strait of Gibraltar) and SERPETTE and MAZÉ (1989; a 2½-D model in the Bay of Biscay). (Fully 3-D rather than 2½-D models are needed to represent internal tide characteristics or vertical structure other than mode 1).

Internal tides are known to occur at many shelf-edge locations world-wide (see HUTHNANCE, 1989). At an extensive subset of these, solitons have been observed. Updating HUTHNANCE (1989), documentation now includes detailed current-meter measurements: on the Australian north-west shelf (HOLLOWAY, 1994); beam structure as expected near generation, with motion concentrated along characteristics of slope $\pm c$, off the Celtic Sea shelf edge (PINGREE and NEW, 1989, 1991); in CODE off northern California (ROSENFELD, 1990); off Vancouver Island where currents reached five times barotropic tidal values and beam structure was also seen, to 40 km offshore from generation on the slope (DRAKOPoulos and MARSDEN, 1993); shear $\sim 0.6$ m/s over Georges Bank (LODER, BRICKMAN and HORNE, 1992; BRICKMAN and LODER, 1993). Trains of large-amplitude internal waves have been recorded over the Guinea Shelf (GORYACHKIN, IVANOV and PELINOVSKY, 1992) and in the Sea of Okhotsk (NAGOVITSYN, PELINOVSKY and STEFANYANTS, 1991). 3-D generation has been demonstrated near the Iceland shelf (PERKINS, SHERWIN and HOPKINS, 1994) and in the Faeroe-Shetland Channel (SHERWIN, 1991) where an internal bore is apparently trapped on the Shetland side by the deep pycnocline at 500 m.
Internal tides are exceptionally large at the Celtic Sea shelf edge, where the cross-slope barotropic tidal currents exceed 0.5 m/s west of Brittany, as shown by the M₃ model of Pingree, MardeLL, Holligan, Griffiths and Smithers (1982). New's (1988) model estimates an energy density exceeding 20 kJ m⁻². The largest instantaneous current observed there, 0.95 m/s downslope, was only 33 m above the steep slope in 548 m (Pingree and Le Cann, 1989) where the M₃ currents are found to be strongest (Pingree, 1988), suggesting enhancement by the local combination of stratification and bottom slope. In the same area, Pingree and New (1989) record internal displacements of amplitude up to 150 m and energy density 8 kJ m⁻².

Contributions to circulation may arise from (i) rectification of internal tidal motion, and (ii) any redistribution of density through associated mixing. (i) Rectified flow scales as $u^2/(\sigma L)$, eg. about 0.02 m/s for a tidal current amplitude $u = 0.1$ m/s, $M₃$ frequency $\sigma$ and length scale $L = 3$ km. (Typically, $L$ may be the internal tidal scale length $(g' h'/\sigma)^{1/2}$ when this is less than the topographic scale $h/h_0$. The factor $u^2$ implies local concentration near tidal current maxima). The cross-slope component is constrained by continuity to have vertical structure rather than any depth-integrated transport. Nevertheless, a down-slope mean reached 0.15 m/s at 500 m off the Celtic Sea (Pingree, 1988); a comparable Stokes Drift (also due to the internal tide) might negate this in the transport of water (Pingree and Le Cann, 1989). Along-slope rectified flow is rather weakly determined by some combination of friction and inertia (a geostrophic along-slope flow being arbitrary). (ii) The interest in sloping-boundary mixing as a source of diapycnal mixing for the ocean interior was mentioned in §1. Internal tide- (and wave-) currents are strong enough to make an important contribution to near-bed mixing (see below).

Energy fluxes integrating to over 10⁸W (eg. 100 W/m for 1000 km) were estimated by Baines (1982) for the internal tides of several coastal sectors, including the north-west European shelf edge, north-east Indian Ocean, Vietnam - East China Sea, Australian north-west shelf, Argentine shelf, Guiana Basin and Gulf of Maine - Grand Banks. A flux of 100 W/m is typical of the estimates reviewed in Huthnance (1989). However, the most energetic locations have fluxes exceeding 1 kW/m, ie. 4.5 x 10⁷ J/m over a semi-diurnal tide. Sherwin (1991) also estimates 2.2-4.7 kW/m in the Faeroe-Shetland Channel. [All these figure aggregate to a negligible energy loss from the barotropic tide (Baines, 1982). However, Morozov (1995) estimates energy fluxes O(10 kW/m) from several ocean ridges reaching relatively close to the ocean surface, and an aggregate energy flux O(10¹²W) which is a significant loss from the barotropic tide.] Where the energy flux is large and carried by solitons, associated with non-linearity, water and its contents are also transported away from the shelf edge.

Fluxes per tide in solitons approximate (c.f. §0)

\[
\text{volume} \quad <\zeta> \lambda, \quad \text{eg.} \quad 5 \times 10^4 \mathrm{m}^2 \sim 1 \mathrm{m}^2 \mathrm{s}^{-1}; \quad \text{energy} \quad \rho g' <\zeta> \lambda, \quad \text{eg.} \quad 2 \times 10^7 \mathrm{J} / \mathrm{m}
\]

Some examples are given by Table 1. Nearly ½ Sv was estimated by Kinder (1984) for tidal internal wave packets east of Gibraltar. Note that the off-shelf Celtic Sea soliton appears to be formed from motion along a bottom-reflected characteristic (New and Pingree, 1990a, 1992), in which case its water is transported only locally, not from the shelf break. The associated supply of nutrients is possibly significant for biological production on the Scotian shelf (Sandstrom and Elliott, 1984) and Pineda (1991) gives evidence of internal tidal bores transporting planktonic larvae (along with cold water) shorewards across the narrow Southern California shelf. Wolanski and Delesalle (1995) describe internal displacements sufficient (100 m) to possibly raise nutrients...
up to the reef system on the shallow Tahiti shelf. Other constituent fluxes are cited in Huthnance (1989). Fluxes of heat and salt are atypical; these quantities are correlated with density which is intrinsic to the motion.

Table 1. Estimates of water volume flux in internal tides.

<table>
<thead>
<tr>
<th>Location</th>
<th>water flux, m³ s⁻¹</th>
<th>&lt;ζ&gt;, m</th>
<th>λ, m</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Georges Bank</td>
<td>1.6</td>
<td>20</td>
<td>4000</td>
<td>Brickman and Loder (1993)</td>
</tr>
<tr>
<td>Scotian shelf edge</td>
<td>0.5</td>
<td>60</td>
<td>2 x 175</td>
<td>Sandstrom and Elliott (1984)</td>
</tr>
<tr>
<td>Mid. Atlantic Bight</td>
<td>0.07</td>
<td>5</td>
<td>6 x 110</td>
<td>Zheng, Yan and Klemas (1993)</td>
</tr>
<tr>
<td>N California</td>
<td>0.16 (If 1/tide)</td>
<td>29</td>
<td>241</td>
<td>Howell and Brown (1985)</td>
</tr>
<tr>
<td>NW Australia</td>
<td>1</td>
<td>30</td>
<td>8 x 200</td>
<td>Holloway (1987)</td>
</tr>
<tr>
<td>Celtic Sea, on-shelf</td>
<td>1</td>
<td>50</td>
<td>1000</td>
<td>Pingree (1984)</td>
</tr>
<tr>
<td>150 km off-shelf</td>
<td>2</td>
<td>50</td>
<td>1700</td>
<td>New and Pingree (1990a)</td>
</tr>
</tbody>
</table>

Thorpe and White (1988) found high nepheloid counts extending ~ 1.4 R₉ offshore from the slope of Porcupine Bank, evidence of cross-slope dispersion.

Internal tides contribute to bottom-intensified currents, stirring, turbulence, friction and dissipation as noted above and reviewed in Huthnance (1989). Several estimates of internal tidal energy loss (i.e. mixing potential) have been made (Table 2). The values in Table 2 generally represent a significant contribution in comparison with bottom frictional dissipation ρC₉V₉ = 3 mW m⁻² for a typical current speed V = 0.1 m/s. The larger figures exceed typical wind energy inputs τV = 10 mW m⁻² (cf. §§3-5). It should be realised that all these figures are reduced by an efficiency factor between energy input and potential energy resulting from mixing: Stigebrandt and Auer (1989) estimate an average 5.6% in fjords from internal tides acting on the sloping bottom and at the thermocline. Internal-tide energy input is distinctive in acting internally and relatively efficiently to cause diapycnal mixing, e.g., to diffuse the summer thermocline (Sherwin, 1988). For example, the modest energy input for Cape Point Valley (South Africa) corresponds to an estimated vertical eddy diffusivity 10⁻⁴ m² s⁻¹ (Largier, 1994) which greatly exceeds adjacent oceanic interior values.

