Cross-shelf exchange

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Abstract:

Cross-shelf exchange dominates the pathways and rates by which nutrients, biota and materials on the continental shelf are delivered and removed. These transports are limited by Earth's rotation, which inhibits flow from crossing isobaths. Thus, cross-shelf transports are generally weak compared to alongshore flows, and this leads to interesting observational issues. Cross-shelf flows are enabled by turbulent mixing processes, by nonlinear processes (such as momentum advection), and by time-dependence. Thus, there is a wide range of possible effects that can allow these critical transports, and different natural settings are often governed by differing mixes of processes. Examples of representative transport mechanisms are discussed, and possible observational and theoretical paths to future progress are explored.

1. Introduction

Cross-shelf exchange is arguably the central problem in coastal physical oceanography. The more energetic alongshore currents often dominate flow fields on time scales longer than about a day, but strong cross-shelf gradients in temperature, salinity or dissolved materials mean that even weak cross-shelf currents have greater impact. Cross-shelf exchanges, in connection with coupled vertical transports, thus deliver nutrients to shelf ecosystems, govern residence times, transport planktonic life stages, enable sediment transports, remove contaminants from the coastal region, and are central in determining biogeochemical budgets. And this is an incomplete list.

Although cross-shelf exchange has been recognized as a central concern for decades (e.g., Allen et al., 1988), progress has been slow for reasons that pervade this review. It is indeed striking to contrast the rapid theoretical progress in understanding alongshore currents during the 1970s (e.g., Huthnance, 1973; Pedlosky, 1974; Gill and Schumann, 1974; Csanady, 1978; Allen, 1980) with the relative difficulty met in finding similarly sweeping syntheses for the dominant components of cross-shelf flow. The cross-shelf problem, of course, does not have a single answer. Rather, there is a range of phenomena involved, and, for each location or time, there occurs a different mix of processes.

In the following, some attention is given initially to explaining, both theoretically and observationally, why cross-shelf exchange is a difficult topic. Next, a range of examples of exchange mechanisms are studied, organized by the physical effects that allow them to operate. Finally, some thoughts are noted about future progress. Given the breadth of the subject material, this review is necessarily incomplete, and earlier contributions, such as that of Huthnance (1995) will help fill the many gaps.

2. The difficulty

Three of the traditional physical oceanographer's most-treasured assumptions about the ocean are 1) that the flow field changes slowly with time, 2) that nonlinear effects (such as the transport of momentum by the flow itself) are not important, and 3) that turbulent stresses and mixing are not too important. The consequences of these assumptions are twofold. First, the velocity is geostrophically balanced, i.e., the flow is strictly parallel to contours of constant pressure. Second, density variations are governed mainly by the flow field and not primarily by mixing.

If one further assumes that the spatial region of interest is small compared to Earth's radius (so that the effective rotation rate is uniform), then the constraints are remarkably powerful (e.g., Brink, 1998). First, the vertical component of velocity is zero everywhere, which in turn requires that any flow near the bottom has to follow isobaths exactly. Second,

$$0 = \boldsymbol{k} \cdot \boldsymbol{v} \times \boldsymbol{v_z} \tag{1}$$

where $\boldsymbol{k}$ is a vertical unit vector, $\boldsymbol{v}$ is the horizontal velocity vector, and $\boldsymbol{v_z}$ is the vertical derivative of the velocity (i.e., the vertical shear): the vertical shear is parallel to the velocity. In other words, the velocity cannot veer with depth as long as $\boldsymbol{v} \cdot \boldsymbol{v_z}$ is nonzero. (There can be variations in the flow direction at depths where $\boldsymbol{v} = 0$). One striking thing about the rule (1) is that it holds even if the ocean is stratified, i.e., if the density varies with depth. Finally, combining (1) with the requirement that near-bottom flow follows isobaths means that (again, unless $\boldsymbol{v} = 0$ or $\boldsymbol{v_z} = 0$ at some depth), flow at all depths has to follow isobaths. This result is the well-known Taylor-Proudman (e.g., Pedlosky, 1979) theorem.

To an oceanographer interested in flow over the continental shelf or slope, the Taylor-Proudman theorem usually has the disturbing implication that, at lowest order, there cannot be flow across isobaths. That is, there can be no exchange between the coastal and open oceans. Of course, there is ample evidence that these exchanges do occur in nature: shelf water properties are often similar to those offshore. For example, even when shelf and oceanic waters differ in salinity, it is rarely by more than a few g/kg.

So, what happens in reality? The assumptions going into the Taylor-Proudman theorem are reasonable under a wide range of circumstances, so only mild violations of the three key assumptions (steadiness, linearity and adiabatic physics) occur and thus enable only relatively weak cross-shelf transports. Stated another way, the cross-isobath transport is often a secondary flow that occurs in the shadow of more energetic alongshore flows.

The component of flow we care most about (i.e., the cross-shelf velocity u) is thus by nature weak compared to more energetic alongshore flows. This leads to all sorts of interesting problems. For example, say that the approximate alongshore velocity is v and one wants to define a new coordinate system that better approximates the local alongshore orientation. Then the new alongshore and cross-shelf velocity components (u', v') are given by

$$u' = u \cos\theta + v \sin\theta \tag{2a}$$
$$v' = -u \sin\theta + v \cos\theta \quad . \tag{2b}$$

Assuming that the directional correction θ is small, and that $|v| >> |u|$, then $v' \approx v$: the exact alongshore flow is unaffected by small directional errors. On the other hand, the same assumptions lead to the conclusion that the new cross-shelf velocity component u' depends sensitively on u, v and θ. Thus, estimates of cross-shelf velocity, hence cross-shelf exchange, are very sensitive to the local coordinate system (e.g., Smith, 1981): errors of only a few degrees in defining the true alongshore direction can lead to changes even in the sign of u'.

A dominant alongshore flow can cause other problems for the observer. Say, for example, that free-floating drifters are used to study a cross-shelf flow pattern. The strong alongshore flow guarantees that the drifter will be carried far alongshore before much can be learned about local cross-shelf flows. Other difficulties for the observer will be mentioned in context below.

Thus, we seek to understand a flow component that is weak, often ageostrophic, and difficult to observe in nature. The following discussion is driven by the idea that violations of the Taylor-Proudman assumptions enable cross-shelf exchange. The goal is not to summarize all of the many known manifestations of cross-shelf transport, but rather to discuss the three key violations and how they play out in nature. Some thoughts about future progress are then offered as a conclusion.

3. Mixing and dissipation

3.1 Wind-driven upwelling

In reality, turbulent mixing and stress transfer are often important factors in parts of the water column. Most notably, in the upper 10s of meters, turbulence is often driven by applied winds, surface cooling, breaking surface waves, and strong vertical shears. The net effect is a near-surface boundary layer (the Ekman layer) where, for reasonably steady flows with negligible momentum advection, the directly wind-driven cross-shelf transport (i.e., depth-integrated cross-shelf velocity) is

$$U_{ES} = \tau_0^y / (\rho_0 f) \qquad (3)$$

where τ_0^y is the alongshore component of the wind stress, ρ_0 is the water's density and f is the Coriolis parameter, a local measure of Earth's effective rotation ($f = 2\Omega \sin \varphi$, where Ω is the angular frequency of Earth rotating about its axis and φ is the latitude). This cross-shelf transport needs to be balanced, at least near the coast if not over the whole shelf, by an onshore, cross-isobath flow, presumably deeper in the water column. When the wind drives an offshore near-surface flow, the compensating flow at depth generally carries water that is cooler and more nutrient-laden than the near-shore, near-surface water it replaces. This cold, upwelled water leads to highly productive ecosystems (e.g., Mackas et al., 2006) in upwelling regions such as the U.S. west coast, the Peru-Chile system, the Benguela, and off northwest Africa.

The above basic explanation of these productive systems dates back to Ekman (1905) and Thorade (1909) and it represents one of the great early successes of dynamical thinking in oceanography. The above explanation is, of course, incomplete. It does not address questions such as the dynamical balances that control the onshore flow: time dependence, geostrophic contributions and/or nonlinear momentum transport (e.g., Lentz and Chapman, 2004) might all be important. Introducing these effects can allow entirely new aspects of the flow to appear (e.g., Clarke, 1977). Further, this basic view is expressed in terms of a rather dreamlike ocean, with straight topography, uniform winds and a smooth coastline.