Table 2. Estimates of internal tidal energy loss, potentially available for mixing.

<table>
<thead>
<tr>
<th>Location</th>
<th>flux, kW/m</th>
<th>distance, km</th>
<th>loss, mW m⁻²</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>NW Australia</td>
<td>3 (x 40%)</td>
<td>0.63 ms⁻¹ x 12.4h</td>
<td>42</td>
<td>Holloway (1987)</td>
</tr>
<tr>
<td>Georges Bank</td>
<td>(8400kW/m/tide)</td>
<td>10</td>
<td>20</td>
<td>Brickman and Loder (1993)</td>
</tr>
<tr>
<td>Scotian shelf edge</td>
<td>(6x10⁷W/m/tide)</td>
<td>10</td>
<td>50</td>
<td>Sandstrom and Elliott (1984)</td>
</tr>
<tr>
<td>Malin shelf</td>
<td>0.1</td>
<td>100</td>
<td>1</td>
<td>Briscoe (1984)</td>
</tr>
<tr>
<td>Cape Point Valley</td>
<td>0.04</td>
<td>15</td>
<td>3</td>
<td>Sherwin (1988)</td>
</tr>
</tbody>
</table>

Effects of mixing over the Celtic Sea shelf edge and beyond have been studied extensively. Pingree, MardeI, Holligan, Griffiths and Smithers (1982) and Pingree, MardeI and New (1986) show cross-shelf sections with a thickened seasonal thermocline and cool near-surface waters with enhanced chlorophyll indicating primary production near the shelf break. Temperature measurements in Pingree (1988) show evidence of overturning (in patches lacking a vertical gradient) substantiated by nutrient and chlorophyll measurements. New (1988) models the internal tide generation to show strong shear locally:
Richardson number \( \text{Ri} \equiv (\text{buoyancy frequency / shear})^2 < 1/4 \)

suggests instability, and during spring tides (but not neaps) this condition persists long enough to predict mixing, particularly in the seasonal thermocline and especially on the ocean side. There is agreement with observation; New and Pingree (1990b) observed cooler sea-surface temperatures at appropriate distances offshore and inshore from the shelf-break source, measured Ri and found evidence of upward mixing of cool water. They also estimated that upward mixing to cool the upper 50 m by 0.3°C over a distance of 20 km across the slope would require only 1% of the internal tidal energy radiated to the offshore near-surface sector during a spring-neap tidal cycle. [The band of cooler surface water along the Celtic Sea shelf edge was earlier seen in satellite infra-red images, eg. Dickson, Gurbett and Pillai (1980). The surface cooling may be enhanced by wind mixing in addition (le Tareau and Maze, 1993). Evidence of Ri < 1/4 has also been found at the Scotian shelf edge (Sandstrom and Elliott, 1984) and Ri < 1 extensively on the Malin shelf (Sherwin, 1988)]. Evidence of mixing, including near-surface nutrients, hundreds of kilometres offshore from the Celtic Sea (Holligan, Pingree and Mardell, 1985) is now attributed to the local generation of solitons (New and Pingree, 1992) where a characteristic from the source reaches the seasonal thermocline after downward propagation (Pingree and New, 1989) and reflection upwards from the bottom (Pingree and New, 1991).

Thorpe and White (1988) found evidence that the offshore nepheloid layer was most extensive at a depth where the slope \( h_x \) of Porcupine Bank matched the characteristic slope \( c \) for the \( M_2 \) tidal frequency. Clearly any such occurrence of bottom-intensified currents encourages the movement of bottom sediments. Heathershaw, New and Edwards (1987) showed that internal tides at the Celtic Sea shelf break could potentially be responsible for sediment transport to the deep ocean. Other qualitative examples are cited in Huthnance (1989).

### 7.2 Internal waves

These occur in the ocean with a complete spectrum of frequencies from the Coriolis frequency \( f \) to the buoyancy frequency \( N \), and all wave directions. In principle the spectrum should be composed of vertical structure modes (over a flat bottom). However, for short waves the wavelength is arbitrary; the wavenumber is constrained only to be normal to the plane of motion (which slopes at the angle \( c \) given in §7.1). The oceanic spectrum is empirically near-universal except very close to source regions (Garrett and Munk, 1979), with an energy density corresponding to an estimated shelfward energy flux of the order of 1 kW/m² (Huthnance, 1981). [In the Arctic ocean, Pisarev (1992) finds internal wave energy less than in other oceans, but still a relative maximum at the shelf edge, and a broad band in wavenumber]. If the sea-floor slope \( h_x \sim c \), then the currents may be amplified several times within a bottom boundary layer (Gordon, 1980) which becomes thick (tens of metres) as \( \sigma \rightarrow f \) (Weatherly, Blumsack and Bird, 1980).

The shelf edge provides an effective source of internal waves: at internal tidal frequencies as discussed in §7.1 with higher-frequency contributions if non-linearities are significant; as standing lee waves (with associated form drag and some bias towards on-shelf energy flux) in longer-period flow along a rough continental slope (Thorpe, 1992b) and analogously around a seamount (Chapman and Haidvogel, 1993). Moreover, internal wave motion along characteristics is reflected off the sloping bottom, with a change of wavelength (Thorpe, 1987a). The shorter waves are subject to distortion by the motion of the longer waves and the \( M_2 \) tide (Thorpe, 1989); waves at the characteristic frequency for the bottom slope may suffer parametric instability (Thorpe, 1994); associated non-linearities imply wave-steepening (Thorpe, 1987a), a possibility of thermal
Exchange at the ocean margin

fronts (THORPE, 1992a) and mean currents (TAYLOR, 1993) which may suffice to invert the stratification, giving static instability. Laboratory experiments (HELFRICH, 1992) have also shown 2-layer solitary waves breaking if the bottom layer shoals to about three times the wave amplitude. Empirically, various uncertainties in stratification and bottom slope, and irregular bathymetry, may mask near-bottom enhancement over a critical slope (GILBERT, 1993) and evidence of non-linearity in the reflected wave may be clearer.

[Internal wave energy may also transmit to the shelf, especially if the bottom slope $h_s$ is less than the characteristic slope $c$ for the wave. Thus a response or oscillation on the shelf is induced. A few papers are cited in HUTHNANCE (1989); GRIMSHAW and CHAPMAN (1992) present theory in the case where oceanic stratification lies below the shelf break].

Effects on circulation may be via mixing (see also below) and form drag. THORPE (1988) speculates that internal-wave input to turbulence at a distance above the bottom may give a special structure to the associated bottom boundary layer for the circulation. Form drag occurs as along-slope currents generate internal lee waves (THORPE, 1987a) and may be comparable with bottom stress. There are few direct measurement of “eddy” fluxes of momentum and buoyancy associated with internal waves and corresponding to stress/drag and mixing. VAN HAREN, OAKEY and GARRETT (1994) carried out an experiment on the sloping side of Emerald Basin in 178 m, and found the main contributions at high frequencies near N.

Cross-slope (exchange) transport, as a result of mixing and spreading, was estimated by FLAGG (1988) at about 0.05 Sv in 900 km in the late-winter context of SEEP-I (Middle Atlantic Bight). Near-bottom current variance was 0.0016 m²s⁻² in 120 m and 0.0009 m²s⁻² in 80 m water depth (15 km distant onshore). From the internal wave energy flux divergence, a loss $\sim 10$ mW m⁻² to energy available for mixing was inferred, comparable with large wind events. SANFORD and GRANT (1987) model the effect of surface waves in dissipating internal waves on the continental shelf. Limits to internal wave propagation on the shelf may also be imposed by other dissipation (e.g. tides and surge currents) or lack of stratification. Typically, 10% of the incoming energy flux (eg. 100 W/m) may be lost from the internal waves in 10 km, a potential input 10 mW m⁻² to mixing. A smaller value, 1 mW m⁻², is probably typical for the deeper waters of the continental slope (GARRETT and GILBERT, 1988) where internal waves may be expected to suffer less dissipation but short reflected waves themselves generate high shears within tens to 100 m of the bottom.

Enhanced mixing is usually the main impact of internal waves at the ocean margin. Steepening and periods of static instability near the bottom facilitate energy input into mixing (THORPE, 1987a; TAYLOR, 1993). THORPE, HALL and WHITE (1990) found that the $M_2$ tide (on the Hebrides slope and Porcupine Bank) modulates the mixing context; “mixed” layers appeared within the tidal cycle, followed by static stability down to a few metres above the bed. In laboratory experiments, TAYLOR (1993) found that $h_s - c$ led to wave-breaking and turbulent mixing, which were sensitive to wave-amplitude increase through shoaling and the effect of currents from the previous wave. The turbulence time-scale was longer than an individual wave period.

Sediment movement may be favoured when internal waves enhance near-bottom currents. Thus nepheloid layers formed off Porcupine Bank, when upwelling induced stratification such that $h_s - c$ (DICKSON and MCCAVE, 1986; on another occasion it was observed that increased internal wave energy occurred near the bottom following upwelling-favourable winds). In SEEP-II (Middle Atlantic Bight), at 131 m water depth, internal waves approaching the shelf frequently added to the lower-frequency flow to take the total above the speed threshold for sediment resuspension (CHURCHILL, WRICK, FLAGG and PIETRAFESA, 1994). Surface waves became of more importance for this, in shallower shelf waters.
8. SURFACE WAVES

Surface waves comprise oscillations of the sea surface, individual contributions of which are sinusoidal in time and space, accompanied by sub-surface currents describing an ellipse in the vertical plane of propagation. The ellipse tends to rectilinearity at the sea bed, and the currents decrease downward, the vertical scale being $\kappa^{-1}$ where the horizontal wavenumber $\kappa (=2\pi/\text{wavelength})$ is related to the wave frequency $\omega_w$ by $\omega_w^2 = g \tanh \kappa h$. Hence only long (low-frequency) waves with $\kappa h < 2$ “feel” the bottom, e.g. periods $> 10, 14, 20$ s approximately for depths $50, 100, 200$ m respectively. There is a greater likelihood of wave amplification through refraction and concentration of energy by a strong current, e.g. the Agulhas current (SMITH, 1976).

Waves may contribute to circulation and exchange (i) via non-linear rectification of their currents, and (ii) by modifying the wind stress. (i) Surface currents $\mathbf{u}$ for waves of amplitude $a \sigma_w$ (e.g. 1 m/s, cf. §0). Hence the Stokes Drift velocity $\langle \mathbf{u} \mathbf{d}t \rangle \mathbf{w}$ (with $\mathbf{V}$ scaled by $\kappa = 0.1 \text{ m}^{-1}$) may commonly exceed 0.1 m/s, typical wave heights being 2 to 3 m (CARTER, FOALE and WEBB, 1991). On the basis of a typical wave spectrum related to the wind speed $w$, KENYON (1969) estimates surface currents 0.015 $w$ to 0.035 $w$, decaying with depth on a scale $w/3$ in mks units. The Stokes Drift (Lagrangian - Eulerian) transport $\sim 0.01 w^2$ is thus comparable with the Ekman transport $\tau/p f \sim 0.01 w^2 \rho_{\text{air}}$ and air/water ratio of densities $\rho \sim 10^3$). Typically, waves and the associated Stokes Drift have an on-shelf component owing to the greater scope for generation provided by the off-shelf area of ocean (exaggerated where prevailing winds blow from ocean to shelf). It should be noted that this transport is concentrated near the surface and represents a difference (tracked water movement - average velocity at one point) rather than determining the absolute circulation. (ii) A developing wave spectrum increases the drag between the wind and the sea surface (e.g. JANSSSEN, 1991). Allowance for this effect has been found to improve the representation of wind-driven flow in shelf-sea models (MASTENBROEK, BURGERS and JANSSSEN, 1993).

Waves may contribute to turbulence and mixing through (i) near-surface shear and breaking, and (ii) currents at the bed. (i) Near-surface shear is $a \sigma_w \kappa$, e.g. 0.1 s$^{-1}$. This is large, but decreases for long waves (smaller $\sigma_w$ and $\kappa$) and rapidly reverses during the wave oscillation. The overall effect on the waves is estimated in the wave model WAM (for example) by an energy dissipation rate $\sim 3 \times 10^{-5} \sigma_w$, for a reasonably well-developed spectrum (WAMDI GROUP, 1988). The wave-energy density being $\frac{1}{2} \rho g a^2$, the potential input to near-surface mixing is thus $1.5 \times 10^{-5} \rho g a^2$ approximately, e.g. 150 mW m$^{-2}$. An alternative (MELVILLE, 1994) relates the energy dissipation $5 \times 10^{-7} \rho w^3$ to the wind speed $w$ (e.g. 500 mW m$^{-2}$ for $w = 10$ m/s). There is no special effect of the shelf edge in comparison with elsewhere in the ocean. (ii) Near-bed currents $a \sigma_w / \sinh (\kappa h)$ are small for typical shelf-edge depths for all but the longest waves and shallowest depths $h$ (Table 3). Where they are comparable with other contributions to near-bed currents, their oscillatory character enhances the power to suspend sediments from the bed (e.g. SOULSBY, HAMM, GLOPMAN, MYRHAUG, SIMONS and THOMAS, 1993). This, and asymmetric flow due to waves’ non-linearity, renders waves particularly influential in the transport of particles across shallower continental shelf waters as reviewed by NITTRUER and WRIGHT (1994).

Table 3. Nearbed currents (m/s) for surface waves of amplitude $a = 1$ m.

<table>
<thead>
<tr>
<th>$h$, m</th>
<th align="right">$1$ (6.3)</th>
<th align="right">$0.5$ (12.6)</th>
<th align="right">$0.3$ (20.9)</th>
</tr>
</thead>
<tbody>
<tr>
<td>50</td>
<td align="right">0.013</td>
<td align="right">0.26</td>
<td align="right">0.38</td>
</tr>
<tr>
<td>100</td>
<td align="right">$10^4$</td>
<td align="right">0.08</td>
<td align="right">0.22</td>
</tr>
<tr>
<td>200</td>
<td align="right">$&lt;10^4$</td>
<td align="right">0.007</td>
<td align="right">0.09</td>
</tr>
</tbody>
</table>
9. CAPES AND CANYONS

Along-shelf curvature of depth contours or changes of depth may affect ocean-margin circulation, exchange and mixing processes in several ways outlined in the following sections. In addition, the extra length of an irregular ocean-shelf boundary increases the scope for ocean-shelf exchange.

9.1 Cross-contour flow

The constraint for geostrophic flow $v$ along depth contours may be relaxed by canyons and shelf-edge irregularities of small-enough scale $L$. (i) Stratification (represented by $N$) decouples the flow from the bathymetry if $L < R_k = h_o N/v$ (eg. 10 km for $N = 10^{-3} s^{-1}$). (ii) Geostrophy is broken by inertia if $L < v/f$ (eg. 1 km); more generally, cross-slope flows scale as the smaller of $v^2/fL$, $fL$. (iii) A small cyclonic bend $\theta$ in the slope (steepness $h_x$) causes an off-shelf displacement $\theta h_x$ in an along-slope current (Huthnance, 1987b).

In cases (i) and (ii), if the relative changes of depth are small, then the constraint is correspondingly weak. In cases (i) and (ii) also, the depth contours may be regarded as effectively interpnd, ie. there is a "topographic western boundary" with an associated boundary layer admitting cross-slope flow as exemplified north of Taiwan (cf. §3.1; Su and Pan, 1987; Qiu and Imasato, 1990). The forcing of open-ocean flow from a narrowing or terminating shelf (Johnson, 1991) may be viewed in this light.