The basic explanation is, however, supported by myriad observations. For example, Figure 1 shows cross-shelf sections of temperature and chlorophyll during upwelling off Oregon. Indeed, cold, nutrient-rich water occurs at the surface nearshore, and the high nutrient content leads to phytoplankton growth, hence high chlorophyll concentrations, near the surface front. Sections such as these, however, fail to emphasize aspects that are poorly understood. For example, direct measurements have been made of surface Ekman velocities, but the compensating onshore flow is often poorly characterized either by observations or theory. There are many good reasons that this happens. One is the uncertainty in coordinate rotation, mentioned above (Smith, 1981). Another is the apparent ubiquity of small-scale (5-10 km) eddies or meanders over many continental shelves that are evident because of short correlation length scales for cross-shelf currents (e.g., Kundu and Allen, 1976; Winant, 1983; Dever, 1997). These mask weaker, larger-scale interior cross-shelf flows. The smaller scale features are evidently related to baroclinic instability of the wind-driven alongshore flow (Brink, 2015; Brink and Seo, 2015) which is surely active in many upwelling regions (Barth, 1989 a, b; Barth, 1994; Durski and Allen, 2005) and perhaps much more widely. The underlying observational issues boil down to observing a weak cross-shelf net flow in a noisy, complex setting.

Two-dimensional (cross-shelf and vertical) models of upwelling and downwelling (e.g., Allen et al., 1995; Allen and Newberger, 1996) have led to considerable insights on the development of fronts and mixing in response to wind driving. However, they also point to fundamental difficulties. Specifically, consider a case where the modeled continental shelf, realistically, borders on a deeper, flat-bottomed ocean. In the flat-bottom outer region, offshore of any fronts or jets, it is easy to expand any theoretical solution in terms of baroclinic modes (e.g., Gill and Clarke, 1974), each of which has a finite length scale (Rossby radius) of O(50 km) or less for all modes except the barotropic. Thus, far from the coast, the two-dimensional flow field must be strictly barotropic (i.e., depth-independent). This means that onshore transport is either evenly distributed through the entire, O(1000m) deep, offshore water column in the time-dependent case, or confined to the bottom boundary layer if the flow is steady. Thus, two-dimensional models imply that upwelled water must come from depths of O(1000m), while observations (e.g., Roughan et al., 2006) typically indicate upwelling source depths of about 100-200m. One conclusion from this discomforting contrast is that two-dimensional models should not be treated as realistic for time scales longer than about the time it takes for a water parcel to move onshore by about an internal Rossby radius of deformation, i.e., about perhaps 10 days. A partial, rather artificial, resolution to the problem can be obtained by allowing an alongshore pressure gradient, but it must be uniform in the cross-shelf direction in order to maintain a two-dimensional volume balance.

The real resolution to this problem, however, is to admit that the real ocean is three-dimensional so that there can be geostrophic, spatially variable, onshore flow (Pringle and Dever, 2009, Rivas and Samelson, 2011) approaching the shelf. This, in turn implies that there is an alongshore pressure gradient over the continental slope: the gradient might be related to conditions far alongshore, to deep-ocean currents or to topographic or wind variations within the domain. Because geostrophy is a diagnostic relation, a pressure gradient should not be thought of as a forcing mechanism in any normal sense, but rather as a signature of some other effect. As Pringle and Dever (2009) point out, it is often not obvious what drives this geostrophic flow in a realistic settling.

3.2 Bottom boundary layer flows.

Just as there is a turbulent near-surface region in the ocean, there is also a layer near the bottom where turbulence is maintained through stresses associated with overlying currents. Because alongshore flows generally predominate in the coastal ocean (at least on time scales longer than the inertial), the bottom stress $\boldsymbol{\tau}_\mathbf{B}$ is expected to be directed mainly alongshore, and to be associated with a cross-shelf Ekman transport, $-\tau_B^y / (\rho_0 f)$. Indeed, a number of theories (e.g., Hsueh and O'Brien, 1971) posit that this bottom transport can be important for cross-shelf exchanges, and there is some clear observational evidence of this occurring in nature (e.g., Badan-Dangon et al., 1986).

These intriguing early theories, however, consistently ignore buoyancy transport in the bottom boundary layer, which can have some remarkable consequences (e.g., Garrett et al., 1993; Brink and Lentz, 2010). For example, if there is onshore flow across isobaths in a bottom boundary layer, the flow transports dense water upslope, and this promotes stronger stratification near the bottom (Figure 2). Similarly, downslope bottom boundary layer flow leads to a thicker boundary layer and weaker stratification near the bottom. This asymmetry has been clearly observed in nature (Lentz and Trowbridge, 1991; Figure 3). Bottom boundary layer structure, however, is only the beginning: because this stabilization occurs over a sloping bottom, it implies horizontal density gradients, which then imply a geostrophically balanced shear in alongshore flow. Regardless of the sense of flow, the shear tends to bring the near-bottom alongshore velocity to rest geostrophically, thus neutralizing the bottom stress, hence cross-shelf Ekman transport.

There are two key parameters that govern buoyancy arrest, the slope Burger number

$$s = \alpha N / f \tag{4a}$$

and a measure of the bottom drag

$$d = c_D N / f \tag{4b}$$

where α is the bottom slope, N is the ambient buoyancy frequency (a measure of the strength of stratification), and c_D is the quadratic bottom drag coefficient. Buoyancy arrest only occurs for time scales longer than the order of

$$T_{down} \approx (1+s^2)/(fds^3) \tag{5}$$

for down-slope flow. Similarly, for upslope flow with larger d,

$$T_{up}^{smooth} = (1+s^2)\Lambda(s)/(s^3 df) \tag{6 a}$$

$$\Lambda(s) = 0.5[\ (1 + a\ s^2)^{½}\ -1] \quad , \tag{6b}$$

where a is a constant (Brink and Lentz 2010). For upslope flow with smaller d,

$$T_{up}^{capped} \approx (1+s^2)\,/\,[(1+s)s^2 d^{\frac{1}{2}} f\,] \tag{7}$$

(Middleton and Ramsden, 1996). Further, arrest can only occur over alongshore distances L_B greater than an adjustment length scale that increases as d becomes smaller, but has a more complex dependence on s and the alongshore transport of the current (Chapman and Lentz, 1997; Chapman, 2000; Brink, 2012). These scales depend sensitively on the parameters, especially high powers of s, but it appears that wider mid-latitude shelves ought to be unaffected by arrest (because L_B is O(1000 km)) but lower latitude shelves, such as off Peru, could potentially feel the effects of buoyancy arrest. Over continental slopes, however, the bottom slope is sufficiently large that shorter L_B scales, of O(100 km), might be typical. In addition, numerical model results (e.g., Brink, 2011) show that, because arrest involves nonlinear scale interactions, the alongshore-averaged flow does not reach arrest in the presence of flow features, such as eddies, that have spatial scales less than L_B. The point is that, based on these adjustment time and space scale estimates, buoyancy arrest over the continental shelf may not be common, and so bottom Ekman transport would seem likely to escape buoyancy arrest over many shelves, especially for flow features with shorter alongshore scales. However, nature appears more complicated than this.

Buoyancy arrest is not the only process that can neutralize bottom Ekman transport. The other potential mechanism is stratified spindown (e.g., St. Maurice and Veronis, 1975). Over a flat bottom, where there is no buoyancy arrest, flow *above* the bottom boundary layer is brought to rest over a vertical scale of

$$O(f\,L/\,N) \tag{8}$$

over a time scale of

$$O(\ h\,/\,r\ \ \text{or}\ \ Lf\,/\,(Nr)\ ), \tag{9}$$

(whichever is shorter), where r is a bottom resistance coefficient (i.e., the linear proportionality constant between bottom stress and near-bottom velocity). The horizontal length scale L is essentially the width of the alongshore current. Under representative shelf conditions, the spindown time scale (9) might be a few days. As is the case with buoyancy arrest, a geostrophically balanced shear brings the near-bottom velocity to rest, but the difference is that buoyancy arrest occurs entirely within the bottom boundary layer, while stratified spindown occurs because of an adjustment outside the turbulent boundary layer. Chapman (2002) provides an enlightening treatment of the interplay of the two mechanisms. It is worth keeping in mind that, in unstratified regions, neither buoyancy arrest nor stratified spindown are able to mitigate bottom stresses.

Thus, on stratified continental shelves, there is conflicting evidence as to whether alongshore bottom stresses can be important over time scales longer that perhaps 10 days. What do the observations say? As mentioned, Trowbridge and Lentz (1991) showed, on time scales of days, that indeed the boundary layer thickness depends markedly on the up- or down-slope direction of the bottom Ekman transport. Probably the most convincing study of the stress itself is that of Trowbridge and Lentz (1998) over the northern California shelf. They show that indeed bottom stress, on seasonal time scales, is neutralized by buoyancy arrest, but that on time scales of a few days, the bottom stress plays an important role in the momentum balances.

Thus, some further clarification is required because theory suggests that bottom Ekman transport may rarely reach arrest, but the existing observations suggest that arrest indeed occurs. Our thinking is conditioned by the relative paucity of high-quality, long term direct bottom stress measurements: more long-term measurements of bottom stress and bottom boundary layer structure are very much needed.

3.3 Surface-to-bottom mixing

During cold wintertime conditions, waters over the continental shelf often mix from the surface to bottom. Models show that if an outward surface heat flux is uniform over the shelf, shallower waters cool more than deeper waters. This follows because surface cooling leads to vertical mixing which distributes the given surface heat loss over a smaller volume in shallower water than in deep water. Thus, a cross-shelf temperature gradient forms and baroclinic instability can lead to the formation of relatively small, O(5-10 km), eddies (e.g., Pringle, 2001). It is important to note that David Chapman visited much of the important physics of this problem in earlier studies (e.g., Chapman and Gawarkiewicz, 1997) that focused on cooling only within a finite-size coastal polynya. The eddies in turn transport heat so that cross-shelf temperature gradients lessen and vertical density stratification is maintained over the shelf.

These results are highly idealized, and do not, for example, usually include the effects of wind forcing or of irregular bottom topography. Experiments that include an isolated canyon across the shelf (e.g., Chapman and Gawarkiewicz ,1995) show that the canyon can concentrate offshore flow of denser waters into the trough, and allow an intermittent density current to form. On the other hand, Spall (2013) shows that a corrugated bottom does not lead to particularly anomalous results: there is obviously a good deal more to be understood about topographic effects on cooling-driven flows. It also seems likely that alongshore winds can affect eddy generation over the shelf. This follows because upwelling or downwelling influence isopycnal slope, hence baroclinic instability. This subject is currently a topic of active research.

The short spatial scales of wintertime shelf eddies raise the same sorts of sampling difficulties as do the summertime eddies implied by the work of Kundu and Allen (1976). While the wintertime eddies may complicate observations, they also represent a promising mechanism for cross-shelf exchanges. To my knowledge, no one has made the sort of complete observations that would test the sort of cooling-driven transport mechanism proposed by Chapman and Pringle.

3.4 The inner shelf

There are numerous ways to define the “inner shelf”, and the different definitions highlight different aspects that distinguish this region from the mid-or outer shelf. One very useful definition is the region offshore of the surf zone, but where the surface and bottom turbulent boundary layers merge (e.g., Mitchum and Clarke, 1986; Lentz and Fewings, 2012). This definition implies that there is no relatively adiabatic interior region, and inner shelf observations (e.g., Lentz et al., 1999) show that the bottom stress usually enters the momentum balances at lowest order. Further, because of proximity to the coastal boundary, cross-shelf velocities are expected to be relatively small (e.g., Lentz and Fewings, 2012).

While the inner shelf is subject to the same sorts of effects as on the mid-shelf, such as forcing by alongshore winds, new mechanisms for cross-shelf exchange also occur in this region. For example, surface Ekman transport is strongly affected by the bottom stress. In deep water (where there is a non-turbulent interior), the net Ekman transport is at right angles to the applied stress, although there can be some veering with depth within the layer. When the entire water column is part of the turbulent boundary layer, there are weakened transverse Ekman transports associated with the surface and bottom stresses as well as a net down-wind flow component. This downwind flow makes it possible for steady cross-shelf wind stress to drive cross-shelf transport, a phenomenon that would not usually occur in deeper water (e.g., Tilburg, 2003). These theoretical notions have been confirmed using long-term observations by Fewings et al. (2008), who found typical exchange velocities associated with cross-shelf winds to be about 2 cm/sec. While these flows are relatively weak, there is only a limited cross-sectional area (roughly half of the water depth times the distance from shore) onshore of their observation site, so such a flow could still flush the inshore waters in O(1 day).

Another mechanism that is important on the inner shelf is undertow (Lentz et al., 2008). Physically, onshore surface wave propagation causes a Stokes drift (particle transport associated with the correlation of free surface height and wave orbital velocity) toward the coast. At the same time (Hasselmann, 1970), the Coriolis force acting on the wave particle motions, averaged over many wave periods, creates a net force over most of the water column. The net effect is a relatively depth-independent offshore flow of again O(2 cm/sec) (Lentz et al., 2008). Again, this seemingly weak flow would appear to be very effective at flushing near-shore waters, but the Lagrangian (particle-following) flow field is more complex, and so net inner shelf flushing times are still poorly understood.

Both of these mechanisms cause relatively weak flows, but sustained long-term observations at a coastal observatory allow a thoughtful data analysis to extract the important signals during favorable conditions. While there is still much to be learned about inner shelf circulation, especially in terms of alongshore variability, the recent progress has been very heartening.

4. Inertial Effects

4.1 Mesoscale structures.

Perhaps the most dramatic and evocative exchange pathways are those associated with energetic flow structures, such as offshore eddies, that drive flows across shelf edges. Either eddies or coherent currents might be involved, but the common thread is that a highly nonlinear (in the sense of the Rossby number: the magnitude of relative vorticity divided by planetary vorticity) flow is involved in a major shelf-edge disturbance.

For example, the energetic, northward-flowing Kuroshio encounters the shelf edge trending east of Taiwan and frequently crosses temporarily onto the shelf before returning offshore (e.g., Chern, et al., 1990; Vélez-Belchí et al.; 2013). Hsueh et al., (1996) provide a mechanistic explanation of these intrusions, while Chern et al. (1990) provide evidence that at least some of the Kuroshio water mixes with ambient shelf waters, thus allowing a lasting impact of the intrusion. This mixing is a crucial step in the process because if offshore water simply flows onto the shelf and back off again, it allows no net onshore flux of water properties.

In terms of seaward transport, there is clear evidence that Warm-Core Rings (100 km-scale mesoscale eddies) offshore of the shelfbreak draw fresher shelf waters offshore into the ambient ocean off the northeastern United States (e.g., Joyce et al., 1992; Lee and Brink, 2010). These offshore "streamers" can penetrate down to at least 100m depth and have instantaneous volume fluxes of O(1-10 $\times$ 10^4 m^3 /sec). These vigorous transports suggest the possibility that these eddy-driven withdrawals could be very important for the regional offshore flux of shelf water (see Lentz, 2010, for current estimates of the required offshore fluxes), but these eddy events remain poorly characterized in terms of their frequency and duration. Nonetheless, Chaudhuri et al. (2009) used remotely sensed data to estimate a time- and space-averaged offshore transport (per m of coastline) of order 0.1 m^2 /sec due only to these features, a number compatible with total offshore flux estimates by Brink (1998) or Lentz (2008). Thus, Ring-driven withdrawal is likely to be an important mechanism, but more refined estimates certainly are desirable. It seems likely that similar events may also occur in other regions, such as off southeastern Australia or northeastern Japan, where highly energetic eddies also occur offshore of the shelf.

A third example of exchanges driven by offshore structures occurs at locations where a western boundary current separates from the shelfbreak and turns offshore. A well-known example is the fresh "Ford water" (Ford, et al., 1952) that is drawn off the shelf near Cape Hatteras and maintains its identity as narrow fresh filaments within the north wall of the separated Gulf Stream. Recent evidence (Guerrero et al., 2014) also suggests that comparable features are likely to exist off the eastern coast of South America where the Brazil Current also turns offshore. As is the case with offshore eddies, there have been questions as to how effective these withdrawals are in terms of shelf water balances, but focused field programs (e.g., Savidge and Savidge, 2014) are shedding a good deal of light on this subject.

4.2 Instabilities

In many cases, horizontal variations in velocity or density can become so strong that a steady flow pattern breaks down into chaotic eddy fields, i.e., instability sets in. The virtually random motions associated with an extensive, developed eddy field can be strong enough to play a prominent role in exchange across the initial gradients. In other cases, such as the first two described below, the instability may simply result in a train of individual eddies that are still capable of interesting cross-shelf transports.

Inshore of the Gulf Stream (in the South Atlantic Bight), small (10-20 km) bottom-hugging eddies form along the coast of Florida and then strengthen as they propagate northward along the shelf edge (Brooks and Bane, 1983; Glenn and Ebbesmeyer, 1994; Luther and Bane, 1985). Near the surface, these eddies cause shallow transient tendrils of warm, salty Gulf Stream water to reach across much of the relatively fresh continental shelf. Further, Lee and Brink (2010) observe a very similar, but isolated, shelf-edge feature (associated with a proximate Warm Core Ring) on the southern side of Georges Bank, and they tentatively suggest that it may represent the same sort of processes as the South Atlantic Bight features. In any case, it remains to be seen as to how great an impact these transient features have on the ambient shelf waters, but there is evidence that these eddies can facilitate an important flux of nutrients onto the adjoining shelf (Lee et al., 1991). Having a lasting impact requires ultimately that the intrusion water mixes with shelf waters, a process that would occur efficiently via wind-driven mixing acting on the shallow surface tendrils. Planning an effective field program to deal with net exchanges due to these eddies would be rather difficult given their transient, mobile character.