Thus Killworth (1978) found that flow followed a gently curving coast, whereas a boundary-current model (Hughes, Ofosu and Hickey, 1990) and 3-layer theory (Klinck, 1989) predict flow passing across the top of a narrow canyon with little deviation. These models also show pressure-induced upwelling, and hence cyclonic flow within and anticyclonic flow above a canyon, represented by flows $O(v)$ along the sides of the canyon and local thereto (length scale $R_k$); for a narrow canyon, the result may be deflection of an alongslope current in the opposite sense from the deviation of depth contours. Flows following tortuous contours on the canyon scale juxtapose shelf and oceanic waters, already appear (from a broader viewpoint) as dispersion, and add scope for genuine exchange via mixing. Nof (1980) finds that an anticyclonic inertial boundary current is compressed over a ridge (axis normal to the coastline) if the beta-effect is relatively weak; the effect is reversed over a canyon; if the latter is deep enough, the current could separate from the coast. Haynes, Johnson and Hurst (1993) consider 1-layer flow (inertial effects) along a curved step, finding a downstream displacement across it at intermediate values of $v/fL$. Gjevik and Moe (1994) analyse and model the effect on along-shelf flow of a local topographic elevation on the shelf; cross-shelf flow and waves are generated, but little flow across the continental slope. A boundary current anticyclonic around the ocean can generate lee waves and hence offshore flow just to the lee of a bluff cape (Freeland, 1990). Laboratory experiments of barotropic flow along a slope around a sharp bend (radius $< v/f$) show separation if $v/fL$ exceeds about 0.1 (Klinger, 1993), with an excursion (before down-stream reattachment) proportional to $v/f$ and to the inverse depth change across the slope. Theory for a single active layer representing a coastal current (Klinger, 1994) similarly shows a critical radius of curvature for which centrifugal force induces separation from a rounded cape. Experiments for stratified flow $v$ past a cape (Boyer and Tao, 1987) show transverse flow; starting and lee eddies scaling with $v$ and with the cape's projection ($L$).

Point Conception appears to be an example where an (upwelling) along-shelf current may fail to curve sharply with the convex coast and instead separates as an offshore jet (Atkinson, Brink, Davis, Jones, Paluszkiewicz and Stuart, 1986). In the Black Sea, Sakarya Canyon almost completely intersects the shelf, intercepting the "rim current" current along the slope and deflecting...
it offshore (Özsoy, Ünlüata and Top, 1993). Standing eddies, with circulation around bodies of water originating further offshore, have been inferred in Pribilof and Juan de Fuca canyons (for example: Kinder and Coachman, 1977; Crawford and Thomson, 1991). Examples of narrow canyons are: Lydonia Canyon south of Georges Bank (Noble and Butman, 1989) which is not penetrated by the subtidal flow along the shelf, but forced by the associated cross-shelf pressure gradient and exhibits an on-offshore "exchange mode" flow at the mouth of the canyon rim; Astoria canyon (Hickey, personal communication) where the flow corresponds to the model findings of Hughes, Ofosu and Hickey (1990) - the flow passes across the top with little deviation, and upwelling within induces deep cyclonic flow and anticyclonic flow above; Monterey canyon (Rosenfeld, Schwing, Garfield and Tracy, 1994) across which cool upwelled water from Pt. Año Nuevo appears to be advected southwards. Durrieu de Madron (1994) inferred that flow past the Grande-Rhône canyon mouth was somewhat deflected into the canyon, with anticyclonic flow at higher levels in the canyon, but also down-slope flow at the bottom. To the south-west, Masó and Tintore (1991) found that the canyons of NE Spain deflect the along-shelf current with the formation of cyclonic and anticyclonic eddies, and allow a cold saline intrusion from the open ocean towards the indented shelf. Church, Mooers and Voorhis (1984) found that dense "cold pool" shelf water moved cyclonically around the rim of Wilmington Canyon (ie. in the sense expected of a gravity current) as well as protruding beyond the rim to mix with slope water which extended into the canyon from off-shelf. Hickey (1992) found that drifters drogued at 15m and 40m were deflected over Santa Monica Canyon (scale L \~ 5 km, H \~ 200 m). Quinault Canyon (Washington State) is larger - up to 30 km broad and indenting the shelf by 15 km with slopes 6° to 30° in 200-600 m; however, Baker and Hickey (1986) found some cross-contour flow in the regional circulation at the head of the Canyon.

In a numerical model of the Kuroshio adjacent to the Japanese shelf, Awa, Akitomo and Imasato (1991) found that the flow set up eddies, ~100 km in diameter, behind submarine spurs projecting from capes. Kuroshio water was trapped in such eddies on the shelf, amounting (for example) to 20% of the shelf-sea volume in one 50-day event.

Other cross-contour flows related to capes, etc. occur in upwelling contexts (§4).

The Northeast Channel (30-40 km wide) to the Gulf of Maine appears to allow relatively free shelf-slope exchange. Ramp, Schlitz and Wright (1985) found inflow averaging about 0.3 Sv on the eastern side below 75 m, and a smaller outflow of less saline water on the western side. This net transport corresponds to exchange across a contour length ~1000 km within the Gulf (as opposed to the much shorter straight-line boundary with the Atlantic). (The corresponding residence time in the deeper waters of the Gulf was estimated as 11 months). Significant heat and nitrogen are thought to be contributed to the budgets in the Gulf of Maine.

Shelf-ocean exchange via a canyon may also occur as a deep-water "return" flow complementing wind-driven flow and set-up in shallow water on the shelf. Such is the interpretation of outflow from the North Sea over the Norwegian Trench, a consequence of prevailing winds driving water into the shallow southern North Sea as modelled by Davies and Heaps (1980). In such a context (Fig. 9) the exchange scales as \ht/\pk, eg. 10 m²s⁻¹ (Appendix C). A refinement (Shepard, McLoughlin, Marshall and Sullivan, 1977) is that set-up may be less across the narrower shelf at the head of the canyon; the alongshelf pressure gradient drives flow convergent and turning off-shelf at the canyon. Fluctuations in the Northeast Channel to the Gulf of Maine were correlated with wind stress and set up on the shelf (Ramp, Schlitz and Wright, 1985).

If along-shelf flow around a canyon rim is in the sense giving a downwelling Ekman layer, then suspended particulate matter (SPM) carried therein is liable to be trapped in the canyon, as posed by Millot (1990) for the Liguro-Provençal current and Lacaze-Duthiers canyon, Gulf of Lions.
The latter canyon is believed to be a "conduit" for much of the SPM export from the Gulf to the deep Mediterranean (MONACO, BISCAYE, SOYER, POCKLINGTON and HEUSSNER, 1990; MONACO, COURP, HEUSSNER, CARBONNE, FOWLER and DENIAUX, 1990); evidence includes higher SPM concentrations than the rest of the continental slope in the Gulf of Lions. SPM from the 10-30m benthic nepheloid layer on the shelf is thought to become an intermediate nepheloid layer at canyon heads, especially where the shelf narrows "downstream" towards Lacaze-Duthiers canyon (DURRIEU DE MADRON, NYFFELE and GODET, 1990).

Barrow Canyon, ~25 km wide on the northern Alaskan shelf, appears to be a focus for Arctic Ocean-shelf exchange in both senses. Flow is well-aligned with the canyon axis and down-canyon in the mean. However, down-canyon flows, of cold relatively-fresh water, alternate with warm saline upwellings onto the shelf (AAGARD and ROACH, 1990). Over short time-intervals of a few days, down- or up-canyon fluxes O(20m thick x 25m wide x 0.2m/s) = 0.1 Sv were recorded, amounting to O(10%) of the time in each sense.

Figure 9. Deep embayment (appendix 3) with wind stress \( \tau \) inducing raised surface elevation at the head (right). The resulting exchange is via inflow in the shallower water "1" driven by the wind, and return flow in the deeper water "2" where the surface-slope-induced pressure gradient is more effective.
9.2 Local upwelling

This is predicted to occur on the equatorward side of a cape in equatorward eastern-margin flow (ARTHUR, 1965). However, ROSENFELD, SCHWING, GARFIELD and TRACY (1994) present a rather different argument, that there will be a maximum offshore Ekman transport and associated minimum coastal sea level, where a convex coastline (cape) is aligned with the wind; then geostrophic flow will tend to enhance offshore transport on the poleward side. Upwelling is also expected to be intensified over a ridge that is normal to the coast (KILLWORTH, 1978) after upwelling-favourable winds. (Correspondingly, downwelling is favoured on the poleward side). Beside a ridge, the associated shoreward (upwelling) transport \(-2(\Delta h/h_0)\tau/pf\) after sufficient time of application of the wind stress \(\tau\); this value is the Ekman transport \(\tau/pf\) (§5) with a reduction factor for the depth change \(\Delta h\) over the ridge relative to the ambient depth \(h_0\). Cross-slope exchange is not necessarily implied around a cape.

The association, albeit unquantified, has been made in observations at several locations. DICKSON, GURBUTT and PILLAI (1980) ascribe the cool band at the Celtic Sea shelf edge to upwelling enhanced by canyons and ridges, although mixing by internal tides (§7.1) is now thought to be important. SHAFFER (1976) has seen evidence of shelf-edge canyons guiding upwelling off north-west Africa. Off north-east Spain, flow to the south-west upwells onto the shelf at a concavity in the depth contours, with subsequent off-shelf flow (FONT, SALAT and JULIA, 1990). There is evidence for topographically-induced upwelling of the East Australia Current around the "cape" formed by Swain Reefs (Capricorn Channel; KLEYPAS and BURRAGE, 1994). Upwelling occurs at Point Conception, southern California (eg. ATKINSON, BRINK, DAVIS, JONES, PALUSZKIEWICZ and STUART, 1986). The association of upwelling with the capes of northern California has been much remarked. Analysing CODE measurements, KELLY (1985) found that the principal sea-surface temperature variability, after seasonal warming, was in spatial structures representing irregular upwelling enhanced at capes. Measurements in 1987-1989 (BRAY and GREENGROVE, 1993; LARGIER, MAGNELL and WINANT, 1993) show upwelling to the south of Point Arena and Cape Mendocino. Offshore flow on the north side and onshore flow on the south side (1 to 2 Sv in each case) was inferred from divergence of the along-slope mean transport. There was a minimum in temperature and equatorward flow, and a maximum poleward undercurrent, south of Cape Mendocino. LAGERLOEF (1992) associates the cold filament off Point Arena with a co-evolving meander and eddy in the California current, manifesting a standing wave phase-locked to Cape Mendocino.

9.3 Wave reflections

Intensified flow may occur in canyons as a result of internal wave reflections and trapping related to the particular bathymetry. Under weak stratification (bottom slope \(h_x < \) characteristic slope \(c\) for the waves, c.f. §7) greater cross-shelf currents are expected at the head of the canyon (BAINES, 1983). For strong stratification (\(h_x > c\)) successive reflections within the canyon, only partial radiation from the mouth, and reflections intensifying external wave forcing, may all serve to intensify oscillations at the bottom of (especially) a steep narrow canyon (GORDON, 1982; GRIMSHAW, BAINES and BELL, 1985). If canyon-ridge bathymetry approaches the coast to refract incoming surface waves, concentrations of wave action and variations of set-up driving longer-period edge waves and circulations may occur (INMAN, NORDSTROM and FICK, 1976).

Observations of intensified current fluctuations have been made in several canyons, in some cases accompanied by increased turbidity. The Lacaze-Duthiers canyon (Gulf of Lions) appears to be a focus for internal wave energy (MONACO, BISCAYE, SOYER, POCKLINTON and HEUSSNER,
1990); 0.1-0.15 m/s currents therein could be actively erosive (MILLOT, 1990) contributing to its large concentrations of suspended particulate matter. In canyons and gullies on the slope of the Middle Atlantic Bight, CSANADY, CHURCHILL and BUTMAN (1988) found reduced energy at long periods (> 5 days) but enhanced high-frequency and inertial motion over the upper slope. In canyons off Southern California, bottom currents up to 0.5 m/s have tidal frequencies and phase propagation up the canyon, with higher frequencies near the head (SHEPARD, 1976). HOTCHKISS and WUNSCH (1982) observed intensified internal wave energy at the head and bottom of Hudson Canyon and relate the effect to the width of the canyon (not larger than the internal deformation scale \( R_I \), c.f. §9.1), bottom intensification and the decreasing depth and width for waves propagating in from the ocean. The observed energy densities actually suggested dissipation from bottom friction and other causes; possibly internal mixing and at the head of the canyon. However, one storm in the 15 weeks measurements, giving sediment-bearing bottom currents \( \approx 0.5 \) m/s for 16 hours, proved to be more effective than internal waves, except at the canyon head. Baltimore Canyon is also deep and reasonably narrow (8 km wide at the shelf break) with currents enhanced by the focusing of internal waves (GARDNER, 1989). Apparent consequences are a thickened thermocline, and increased turbidity in the canyon. The influence of storms was not seen to penetrate much deeper than shelf depths, but mean currents were down the canyon axis at 275 m, up-axis at 600 m. Resuspension occurred between these depths with an offshore flux along density surfaces. Here and in Quinault Canyon, which lacked suspended sediment transport along the bottom (BAKER and HICKEY, 1986), there is evidence of fast settling in aggregated particles. SHEPARD, MCLoughLIN, MARSHALL and SULLIVAN (1977) record observations of currents \( \approx 0.5 \) m/s at the bottom of varied canyons with a sequence (i) large up-canyon flow, (ii) rapid build-up of a turbidity current down the canyon (iii) slow decay, (iv) an interval with relatively small currents.

SEYMOUR (1990) reviews turbidity flows, which typically empty Scripps canyon of accumulated sand \( (> \, 10^6 \, m^3) \) by a single event each year (for example). Other examples are cited, of currents \( \approx 1 \) m/s or more, duration \( O(1/2 \) hour) in the bottom 2 to 5 metres. Earthquakes may be a trigger, but their overall contribution is unknown.

In summary, canyons may be the focus of enhanced wave energy, mixing, suspended sediment and shelf-sea export, which may be transported further off-shelf to the ocean if enhanced energy levels suspend material in the water column as opposed to settling over the slope. BISCAYE and ANDERSON (1994) conceive canyons of the Middle Atlantic Bight as locations of more frequent sediment resuspension which receive material from shelf-break “events” and in turn act as sources that are steadier in time. However, there is little evidence of significant contributions to circulation or large exchanges of water. Whereas focusing affects the representativeness of measurements, it may not increase the “global” values of internal wave energy flux available as input to mixing, or the particulate matter available for export.
10. COMPARISON OF PROCESSES

We gather together the estimates from previous sections, for process contributions to circulation, exchange and energy potentially available for mixing. The numerical values derive from the given scale using “typical” values for the context as given in §0 or as found empirically in the case of certain velocity scales. It is emphasised that these estimates are not uniformly applicable: they only apply where the process or phenomenon occurs; the magnitude scales according to the controlling variables via the formula given under “Scale”.

10.1 Circulation

Table 4. Scales and estimated values of process contributions to shelf-edge circulation.

<table>
<thead>
<tr>
<th>Process</th>
<th>Scale</th>
<th>eg. m/s</th>
</tr>
</thead>
<tbody>
<tr>
<td>coastal current</td>
<td>?</td>
<td>0.1-1</td>
</tr>
<tr>
<td>slope current forced by:</td>
<td></td>
<td></td>
</tr>
<tr>
<td>JEBAR</td>
<td>$h_{o}^{2}p^{1/4}V_{pl}g/8k$</td>
<td>0.1</td>
</tr>
<tr>
<td>steady wind</td>
<td>$v/pk$</td>
<td>0.1</td>
</tr>
<tr>
<td>unsteady wind$^{(1)}$</td>
<td>$v/p\rho$</td>
<td>0.1</td>
</tr>
<tr>
<td>biased form drag</td>
<td>$(v/2p\pi) \min(1/k, v/h)$</td>
<td>0.01</td>
</tr>
<tr>
<td>wave rectification$^{(2)}$</td>
<td>$u^{2}/u_{*}^{2} \sigma^{2}$</td>
<td>0.01</td>
</tr>
<tr>
<td>eddy momentum</td>
<td>$uvb/L_{e}k$</td>
<td>0.1</td>
</tr>
<tr>
<td>western boundary current</td>
<td>$(L_{x}/L_{y})v/pk_{l}$</td>
<td>1</td>
</tr>
<tr>
<td>eddies, warm-core rings, jets</td>
<td>$\gamma$</td>
<td>0.5</td>
</tr>
<tr>
<td>tides</td>
<td>$\zeta \max { (g/h)^{1/2}, \sigma W_{g}/h }$</td>
<td>0.3</td>
</tr>
<tr>
<td>in strait to marginal sea</td>
<td>$\sigma \zeta A/hb$</td>
<td>&gt;1</td>
</tr>
</tbody>
</table>

Notes to Table 4

1) In the presence of seasonal stratification, the effective depth $h$ will be that of the upper layer down to the thermocline, $h'$ (smaller) so that the upper-layer current (only) is greater.

2) Internal motion, notably the internal tide, may reduce the effective length scale $L_{e}$ to $(g/h)^{1/2}/\sigma$ so that the rectified flow is locally greater.

Overall, there are several agents of currents ~ 0.1 m/s, but in particular contexts western boundary currents, instabilities manifested as eddies, warm-core rings and jets, and tidal currents may be very much stronger.

10.2 Exchange

Notes to Table 5.

Estimates in parentheses indicate exchange per “event” as distinct from a value per unit length of shelf.

1) An along-shelf average is estimated but cascading is liable to be localised down shelf-edge depressions.