At the shelf edge south of New England, i.e., well away from the western boundary current, there is also a persistent front that separates fresher, usually colder, shelf waters from ambient oceanic waters. This front has long been understood to be unstable (e.g., Flagg and Beardsley, 1978; Lozier and Reed, 2005), and there is plentiful evidence for small, O(10 km radius) eddies tied to this feature (e.g., Garvine et al., 1989). The net cross-shelf eddy fluxes associated with this process have been extremely hard to quantify, however, because of the low correlation of water properties with velocity and the highly variable spatial structure (e.g., Gawarkiewicz et al., 2004). This daunting difficulty in determining the consequences of the observed eddy field is apparently a fairly universal challenge in assessing oceanic eddy fluxes.

Inshore of the shelfbreak, i.e. over the shelf proper, a number of mechanisms can lead to eddy development, hence presumably cross-shelf fluxes. Previous subsections already discussed instabilities associated with alongshore wind forcing and with surface cooling over a bottom slope. In addition, spatial differences in tidally-driven mixing lead to tidal mixing fronts that separate shallower, vertically homogenized waters from deeper stratified waters (e.g., Loder et al., 1993). There is growing evidence (e.g., Simpson and James, 1986; Badin et al, 2009; Brink and Cherian, 2013) that these fronts are also unstable and lead to eddy generation. However, preliminary modeling results (Brink, 2013) suggest that, in at least some cases, the enhanced lateral mixing does not noticeably affect the biological activity, evidently because shallow and deep cross-frontal eddy nutrient fluxes roughly cancel. In addition, it seems likely that other sorts of fronts, such as those associated with buoyant outflows, are at least sometimes unstable as well (e.g., Ahlnäs et al. 1987; Mertz et al., 1988).

All considered, there is growing modeling evidence that instabilities and their consequent eddy fluxes may be ubiquitous over the continental shelf and slope. Demonstrating that these transient, coherent features actually exist (or estimating the exchange that they drive) in the ocean is a rather difficult matter, however. A horizontal eddy flux merely requires a correlation between velocity and some quantity like salinity, but it does not imply the existence of eddies in the sense of coherent, confined three-dimensional features. An eddy flux can be found, for example in time-dependent two-dimensional flow fields that are spatially very simple, and certainly not eddy-like. The point is that the existence of an eddy flux does not require the existence of eddies in the familiar sense. An eddy field is characterized, among other things, by finite spatial scales in the sense of an autocorrelation that decreases with observational separation. A proper, direct, observational demonstration of the importance of an eddy field over the shelf might well require synoptic, three-dimensional observations of small, mobile features that may be difficult to track via remote sensing. This is a daunting observational problem, and so it seems likely that a clever indirect approach, dealing with statistical properties rather than resolved maps, will be more profitable.

4.3 Topographic effects

Fluctuating cross-isobath flows, such as tides, are known to drive time-mean alongshore flows when bottom friction is present (e.g., Huthnance, 1973; Stern and Shen, 1976; Loder, 1980). Physically, bottom friction modifies the phase difference between the oscillating along-and cross-isobath flow components, so that a net cross-isobath eddy flux of along-isobath momentum results. (Note that, in this context, “eddy flux” does not imply the existence of closed eddies, but only that oscillating currents lead to a net transport). A cross-isobath divergence of this momentum flux occurs when the water becomes systematically shallower in one direction, and this flux divergence leads to a time-mean flow (directed alongshore with the coast on the right in the northern hemisphere). In the presence of density stratification, the rectified flow is bottom-intensified and strongest where the bottom slope is strongest (Maas and Zimmerman, 1989). It has further been shown that a similar mechanism applies in the absence of bottom friction if nonlinear effects are strong enough to generate the required phase offset (Robinson, 1981). This rectification process is relatively well-understood in the two-dimensional context where along-isobath variations are negligible.

Topographic rectification has interesting implications for the cross-shelf transport. The mean along-isobath flow leads to an Eulerian mean downslope flow in the bottom Ekman layer, and, if (in a stratified setting) the bottom-trapped rectified flow is geostrophically balanced, thermal wind requires isopycnals to slope upward toward the coast (Figure 4). This density distribution (isopycnals upwarped toward shallow water) appears to be inconsistent with the required downslope bottom Ekman transport. Observations and models for a seamount (Brink, 1995; Beckmann and Haidvogel, 1997) or a shelf (Chen and Beardsley, 1995) resolve this conundrum by showing that a balance is maintained through an up-gradient onshore eddy flux of dense water above the bottom boundary layer. Thus, a density distribution normally associated with upwelling could actually be associated with time-mean Eulerian downwelling!

The same sort of results apply in a three-dimensional case, where there are alongshore variations in the bottom slope (Merryfield and Holloway, 1999), although the strength of the rectified flow depends on the length scale of the topographic variations (Zimmermann, 1980; Brink, 2011). A series of intriguing theoretical and modeling studies (e.g., Salmon et al., 1976; Holloway 2009) have shown that the rectification and upwelling can be rationalized in terms of a chaotic eddy field approaching a state of maximum entropy, although Zimmermann (1980) showed that similar results are obtained in terms of the momentum balance described at the outset of this section. Regardless of the mechanistic explanation, it seems clear that fluctuating currents, tidal or not, over a shelf can lead to a mean alongshore flow and to a systematic and somewhat unintuitive pattern of cross-shelf material transports. It is apparently this transport mechanism that maintains the sustained high biological productivity on the tidally dominated Georges Bank, for example, by transporting nutrients into shallower water with the denser water (Franks and Chen, 1996). These considerations lead to the intriguing conjecture that, because fluctuating currents are ubiquitous, the coastal ocean is bound to be biologically productive simply because there is a sloping bottom to assure that eddy fluxes convey nutrients onshore.

The preceding paragraphs discuss bottom topography in a regional sense, focusing on net exchanges driven by fluctuating currents. There is, however, a fascinating array of effects that occur around isolated features such as ridges or canyons that cross the shelf. Flow disruptions can be either localized or widespread, and flow can involve rapid time variations or steady perturbations. Generally speaking, topographic features with short spatial scales are more likely to be associated with nonlinear dynamics (as is evident from the Rossby number $V / (f L_C)$, where V is a typical velocity magnitude, and L_C is an alongshore topographic length scale) and with bottom trapping (see the discussion in section 3.2). Further, the blockage of alongshore flow within a canyon can allow frictional ageostrophic terms to become important, and thus enable stronger cross-shelf exchange without enhanced nonlinearity (e.g., Lentz, et al., 2014). There is a very substantial literature on localized topographic features, such as canyons, which has been reviewed by Allen and Durrieu de Madron (2009), so no detailed attention is paid to these important issues here. However, it is worth pointing out that, to the extent that canyons represent a preferred location for violating the Taylor-Proudman theorem, hence expediting cross-shelf exchanges, we need to allow for the possibility that even a handful of these localized features can dominate shelf-ocean exchanges on a regional scale. Using observations to evaluate this possibility, however, represents a daunting task given the apparent need to resolve both short-scale and regional-scale flow features effectively.

5. Time dependence

The Taylor-Proudman theorem does not apply to flows that vary substantially in time (e.g., over an inertial period $2\pi/f$). However, in a simplistic sense, flows that simply slosh back and forth will not contribute to any net exchange. Yet time-dependent flows are important for generating cross-shelf exchanges, as the example of topographic rectification has already suggested.

Mixing processes are central to establishing a lasting cross-shelf transfer due to oscillating flows over the shelf. Sloshing a water parcel back and forth can lead to a lasting net transfer, if there is some vertical mixing. This is expressed quantitatively in terms of shear dispersion (e.g., Young et al., 1982): the way that vertical mixing and vertical shear interact to give an effective lateral mixing. Thus, for example, the weak, oscillating cross-shelf flows associated with coastal-trapped waves can lead to some horizontal exchange. This consideration puts aside the rather difficult concept of strictly horizontal mixing, which can usually be expressed in terms of eddies or fluctuating flows associated with some form of vertical mixing.

Perhaps the most obvious time-dependent effect is the passage of energy and momentum from the deep ocean onto the shelf. For example, winds over the open ocean drive the development of surface gravity waves, and these transfer momentum from offshore onto the shelf and into the surf zone. One consequence of this momentum flux is the generation of steady cross-shelf circulations as described in section 3.4. Further, the waves generate turbulence on the inner shelf that in turn affects how materials are transported vertically. At a lower frequency, ocean tides are astronomically generated primarily over the open ocean, and the ocean-basin tides in turn drive those over the continental shelf (e.g., Clarke, 1991). Often, tides over the shelf are amplified or even resonant, so that rectified currents, turbulence (hence vertical mixing) and tidal mixing fronts can result. As noted in section 4.3, these phenomena act to drive cross-shelf exchanges in their own right. The point is that time dependence allows energy to pass onto the shelf, and the resulting transformations lead to a lasting exchange mechanism.

There are, of course, other examples of time-dependent motions in the open ocean leading to shelf variability. Equatorial wind forcing, for example, can lead to alongshore-propagating shelf motions on time scales ranging from days (Cornejo-Rodriguez and Enfield, 1987) out to at least those of El Niño (e.g., Clarke, 2008). For the longer time scales, the resulting motions have clear biological implications, thus indicating that material transports are fairly probable. Some questions arise, however, as to how these transports occur. Wave-like motions on time scales of days and longer are constrained to have flow primarily along isobaths (e.g., Gill and Schumann, 1974), and there is evidence that El Niños are accompanied by strengthened poleward flows over the continental margin (e.g., Smith, 1983; Huyer et al., 2002). Thus, it is possible that the substantial biological impacts at mid-latitude are generated by alongshore advection and that cross-shelf transports occur primarily near the equator. Further, there is some evidence that wind-driven westward propagating baroclinic Rossby waves, which have time scales of years, affect coastal sea level on the western boundary of the North Atlantic Ocean (Hong et al., 2000). In all of these cases, it is clear that motions generated in the open ocean lead to a physical signal over the shelf itself. Further, models (e.g., Clarke and Van Gorder, 1994; Hong et al., 2000) do involve (explicitly or implicitly) some degree of cross-shelf flow. What remains elusive, especially observationally, is an assessment of whether these lower frequency oceanic motions lead to substantial cross-shelf exchanges.

6. Musings

6.1 Goals

There is much yet to be done. Many interesting and important aspects of cross-shelf exchanges are now areas of active work, or probably will be soon. One question is what determines the alongshore scale over which buoyancy currents due to river outflows dissipate (e.g., Lentz, 2004)? Another is how does topography affect the rate of cascading of dense shelf waters into the deeper ambient ocean (e.g., Spall, 2013)? These are fairly specific, process-centered possibilities where considerable progress can be made using numerical models, although the conclusive field measurements may be harder to achieve. There are aspects of cross-shelf exchange that are more poorly defined at this time, and that seem to call for a more exploratory approach. For example, there are biologically productive shelf regions, such as off southern Chile (Mackas et al., 2006), where the transport pathway that provides nutrients to sustain high productivity is as yet not as well characterized as one might hope.

Beyond these fairly specific issues, there are at least two grand challenges in synthesizing our knowledge. One important direction can be characterized as “prediction”. This goes beyond describing what the major cross-shelf pathways might be for a particular region, and demands that we be able to predict, with confidence, what the transports are under particular conditions.

Obviously, numerical models are extremely helpful tools for this sort of estimation, and their credibility will only increase in the coming years. However, we can never be entirely satisfied with these model outputs unless they are shown to be quantitatively consistent with observed transport estimates. An even more ambitious goal than prediction is to use readily available inputs to drive reliable simple parameterizations of cross-shelf fluxes that can be used for, say, global biogeochemical models. Meaningful simplicity is, in itself, a challenge.

A second grand challenge in cross-shelf exchange is to understand how the physical behavior of a coastal ocean determines (or fails to determine) the structure of the shelf ecosystem e.g., Steele and Ruzicka (2011). Such things as the quantity of nutrients delivered to a system, the residence times of water over the shelf, water temperature and the amount of vertical mixing are all likely to be influential in determining the amount of biological activity and the sorts of species that will be best adapted to the setting. All of these quantities are at least potentially strongly affected by cross-shelf transports, but the question remains as to how sensitive the ecosystem might be to transport compared to the myriad other chemical or purely biological interactions within this highly complex system. Further, one can ask how well one needs to characterize the physical system in order to obtain reasonable results from an ecosystem model: for example, do we need to know winds on a daily basis, or are seasonal values sufficient? While preliminary efforts are being made to address these questions (e.g., the Northern California Current model of Steele and Ruzicka, 2011), a real resolution appears to be off in the distant future.

6.2 Getting there

Our tools for studying the ocean become ever more impressive. Two decades ago, our ability to carry out quantitative interdisciplinary ocean measurements was severely limited by some essential incompatibilities. For example, physical oceanographers could measure currents using moorings or underway acoustic Doppler current sensors with good accuracy and almost arbitrarily fine space and time resolution. At the same time, a biologist studying nutrients could only use laborious discrete bottle samples or surface underway values. Or, a zooplankton specialist could use painstaking net tow information or gross (only certain size classes, and no species information) measurements based on acoustic signal returns. It was thus impossible to measure meaningful transports on the space and time scales dictated by physical reality. At present, however, the situation has improved radically, even if we are not yet at an ideal state. Some nutrients can now be measured with optically based sensors (Moore et al., 2009), and with space and time resolution at the level of the physical oceanographer's CTD (conductivity-Temperature-Depth probe). A zooplankton specialist can now use the highly sophisticated Video Plankton Recorder (VPR: Davis et al., 2005) to sample zooplankton, identified to the level of taxa, on similarly fine scales. These are but two examples. The introduction of biological and chemical sensors that operate on the same space and time scales as traditional physical sensors opens up entirely new vistas for quantitative oceanography.

Purely physical instrumentation has also improved greatly. One particularly noteworthy development is the arrival of coastal radar systems (e.g., Paduan and Washburn, 2013; Kirincich et al., 2012) that can measure surface currents with horizontal resolutions down to the order of 1 km, and accuracies of a few cm/sec. These systems allow routine resolution of features not previously observed and open the door to regional-scale data analyses that are resolved in both space and time (e.g., Kim et al., 2011).

To these marvelous new sensors, one can add new platforms for carrying out coastal ocean research. Much has been said recently, for example, about the upcoming U.S. Ocean Observatory (Cowles et al., 2010), and its near-permanent presence in the ocean. Further, dedicated, needs-based observations (the Ocean Observing Systems, e.g., Malone and Cole, 2000) can contribute useful, sustained time series in a range of locations. Beyond these entities, however, is a revolutionary change in observing platforms: new autonomous ways to access the ocean. In the coastal context, there are two particularly intriguing platform classes. First, are powered AUVs (Autonomous Underwater Vehicles: e.g., Moline et al., 2005) which can generally move fast enough (a few knots) to sample on a truly set program, but that have energy requirements that limit mission duration. Underwater docking capabilities are likely to make these vehicles highly desirable. A second platform class is the unpropelled AUV, or "gliders" (e.g., Rudnick et al., 2004). Several types are presently in use, and these systems are very attractive because their low power requirements allow them to stay at sea for extended periods (a month or more), while still profiling along a roughly defined path. Issues related to slow vehicle motion and the need to use low-powered sensors create some limitations, but the potential for a sustained, low-cost ocean presence is very attractive.

New tools, however, are unlikely to be the only requirement for progress. Even with perfect instruments, the often weak character of cross-shelf flows, short correlation scales, and the inefficiency of eddy transports (in the sense of low correlations of velocity and a quantity such as salinity) all mean that the "signal to noise ratio" in observations is low. This, in turn implies that a direct approach to measuring cross-shelf transports may prove painfully difficult. In any case, it now seems obvious that the traditional approach of running only a single line of moorings or stations, while highly informative, is unlikely to account for alongshore variability adequately (e.g., Lentz, 1987). The lesson here is that we need to think about new, and possibly indirect, approaches to estimating important transports. For example, Lentz (2008, 2010) pieced together historical current meter measurements in the middle Atlantic Bight to treat volume, heat and salt balances, and these included estimates of cross-shelf transports and their uncertainties. There are two aspects of this approach that are noteworthy. One is the use of integrated alongshore divergences to estimate cross-shelf transports as a residual (i.e., what passes into a box through the surface and through the cross-shelf boundaries has to be balanced by cross-shelf flow). The other is the creative use of a simple synthesis to unify the rather heterogeneous historical data sets. While this approach yields an incomplete answer (no information is gained about process or spatial structure), it is nonetheless an important advance.

To a young scientist seeking to learn about cross-shelf exchanges, there is much to be happy about. First, there is no single answer to this problem, and so there still remains open a wonderful range of interesting processes, and their combinations, to study in a variety of important contexts. Second, our ability to model oceanic processes is growing as computational power and conceptual understandings both improve dramatically. Third, there are now wonderful new observational tools and platforms that were barely imaginable a generation ago. Hard work, imagination and (yes) luck will lead to wonderful new results!