2) Tidal flows return only 6 hours later; the exchange may be only temporary.

3) No value is typical; that given is large but may occur locally ($\sigma$).

4) Such values are localised to broad shelves such as the North Sea with an on-offshore canyon axis across the shelf width. However, canyons also facilitate many of the other processes and increase the length of shelf-edge boundary for exchange associated with slope currents and fronts.
Overall, many processes contribute $O(1 \text{ m}^2\text{s}^{-1})$ exchange, albeit distributed in different ways through the water column (which will be significant according to the constituent transported, §1.3). On this basis, internal tides (for example) are important only where exceptionally large, but upwelling should be important wherever it occurs. Larger exchanges potentially accompany boundary current divergence (although there is no evidence of values as large as tabulated), tides (but the return flow $1/2$ cycle later tends to reduce any longer-term transport, §7) and canyons locally. Perhaps surprisingly, the Gulf Stream does not in practice seem to be the cause of larger exchanges than (for example) the relatively weak slope current around Scotland.

### Table 5. Scales and estimated values of process contributions to shelf-edge exchange.

<table>
<thead>
<tr>
<th>Process</th>
<th>Scale</th>
<th>$\text{eg. m}^3/\text{s}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>slope current</td>
<td>$kV/\tau$</td>
<td>1</td>
</tr>
<tr>
<td>eg. Atlantic inflow Malin-Lewis</td>
<td></td>
<td></td>
</tr>
<tr>
<td>total Scottish slope</td>
<td></td>
<td>1</td>
</tr>
<tr>
<td>topographic irregularities</td>
<td>$V\Delta h$</td>
<td>1</td>
</tr>
<tr>
<td>eddy</td>
<td>$\frac{\partial h}{\partial x}$</td>
<td>1</td>
</tr>
<tr>
<td>warm-core ring streamer</td>
<td></td>
<td></td>
</tr>
<tr>
<td>aggregate (Middle Atlantic Bight)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>impulsive wind</td>
<td>$\tau/\rho f$</td>
<td>1</td>
</tr>
<tr>
<td>upwelling - wind</td>
<td>$\tau/\rho f$</td>
<td>1</td>
</tr>
<tr>
<td>- div. W boundary current</td>
<td>$2h_o^2V_w \partial_z(V_w/\partial_z h_o)\text{div}/f$</td>
<td>20</td>
</tr>
<tr>
<td>jets (narrow-shelf upwelling areas)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>aggregate</td>
<td></td>
<td></td>
</tr>
<tr>
<td>cascading$^{(1)}$</td>
<td>$(0.6)^{-1}(\rho\alpha/\rho c_p)^{2/3}h(H^2/W_s)^{1/3}/f$</td>
<td>0.25</td>
</tr>
<tr>
<td>front</td>
<td>$\alpha^2(g'h')^{1/2}$</td>
<td>0.3</td>
</tr>
<tr>
<td>eg. along isopycnals, Middle Atlantic Bight</td>
<td></td>
<td>0.2</td>
</tr>
<tr>
<td>tides$^{(2)}$</td>
<td>$\sigma\xi W_s$</td>
<td>10</td>
</tr>
<tr>
<td>strait to marginal sea</td>
<td>$\sigma\xi A$</td>
<td>($&gt;1 \text{ Sv}$)</td>
</tr>
<tr>
<td>shear dispersion ($hu = \sigma\xi W_s$)</td>
<td>$t_p u h_u/\sigma_T$</td>
<td>0.1</td>
</tr>
<tr>
<td>internal tide solitons$^{(3)}$</td>
<td>$&lt;\zeta&gt;\lambda/\text{tide}$</td>
<td>1</td>
</tr>
<tr>
<td>waves' Stokes drift</td>
<td>$0.01w^2$</td>
<td>1</td>
</tr>
<tr>
<td>W boundary current and bend</td>
<td>$(h/h_o)^2(L_x/L_y)\tau/\rho f$</td>
<td>($1/2 \text{ Sv}$)</td>
</tr>
<tr>
<td>slope current and bend</td>
<td>$(V_{0L}/L_T)\tau/\rho f$</td>
<td>($0.01 \text{ Sv}$)</td>
</tr>
<tr>
<td>cape eddy</td>
<td>$bv_{L_T}$</td>
<td>($0.1 \text{ Sv}$)</td>
</tr>
<tr>
<td>canyon return flow$^{(4)}$</td>
<td>$b\tau/\rho k$</td>
<td>10</td>
</tr>
<tr>
<td>ridge-associated upwelling</td>
<td>$(2\Delta h/h_o)\tau/\rho f$</td>
<td>1</td>
</tr>
</tbody>
</table>
10.3 Energy potentially available for mixing

Table 6. Scales and estimated values of process contributions to energy potentially available for mixing.

<table>
<thead>
<tr>
<th>Process</th>
<th>Scale</th>
<th>eg. mW/m²</th>
</tr>
</thead>
<tbody>
<tr>
<td>surface waves</td>
<td>$1.5 \times 10^{-5} \rho g \sigma_{wa}^2$</td>
<td>150</td>
</tr>
<tr>
<td></td>
<td>or $5 \times 10^{-7} \rho w^3$</td>
<td>500</td>
</tr>
<tr>
<td>wind</td>
<td>$\tau v$</td>
<td>10</td>
</tr>
<tr>
<td>internal tides&lt;sup&gt;(1)&lt;/sup&gt;</td>
<td>$\rho g \zeta^2 \lambda L_r$ per tide</td>
<td>50</td>
</tr>
<tr>
<td>internal waves</td>
<td>$0.1 \times 1 , kW/m / L_r$</td>
<td>10</td>
</tr>
<tr>
<td>bottom-reflected internal waves</td>
<td>$fn(h, f/N) \times 30 , mW , m^2 , flux \downarrow$</td>
<td>1</td>
</tr>
<tr>
<td>bottom friction</td>
<td>$\rho C_D v^3$</td>
<td>3</td>
</tr>
<tr>
<td>tidal&lt;sup&gt;(2)&lt;/sup&gt; (currents 0.3 or 0.7 m/s)</td>
<td>$\ldots$</td>
<td>100 or 1000</td>
</tr>
<tr>
<td>canyon-intensified internal waves&lt;sup&gt;(3)&lt;/sup&gt;</td>
<td>$&lt;\rho C_D u^3&gt;$</td>
<td>150</td>
</tr>
</tbody>
</table>

Notes to Table 6.

(1) No value is typical; that given is large but may occur locally (§7).
(2) These dissipation estimates correspond respectively to an average and locally greater values for the north-west European shelf (Flather, 1976).
(3) This estimate is local to the canyon floor where an empirical value 0.5 m/s is supposed for the intensified internal wave current.

These values highlight the importance of waves for surface mixing, of internal motions for mixing in the interior, and the highly variable importance of tidal currents and internal waves near the bottom, according to context. The rates are of energy potentially available, before reduction by an efficiency factor for conversion to potential energy. [This efficiency is usually small, eg, an estimated 5.6% (Stigebrandt and Auri, 1989) for internal tide mixing distributed between the thermocline and sea bed].

10.4 Relation to context

The estimates in §§10.1-10.3 depend on the context (e.g. shelf width $W_s$, depth $h$, slope $h_s$, latitude through $f$, seasonality/N, winds through $\tau$, waves, etc.); the dependencies differ (eg, some estimates increase with $W_s$ and others do not). Explicit dependence on capes, canyons etc. is discussed in §9. Therefore the relative importance of different processes differs according to context. There is no one ranking of processes in order of importance. Paradoxically, the common estimate $\sim 1 \, m^2/s^1$ for exchange may assist ranking in a particular context where there is a departure from typical values. Wind-driven exchanges vary as $\tau/f$, for example, and are larger in equatorial regions (for a given wind stress and appropriate direction) while tidal exchanges are even larger for a wider shelf.
11. DISCUSSION AND SCOPE FOR FUTURE WORK

11.1 Non-conservative processes

Although it was emphasised in §1.1 that common physical processes underlie all transports, nevertheless individual constituents can show distinct effects. In general, these may show in the transport equation as relative velocity (sinking, rising or swimming), sources or sinks. Any suspended material denser than the sea-water will descend relative to the ambient water, with a speed increasing with density difference and particle size; aggregation to flocs causes faster sinking and may be favoured by moderate flow conditions. Sediments, being denser than sea-water, are particularly sensitive to the presence and character of the sources; their presence in the water column depends on generation above (biological or through precipitation) or sufficient bottom currents and an erodible bed upstream; the latter in turn depends on sedimentation "history".