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Literature Cited:

Ahlnäs K, Royer TC, George TH 1987. Multiple dipole eddies in the Alaskan Coastal Current detected with Landsat thematic mapper data. *J. Geophys. Res.* 92**:**13041-13047

Allen JS 1980. Models of wind-driven currents on the continental shelf. *Ann. Rev. Fluid Mech.* 12:389-433

Allen JS, Bane J, Brink KH (chair), Brooks D, Brown W, Caldwell D, Collins C, Guza RT , Jumars P, Royer T (Steering Committee) 1988. Coastal Physical Oceanography: Towards a National Plan, Report of a meeting of the Coastal Physical Oceanography Community held January23-26, 1988 in Gulf Park, Mississippi, sponsored by the National Science Foundation, 118pp

Allen JS, Newberger PA 1996. Downwelling circulation on the Oregon continental shelf. Part I: Response to idealized forcing. *J. Phys. Oceanogr.* 26:2011-2035

Allen JS, PA Newberger PA, Federiuk J 1995. Upwelling circulation on the Oregon continental shelf. Part I: Response to idealized forcing. *J. Phys. Oceanog*r. 25:1843-1866

Allen SE, Durrieu de Madron X 2009. A review of the role of submarine canyons in deep-ocean exchange with the shelf. *Ocean Sci*. 5:607-620

Badan-Dangon A, Brink KH, Smith RL 1986. On the dynamical structure of the midshelf water column off northwest Africa. *Continental Shelf Res*. 5:629-644

Badin G, Williams RG, Holt JT, Fernand LJ 2009. Are mesoscale eddies in shelf seas formed by baroclinic instability of tidal fronts? *J. Geophys. Res*. 114, C10021, doi:10.1029/2009JC005340

Barth JA 1989a. Stability of a coastal upwelling front. Part 1: Model development and stability theorem. *J. Geophys. Res*. 94:10,844–10,856

Barth JA 1989b. Stability of a coastal upwelling front. Part 2: Model results and comparison with observations. *J. Geophys. Res*. 94:10,857–10,883

Barth JA 1994. Short-wavelength instabilities on coastal jets and fronts. *J. Geophys. Res.* 99:16,095–16,115

Barth JA, Pierce SD, Cowles TJ 2005. Mesoscale structure and its seasonal evolution in the northern California Current system. *Deep-Sea Research II* 52:5-28

Beckmann A, Haidvogel DB 1997. A numerical simulation of flow at Fieberling Guyot. *J. Geophys. Res*. 102:5595-5614

Brink KH 1995. Tidal and lower frequency currents above Fieberling Guyot. *J. Geophys. Res*. 100:10817-10832

Brink KH 1998. Deep-sea forcing and exchange processes. Chap. 6 In: *The Sea*, Vol. **10**, *The Global Coastal Ocean: Processes and Methods,* KH Brink and AR Robinson, editors, John Wiley & Sons, New York, pp. 151-167

Brink KH 2011. Topographic rectification in a stratified ocean. *J. Marine Res*. 68:337-368

Brink KH 2012. Buoyancy arrest and cross-isobath transport. *J Phys. Oceanogr*. 42:644-658

Brink KH 2013. Instability of a tidal mixing front in the presence of realistic tides and mixing. *J. Marine Res.* 71:227-252

Brink KH 2015. Continental shelf baroclinic instability 1: impulsive wind driving. Submitted to *J. Phys. Oceanogr.*

Brink KH, Cherian DA 2013. Instability of an idealized tidal mixing front: symmetric instabilities and frictional effects. *J. Marine Res.* 71:425-450

Brink KH, Lentz SJ 2010. Buoyancy arrest and bottom Ekman transport, Part I, Steady Flow. *Journal of Physical Oceanography* 40**:**621-635

Brink KH, Seo H 2015. Continental shelf baroclinic instability 2: oscillating wind forcing. Submitted to *J. Phys. Oceanogr.*

Brooks DA, Bane JM 1983. Gulf Stream meanders off North Carolina during winter and summer 1979. *J. Geophys. Res*. 88:4633-4650

Chapman DC 2000. A numerical study of the adjustment of a narrow stratified current over a sloping bottom. *J. Phys. Oceanogr*. 30:2927-2940

Chapman DC 2002. Deceleration of a finite-width, stratified current over a sloping bottom: frictional spindown or buoyancy shutdown?. *J. Phys. Oceanogr*. 32:336-352

Chapman DC, Gawarkiewicz G 1995. Offshore transport of dense shelf water in the presence of a submarine canyon. *J. Geophys. Res.* 100**:**13,373-13,387

Chapman DC, Gawarkiewicz G 1997. Shallow convection and buoyancy equilibration in an idealized coastal polynya. *J. Phys. Oceanogr*. 27:555-566

Chapman DC, Lentz SJ 1997. Adjustment of stratified flow over a sloping bottom. *J. Phys. Oceanogr*. 27:340-356

Chaudhuri AH, Bisagni JJ, Gangopadhyay A 2009. Shelf water entrainment by Gulf Stream warm-core rings between 75° W and 50° W during 1978-1999. *Cont. Shelf Res.* 29:393-406

Chen C, Beardsley RC 1995. A numerical study of stratified tidal rectification over finite-amplitude banks. Part I: symmetric banks. *J. Phys. Oceanogr.* 25:2090-2110

Chern C-S, Wang J, Wang D-P 1990. The exchange of Kuroshio and East China Sea shelf water. J. Geophys., Res., 95:16017-16023

Clarke AJ 1977. Observational and numerical evidence for wind-forced coastal-trapped waves. *J. Phys. Oceanogr.*, 7: 231-247

Clarke AJ 1991. The dynamics of barotropic tides over the continental shelf and slope (review), in *Tidal Hydrodynamics*, BB Parker, editor, J. Wiley & Sons, Inc., 79-108

Clarke AJ 2008. *An introduction to the dynamics of the El Niño and the Southern Oscillation*. Academic Press, Elsevier, London, 308pp

Clarke AJ, Van Gorder S 1994. On ENSO coastal currents and sea levels. *J. Phys. Oceanogr*. 24:661-680

Cornejo-Rodriguez M, Enfield DB 1987. Propagation and forcing of high-frequency sea-level variability along the west coast of South America. *J. Geophys. Res*. 92:14,323-14,334

Cowles T, Delaney J, Orcutt J, Weller R 2010. The Ocean Observatories Initiative: sustained ocean observing across a range of spatial scales. *Marine Technology Society Journal* 44:54-64

Csanady GT 1978. The arrested topographic wave. *J. Phys. Oceanogr*. 8:47-62

Davis CS, Thwaites FT, Gallager SM, Hu Q 2005. A three-axis fast-tow digital plankton recorder for rapid surveys of plankton taxa and hydrography. *Limnol. Oceanogr.: Methods* 3:59-74

Dever EP 1997. Subtidal velocity correlation scales on the northern California shelf. *J. Geophys. Res*. 102:8555-8572

Durski SM, Allen JS 2005. Finite-amplitude evolution of instabilities associated with the coastal upwelling front. *J. Phys. Oceanogr*. 35:1606-1628

Ekman VW 1905. On the influence of the Earth's rotation on ocean currents. *Arch. Math. Astron. Phys.* 2:1-52

Fewings M, Lentz SJ, Fredericks J 2008. Observations of cross-shelf flow driven by cross-shelf winds on the inner continental shelf. *J. Phys. Oceanogr*. 38:2358-2378

Flagg CN, Beardsley RC 1978. On the stability of the shelf water/slope water front south of New England. *J. Geophys. Res*. 83:4623–4631

Ford WL, Longard JR, Banks RE 1952. On the nature, occurrence and origin of cold, low-salinity water along the edge of the Gulf Stream. *J. Marine Res*. 11:281-293

Franks PJS, Chen C 1996. Plankton production in tidal fronts: a model of Georges Bank in summer, *J. Marine Res.* 54:631-651

Garrett C, MacCready P, Rhines PB 1993. Boundary mixing and arrested Ekman layers: rotating stratified flow near a sloping bottom. *Annual Reviews of Fluid Mechanics* 25:291-324

Garvine RC, Wong K-C, Gawarkiewicz G 1989. Quantitative properties of shelfbreak eddies. *J. Geophys. Res*. 94:14,475-14,483

Gawarkiewicz G, Bahr F, Brink KH, Beardsley RC, Caruso M, Lynch JF, Chiu C-S 2004. A large-amplitude meander of the shelfbreak front during summer south of New England: Observations from the shelfbreak PRIMER experiment. *Journal of Geophysical Research-Oceans* **109**, C03006, doi: 10.1029/2002JC001468

Gill AE, Clarke AJ 1974. Wind-induced upwelling, coastal currents and sea-level changes. *Deep-Sea Res.* 21:325-345

Gill AE, Schumann EH 1974. The generation of long shelf waves by the wind. *J. Phys. Oceanogr*. 4:83-90

Glenn SM, Ebbesmeyer CC 1994. The structure and propagation of a Gulf Stream frontal eddy along the North Carolina shelf break. *J. Geophys. Res.* 99:5029-5046

Guerrero RA, Piola AR, Fenco H, Matano RP, Combes V, Chao Y, James C, Palma EB, Saraceno M, Strub PT 2014. The salinity signature of the cross-shelf exchanges in the southwestern Atlantic Ocean: satellite Observations. *J. Geophys. Res*. 119, doi:10.1002/2014JC010113

Hasselmann K 1970. Wave-driven inertial oscillations. *Geophys. Fluid Dyn.* 1:463-502

Holloway G 2009. Entropic forces in geophysical fluid dynamics. *Entropy* 11:360-383

Hong BG, Sturgis W, Clarke AJ 2000. Sea level on the U.S. east coast: decadal variability caused by open ocean wind-curl forcing. *J. Phys. Oceanogr.* 30:2088-2098

Hsueh Y, Lie H-J, Ichikawa H 1996. On the branching of the Kuroshio west of Kyushu. *J. Geophys. Res.* 101:3851-3858

Hsueh Y, O'Brien JJ 1971. Steady coastal upwelling induced by and along-shore current. *J. Phys. Oceanogr*. 1:180-186

Huthnance JM 1973. Tidal current asymmetries over the Norfolk Sandbanks. *Estuar. Coastal Mar. Sci.* 1:89-99

Huthnance JM 1995. Circulation, exchange and water masses at the ocean margin: the role of physical processes at the shelf edge. *Prog. Oceanogr*. 35:353-431

Huyer A, Smith RL, Fleischbein J 2002. The coastal ocean off Oregon and Northern California during the 1997-1998 El Niño. *Prog. Oceanogr*. 54:311-341

Joyce TM, Bishop KB, Brown OB 1992. Observations of offshore shelf water transport induced by a warm-core ring. *Deep-Sea Res.* 39(1):S97-S113

Kim SY, Terrill EJ, Cornuelle BD, Jones B, Washburn L, Moline MA, Paduan JD, Garfield N, Largier JL, Crawford G, Kosro PM 2011. Mapping the U.S. west coastal surface circulation: a multiyear analysis of high-frequency radar observations. *J. Geophys. Res*., **116**, C03011,doi:10.1029/2010JC006669

Kirincich AR, dePaolo R, Terrill E 2012. Improving HF radar estimates of surface currents using signal quality metrics, with application to the MVCO high-resolution radar system. *J. Atmos. and Oceanic Tech*., 29, 1377-1390

Kundu PK, Allen JS 1976. Some three-dimensional characteristics of low-frequency current fluctuations near the Oregon coast. *J. Phys. Oceanogr*. 6:181-199

Lee TN, Yoder JA, Atkinson LP 1991. Gulf Stream frontal eddy influence on productivity of the southeast U.S. continental shelf. *J. Geophys. Res*. 96:22191-22205

Lee C, Brink KH 2010. Shelf edge variability over Georges Bank: Winter and summer 1997. *J. Geophys. Res*. **115**, C08008, doi.10.1029/2009JC005706

Lentz SJ 1987. A heat budget for the northern California shelf during CODE-2. *J. Geophys. Res.* 92:14,491-14,509

Lentz SJ 2004. The response of buoyant coastal plumes to upwelling-favorable winds. *J. Phys. Oceanogr*. 34:2458-2469

Lentz SJ 2008. Observations and a model of the mean circulation over the Middle Atlantic Bight continental shelf. *J.Phys. Oceanogr.* 38:1203-1221, DOI 10.1175/2007JPO3768.1

Lentz SJ 2010. The mean along-isobath heat and salt balances over the Middle Atlantic Bight continental shelf. *J. Phys. Oceanogr.* 40:934-948

Lentz SJ, Butman B, Harris C 2014. The vertical structure of the circulation and dynamics in Hudson Shelf Valley. *J. Geophys. Res*. 119:3694-3713, doi:10.1002/2014JC009883

Lentz SJ, Chapman DC 2004. The importance of nonlinear cross-shelf momentum flux during wind-driven coastal upwelling. *J. Phys. Oceanogr*. 34:2444-2457

Lentz SJ, Fewings M 2012. The wind- and wave-driven inner-shelf circulation. *Ann. Rev. Marine Sci.* 4:317-344

Lentz SJ, Fewings M, Howd P, Fredericks J, Hathaway K 2008. Observations and a model of undertow over the inner continental shelf. *J. Phys. Oceanogr.* 38:2341-2357

Lentz SJ, Guza RT, Elgar S, Fedderson F, Herbers THC 1999. Momentum balances on the North Carolina inner shelf. *J. Geophys. Res*. 104:18,205-18,226

Lentz SJ, Trowbridge JH 1991. The bottom boundary layer over the northern California shelf. *J. Phys. Oceanogr*. 21:1186-1201

Loder JW 1980. Topographic rectification on the sides of Georges Bank. *J. Phys. Oceanogr*. 10:1399-1416

Loder JW, Drinkwater KF, Oakey NS, Horne EPW 1993. Circulation, hydrographic structure and mixing at tidal fronts: the view from Georges Bank. *Phil. Trans.: Physical Sciences and Engineering* 343:447-460

Lozier MS, Reed MSC 2005. The influence of topography on the stability of shelfbreak fronts. *J. Phys. Oceanogr*. 35:1023-1036

Luther ME, Bane, Jr. JM 1985. Mixed instabilities in the Gulf Stream over the continental slope. *J. Phys. Oceanogr.* 15:3-23

Maas LRM, Zimmerman JTF 1989. Tide-topography interactions in a stratified shelf sea II. Bottom trapped internal tides and baroclinic residual currents. *Geophys. Astrophys. Fluid Dyn. 45:*37–69

Mackas DL, Strub PT, Thomas A, Montecino V 2006. Eastern ocean boundaries pan-regional overview (E). In *The Sea*, vol. 14. (AR Robinson and KH Brink, editors), Harvard University Press, Cambridge MA, 21-5

Malone TC, Cole M 2000. Toward a global scale coastal ocean observing system. *Oceanography* 18:7-11

Mertz G, El-Sabh MI, Proulx D, Condal AR 1988. Instability of a buoyancy- driven coastal jet; The Gaspé Current and its St. Lawrence precursor. *J.Geophys. Res.* 93:6885-6893

Merryfield WJ, Holloway G 1999. Eddy fluxes and topography in stratified quasi-geostrophic models*. J. Fluid Mech.* 380:59-80

Middleton JF, Ramsden D 1996. The evolution of the bottom boundary layer on the sloping continental shelf: a numerical study. *J. Geophys. Res*. 101:18,061-18,077

Mitchum GT, Clarke AJ 1986. The frictional nearshore response to forcing by synoptic scale winds. *J. Phys. Oceanogr*. 16:934-946

Moline MA, Blackwell SM, von Alt C, Allen B, Austin T, Case J, Forrester N, Goldsborough R, Purcell M, Stokey R 2005. Remote environmental monitoring units: an autonomous vehicle for characterizing coastal environments. *J. Atmos. Oceanic Tech*. 22:1797-1808

Moore TS, Mullagh KM, Holyoke RR, Madison AS, Yücel M, Luther GW III 2009. Marine chemical technology and sensors for marine waters: potentials and limits. *Annu. Rev. Mar. Sci*. 1:91-115

Paduan JD, Washburn L 2013. High-frequency Radar observations of ocean surface currents. *Annu. Rev. Mar. Sci.*, 5, 115-136

Pedlosky J 1974. Longshore currents, upwelling and bottom topography. *J. Phys. Oceanogr*. 4:214-226

Pedlosky J 1979. *Geophysical Fluid Dynamics*. Springer-Verlag, New York, 624 pp

Pringle JM 2001. Cross-shelf eddy heat transport in a wind-free coastal ocean undergoing winter time cooling. *J. Geophys. Res.* 106:2589-2604

Pringle JM, Dever EP 2009. Dynamics of wind-driven upwelling and relaxation between Monterey Bay and Point Arena: local-, regional-, and gyre-scale controls. *J. Geophys. Res*. 114 C07003, DOI 10.1029/2008JC005016

Rivas D, Samelson RM 2011. A numerical modeling study of the upwelling source waters along the Oregon coast during 2005. *J. Phys. Oceanogr*. 41:88-112

Robinson IS 1981. Tidal vorticity and residual circulation. *Deep-Sea Res. A* 28:195-212

Roughan M, Garfield N, Largier J, Dever EP, Dorman C, Peterson D, Dorman J 2006. Transport and retention in an upwelling region: the role of across-shelf structure. *Deep-Sea Res., Part II* 53:2931-2955

Rudnick DL, Davis RE, Eriksen CC, Fratantoni DM, Perry MJ 2004. Underwater gliders for ocean research. *Marine Technology Society Journal* 38:73-84