11.2 Shelf-sea budgets

A given rate of exchange across the shelf edge will have a greater effect on the waters of a narrow shelf (western Iberia, for example) than for a wide shelf (such as the Celtic Sea SW of Britain). 1 Sv/1000km suffices to replace shelf water of average depth 86 m to a width of 1 km in a day. 100 km width having the same average depth would take 100 days to replace at this rate. In practice, however, replacement across distances O(100 km) or more is unlikely to be direct; processes rarely cause such extensive cross-shelf excursions of water in one movement. The aggregation of several movements, as a pseudo-diffusion (cf. §1), is a more likely cause of exchange on such scales. Then

\[
\text{cross-shelf exchange time} \approx \frac{(\text{shelf width})^2}{\text{effective diffusivity}}
\]

This rapid increase with width gives broader shelf seas an interior, across which transports should be considered in distinction to the effectively separate oceanic and coastal boundaries.

The extension of fresh-water effects from the coast to the shelf edge is subject to the same considerations.

A distinction must be drawn between excursions of water (as above) and the extent (length scale) of "events". MONACO, COURP, HEUSSNER, CARBONNEZ, FOWLER and DENIAUX (1990) found near-simultaneous occurrences of large sediment flux on the shelf and slope (for example, in association with winter storms in the western Gulf of Lions). Such occurrences can be explained by the large extent of storm-forced flow over the shelf and slope, without assuming the sediment to have traversed the area during one event.

11.3 Future needs for research

Individual processes are discussed briefly in turn; an overall view is taken in §12.

Coastal-trapped wave model calculations hitherto are limited, assuming zero friction, zero beta and sub-inertial frequencies for trapped waves harmonic in time. In particular, they do not tackle the form of "natural" wave motion at super-inertial frequencies (HUTHNANCE, 1989) although there is evidence of nearly-trapped edge waves, continuity of forms through the inertial frequency and the possibility of bottom-trapped waves. Thus outstanding problems are: to propose criteria (eg. minimal radiation) for natural wave modes in the dissipative and radiative stratified slope context (HUTHNANCE, 1989) and to determine the waves' form and generation.
Dominant dynamical processes of slope-current generation have not yet been fully resolved in different contexts (California, S America, W Australia, E Atlantic). Distinction may be sought via cross-slope structure, seasonality and relation to the forcing context. There is a need to determine relationships between the slope current and cross-slope transports, involving consideration of cross-slope structure and estimates of the various forcing contributions (not only the transports), the ageostrophic component of the dynamics and the role of the bottom boundary layer when forcing, alongslope gradients or time-dependence are factors. There is also a need to allow for planetary vorticity $\beta$, and to determine the dependence of slope-current stability on $\beta$ and on any flow reversals (slope currents may be unstable, with effects on the adjacent eddy field). There is scope for investigation of these questions using a model with bathymetry uniform along the shelf.

It is not clear whether the slope current’s ubiquity implies that a particular parcel of water travels far along the slope; the circulation might perhaps comprise a number of (plan-view) cells with limited continuity between. Observations off Iberia show that there may be quite considerable variation in water masses at very small along-slope separations. New measurement arrays and perhaps the use of drifters are wanted to test slope-current continuity, distinguishing dynamical, transport and water-mass aspects. A concerted effort is being made under the European Community-funded project SEFOS (Shelf-Edge Fisheries and Oceanography Study) to assemble these measurements and study slope-current continuity around the west European shelf edge from Iberia to Norway.

These considerations regarding the slope current and associated transports suggest some uncertainty and scope for exploring the consequences of varied eastern boundary conditions on the idealised thermocline model for ocean circulation, particularly in relation to cross-slope flows and the nature of vertical exchange at the eastern boundary.

The magnitude, space and time scales of up/down-welling variability are rather poorly known at present, especially for cross-slope flow; the need is for measurements over long periods for improved statistics. Cascading, in particular, tends to have been inferred rather than observed directly. Time series for many years are needed, to determine seasonal cycles and variability for currents over the shelf and slope. Both different seasons and long records will help to gain the desired wide range of measured conditions: so will varied locations, which can also complement drogued buoy deployments to investigate slope-current continuity. It is not clear at present, what are the limits to the growth of eddies and filaments which may be so important locally to cross-slope exchange.

There is no general formulation for predicting the occurrence of fronts, taking account of both buoyancy factors (surface heating, lateral freshwater input) and stirring both at the surface (by wind and waves) and at the bottom by tides.

Work is under-way to develop a 3-D internal tide and wave model extending the 2-D approach of SHERWIN and TAYLOR (1989). However, there are few moored time-series of currents in sufficient density to enable a rigorous test of such models. Exceptions are perhaps ROSENFELD (1990; 50 current meters in CODE) and moored ADCP data from the Iceland-Faeroes ridge (PERKINS, SHERWIN and HOPKINS, 1994).

There appears to have been no concerted effort to use moored ADCPs to study the detailed form of an internal tide at a shelf break where it should be most deterministic. There also remains scope for in situ sub-surface measurements coincident with ERS SAR imagery. Requirements are (a) to estimate variability in time and space of the internal wave field, the location and extent of the regions of highest shear and mixing, (b) to estimate wave drag on larger-scale currents, entailing a good survey of slope topography, and (c) intensive measurements over a 15-day spring-neap tidal cycle at different seasons, for an energy budget (evaluating energy fluxes as a vertical integral of pressure x velocity) for shelf-edge exchange.
12. CONCLUSION

Flux estimated naively from measurements as $\langle uC \rangle$ is uncertain, owing to the need for a comprehensive yet intensive array in space and time. This derives from the geostrophic constraint on overall cross-slope flow, and hence a tendency for cross-slope flow to be relatively small in magnitude, with small time and space scales for structure and coherence. It is also necessary for the array to detect small correlations of $u$ and $C$.

An approach through process understanding is therefore suggested. This paper has reviewed present knowledge of processes' contributions to fluxes. On this basis, a classification of shelves might be attempted, firstly on local physical grounds according to shelf width, depth etc. An assessment of processes' global contribution should then take into account the length of shelf where a process is important.

Processes interact, and the ocean margin topography is complex. Therefore, it seems that numerical models must eventually be invoked to provide the sought-after synthesis over processes, time and space.

Measurements are essential to test models. The sequence of hypothesis (embodied in a model) suggesting experimental arrays, and measurements in turn causing model revision, is the essence of the scientific method. In the shelf-edge context, the choice of the model area and the need to initialise the model place a critical demand on measurements around the boundary of the experiment. Other competing demands on the experimental array are the need for duration to provide statistics of intermittent processes, versus detail to define their form; a compromise may be detailed measurements for a shorter duration to add value to measurements from a sparser long-term array.

Water mass analysis provides a valuable complement to process studies in the form of an integrated (but uncontrolled) measure of cross-slope transport. Drogued buoys may provide an intermediate Lagrangian view, less integral but more controlled through the choice of deployment times and locations.

Measurements may be wanted in a variety of locations so that individual processes are well-developed to test their representation by models; yet there is a case for locations to be representative of significant lengths of ocean margin.

The developing interest in processes of shelf-ocean exchange, and the relationship between the nature of ageostrophy and the cross-slope structure of quasi-geostrophic flow, are reducing model resolution and the accompanying scale of required measurements to the internal deformation radius $R_I$ and in the vertical to an emphasis near the sloping sea-floor.

Model requirements on shelf-wide scales have been discussed elsewhere (HUTHNANCE, 1992). For finer-scale models, there are additional concerns. The model coordinates and advection scheme need to address simultaneously (i) the distribution of stratification and bathymetry such that the maximum of $\nabla \nabla h$ over the sea bed can be described (ii) advection over the sloping bed (iii) dynamical balance of the JEBAR term and of initialising data.

Although fine resolution is needed, shelf-edge models need to take account of phenomena in a large area; the extent of influence from ‘upstream’ may be 1000 km. This may be through open boundary conditions or embedding in a wider area model. Either there are technical problems to be addressed in making the wider area and nested models (two-way) interactive, or there is an intellectual challenge to finding ‘off-line’ open-boundary or matching conditions that are not interactive yet satisfy the needs of both the wider-area and fine shelf-edge models. One factor which may assist is that the distance 1000 km relates to the decay distance for a first mode coastal-trapped wave; if the large-area model has a good approximation to this (a relatively easy requirement on
gross stratification and/or shelf/ocean depth ratio and shelf-slope widths) then the effect of poor boundary conditions in the fine model may relate only to higher modes and penetrate a much shorter distance before decay.

Hence there remains full scope for the application of the scientific method through models and experiments - the latter at sea, in the laboratory and with simpler models. The emphasis is on fine scales $R_f$ and near the sea bed, on the model grid in relation to stratification, bathymetry and advection, and on accounting for external influences in both experiment and model. In view of the prospective application of models, we have not attempted to be fully quantitative in estimating circulation, exchange transports and mixing rates. Rather, the aim has been to provide a basis for deciding what may be the important features and physics to include in a model of a chosen region, and the processes to be resolved in measurements to test models.

13. ACKNOWLEDGMENTS

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14. APPENDIX

A. Equations of motion

The equation for horizontal momentum of hydrostatic flow $(u, w)$ may be written

$$(f+\omega)k \times u = -u_{\tau} - (wu)_{z} - u \nabla u - g \nabla \eta + \frac{1}{2} \nabla (u^2+w^2) + \rho^2 \nabla \cdot F + (\tau/p)_{z} - (g/p) \nabla f \rho dz$$

where $f$ = Coriolis parameter, $\omega$ = vertical component of relative (flow) vorticity, $k$ = vertical unit vector, $g$ = acceleration due to gravity, $\eta$ = surface elevation, $\rho$ = density, $F$ = lateral stress (tensor), $\tau$ = vertical stress and subscripts denote differentiation. The continuity equation is

$$\nabla \cdot u + w_{\tau} = 0.$$  

Averaging the momentum equation through the water depth $h$ gives

$$h^{-1} \int_{h}^{0} (f+\omega)k \times u dz = -h^{-1} \nabla [\chi - \eta] - \nabla [g \eta + (g/p) \int_{h}^{0} (\rho - \eta) dz]$$

$$+ (\tau ph)_{h} - [wu/h]_{h} + h^{-1} \int_{h}^{0} \{-u_{\tau} + \rho \nabla \cdot F - u \nabla \cdot u\} dz$$

where $\chi = (g/p)_{h} \rho dz$, $nl$ = non-linear terms deriving from $\frac{1}{2}(u^2+w^2)$. All but the first two terms on the right-hand side represent ageostrophic effects; moreover, the second term is removed by integration around any closed contour:
\( \oint (f+\omega)u\,\text{d}s = -\oint h^1\text{d}l\nabla[\chi \cdot n] \)
\[
\quad + \left[ \oint (ph)^+\tau\cdot\text{d}l \right] \n + \left[ \oint h^1\omega u\cdot\text{d}l \right] \n + \oint \left\{ -\frac{\partial u}{\partial t} + \rho^{-1} \nabla \cdot F \right\} \nabla u \nabla u \]
where the overbars denote a depth average. Note that the density field (through \( \chi \)) and wind stress (through \( \tau \) at the surface) give separate contributions to the cross-contour flow (even if the density gradients ultimately result from wind forcing). The first term on the right-hand side is zero if the integration is around a depth contour; net cross-contour flux is entirely attributable to ageostrophy.

**B. Upwelling depth**

We use linear hydrostatic inviscid equations for flow \((u,v,w)\) uniform in the alongshelf direction \(y\) (\(x\) is onshore with the coast at \(x=0\); \(z\) is vertically up with the sea surface at \(z=0\)):

\[
-fv = -\rho \frac{\partial p}{\partial y}; \quad \nabla \cdot u = 0; \quad \rho = -\rho_0 g; \quad u_z + w_x = 0; \quad \rho - N^2 \rho_0 w/g = 0
\]

Here \(f\) is the Coriolis parameter, \(p\) the perturbation pressure, \(\rho\) is the perturbation density (mean \(\rho_0\)), \(t\) is time, \(g\) is the acceleration due to gravity, \(N\) is the buoyancy frequency and subscripts denote differentiation. Flow quantities can all be related to \(p\):

\[
v = \frac{\rho}{\rho_0} f; \quad p = -\frac{\rho}{g} \quad w = -\rho_0 \frac{N^2}{\rho_0} \quad u = -\frac{\rho}{f^2} \rho_0.
\]

Then

\[
\{p_{xx} + f^2(p_z/N^2)_{zz}\} = 0 \quad \text{or} \quad u_{xx} + f^2(u_z/N^2)_{zz} = 0.
\]

These are to be solved for flow into \(x=0, z=0\) to replace surface Ekman transport offshore.

(a) *Infinite depth and uniform stratification.* Expanding the vertical scale by \(N/f\), we have to solve

\[
p_{xx} + p_{zz} = 0
\]

or

\[
(rp_z)_r + p_{zz} = 0 \quad \text{in radial co-ordinates} \quad (x = r \cos \theta, z = r \sin \theta).
\]

The solution is radial flow into the corner:

\[
(u,w)/U = -(x,z)/r^2
\]

so that the ratio of horizontal and vertical scales is \(N/f\) (buoyancy/Coriolis frequency). There is no absolute length scale in this idealised problem.

(b) *Finite depth \(H\) and uniform stratification.* We require \(0 = w \sim p_z\) on \(z = 0, -H\), and boundedness as \(x \to -\infty\) in the ocean. Separable solutions for \(p_t\) or \(u\) are then

\[
\exp(n\pi fx/NH) \cos(n\pi z/H) \quad (n = 0, 1, 2, \ldots)
\]
The full solution is a linear combination of these, to satisfy $u = 0$ on $x = 0, z < 0$. However, only the $n=0$ component ($u$ uniform in depth and offshore) contributes a net inflow towards the coast. The other components decay away from the coast (rapid decay for large $n$) and serve only to direct the inflow upwards to the corner $x=0, z=0$. Clearly the vertical scale is $H$, and the horizontal scale for redirection of the flow to the corner is $HN/f$.

(c) **Uniform stratification $N_1$ in depth $H_1$ above uniform stratification $N_2$ in depth $H_2$.** We require boundary conditions as (b) but also $u$ and $u_z/N$ continuous at $z = -H_1$. Thus separable solutions are

$$
\exp(\lambda_n x) \cdot \begin{cases} 
U_1 \cos(\lambda_n N_1 z/f) & 0 > z > -H_1 \\
U_2 \cos(\lambda_n N_2 (z+H_1+H_2)/f) & -H_1 > z > -H_2-H_1,
\end{cases}
$$

where

$$-N_2 \tan(N_1 \lambda_n H_1/f) = N_1 \tan(N_2 \lambda_n H_2/f).$$

Again the full solution is a linear combination of these separable solutions, of which only that with $n = 0 = \lambda_0$ ($u$ uniform in depth and offshore) contributes a net inflow towards the coast. The other components decay away from the coast with rate $\lambda_n \sim n\pi f/(N_1 H_1 + N_2 H_2)$ and again serve to direct the inflow upwards to the corner $x=0, z=0$. In particular, the $n=1$ component has effect furthest offshore, directing much of the inflow into the upper layer if $N_1 H_1 > N_2 H_2$ at $-x$ of the order of $H_1 N_1/f$.

**C. Shelf-canyon exchange**

Consider the $x$-momentum balance towards the head of a deep embayment, depth $h(y)$:

$$0 = -g \zeta_x - ku/h + \tau/\rho h$$

where $g$ is gravitational acceleration, $\zeta_x$ is the surface slope, $k$ is the (linear) bottom drag coefficient, $u$ is the velocity component into the embayment, $\tau$ is the wind stress (supposed uniform) and $\rho$ is the density. $\zeta_x$ is determined by the constraint $\int h dy = 0$ (no net flow into the embayment). Then the local transport is $hu = (ht/\rho k) \{ 1 - h(\zeta_x dy)/(h^2 dy) \}$. If $h(y)$ is $h_1$ over a width $w_1$ of embayment and $h_2$ over a width $w_2$, then the shelf-canyon exchange is

$$w_1 w_2 h_1 h_2 (h_1-h_2)(w_1 h_1^2+w_2 h_2^2)^{-1/2} \tau/\rho k$$

over a sector $w_1+w_2$ spanning the embayment. If the canyon ("1") is relatively deep and not too narrow ($w_1 h_1^2 >> w_2 h_2^2$), this simplifies to $w_1 h_1 \tau/\rho k$, i.e. the value for steady wind-driven transport on the shallow shelf, constrained only by bottom friction.
15. REFERENCES


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