St. Maurice JP, Veronis G 1975. A multi-scaling analysis of the spin-up problem. *J. Fluid Mech*. 68:417-445

Salmon R, Holloway G, Hendershott MC 1976. The equilibrium statistical mechanics of simple quasi-geostrophic models. *J. Fluid Mech.* 75:691–703

Savidge DK, Savidge WB 2014. Seasonal export of South Atlantic and Mid-Atlantic Bight shelf waters at Cape Hatteras. *Continental Shelf Res*. 74:50-59

Simpson JH, James ID 1986. Coastal and estuarine fronts, *in Baroclinic Processes on Continental Shelves,* CNK Mooers, ed., American Geophysical Union, Washington, DC, 63–94

Smith RL 1981. A comparison of the structure and variability of the flow field in three coastal upwelling regions: Oregon, northwest Africa and Peru. In: *Coastal Upwelling*, F.A. Richards, ed., American Geophysical Union, Washington, D.C., 107-118

Smith RL 1983. Peru coastal currents during El Niño: 1976 and 1982. *Science* 221:1397-1399

Spall MA 2013. Dense water formation around islands. *J. Geophys. Res,* 118:2507-2519, doi:10.1002/jgrc20185

Steele JH, Ruzicka JJ 2011. Constructing end-to-end models using ECOPATH data. *J. Marine Syst.* 87:227-238

Stern ME, Shen CY 1976. Displacement and rectification of planetary fluids. *Geophys. Fluid Dyn.* 7:81–118

Tilburg CE 2003. Across-shelf transport on a continental shelf: Do across-shelf winds matter? *J. Phys. Oceanogr*. 16:1165-1178

Thorade H 1909. Ueber die Kalifornische meeresstromungen, oberflachentemperaturen und stromungen an der Westkuste Nord- amerikas. *Ann. Hydr. u. marit. Meteor.* 37:17-34. 63-77

Trowbridge JH, Lentz SJ 1991. Asymmetric behavior of an oceanic boundary layer above a sloping bottom. *J. Phys. Oceanogr.* 21:1171-1185

Trowbridge JH, Lentz SJ 1998. Dynamics of the bottom boundary layer on the northern California shelf. *J. Phys. Oceanogr.* 28:2075-2093

Vélez-Belchí P, Centurioni LR, Lee D-K, Jan S, Niiler PP 2013. Eddy induced Kuroshio intrusions onto the continental shelf of the East China Sea. *J. Marine Res.*, 71: 83-108

Winant CD 1983. Longshore coherence of currents on the southern California shelf during the summer. *J. Phys. Oceanogr.* 13:54-64

Young WR, Rhines PB, Garrett CJR 1982. Shear-flow dispersion, internal waves and horizontal mixing in the ocean. *J. Phys. Oceanogr.* 12:515-527

Zimmermann JTF 1980. Vorticity transfer by tidal currents over irregular topography. *J. Mar. Res.* 38:601–630

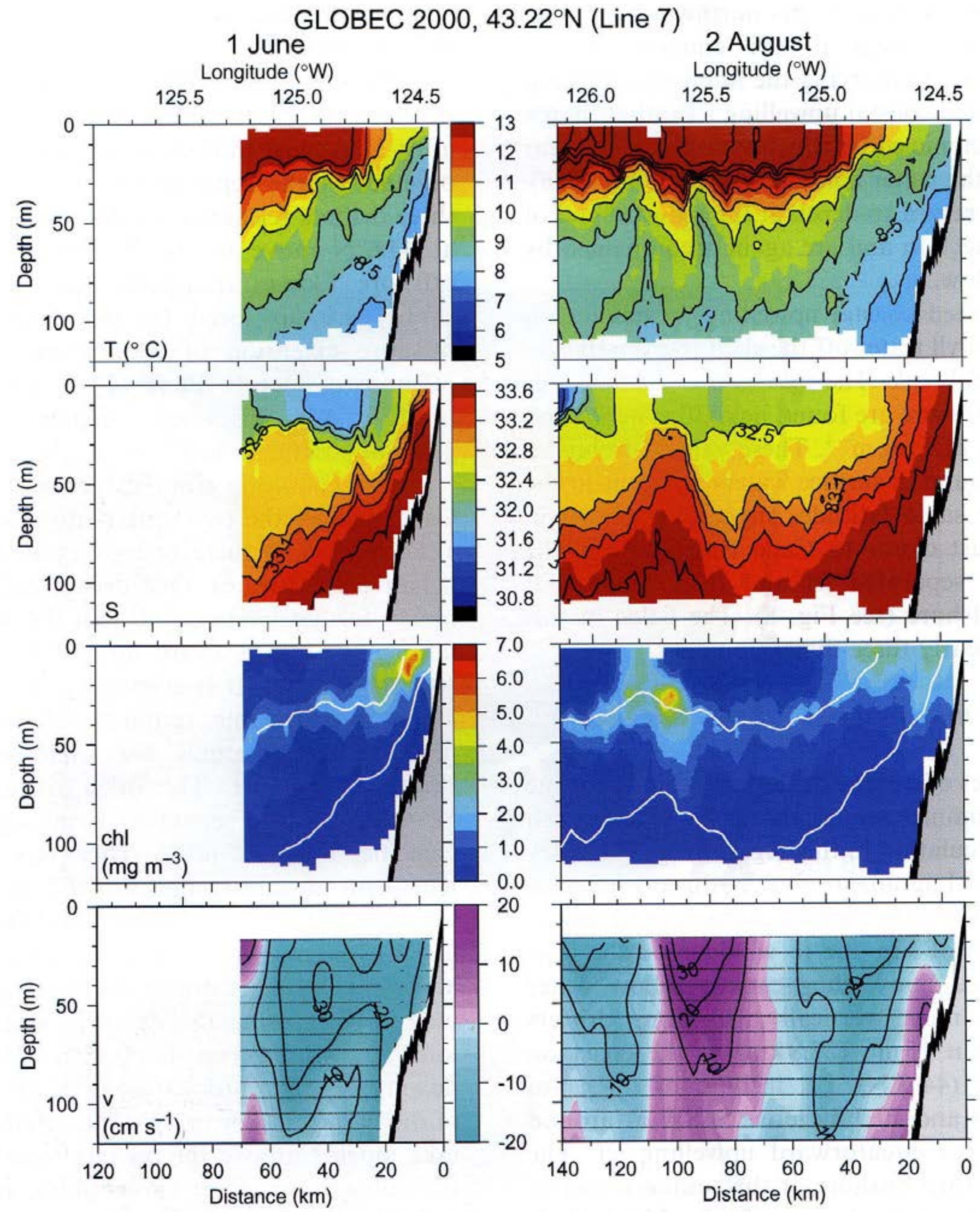

Figure 1: [Thin this down to left side 1st and 3rd panels only] Cross-shelf sections of temperature (top) and chlorophyll *a* concentration (lower) off central Oregon during active coastal upwelling. from Barth et al. (2005).

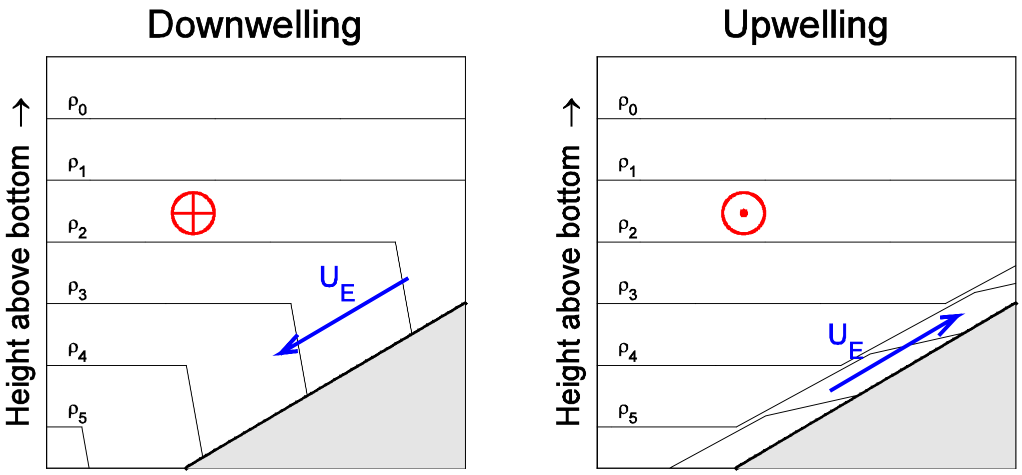

Figure 2: Schematic showing the density structure (black contours) and bottom boundary layer currents (blue vectors) for the cases of alongshore flow in the direction of coastal-trapped wave propagation (downwelling: left panel), and in the opposing alongshore direction (upwelling: right panel). U_E is the bottom Ekman transport.

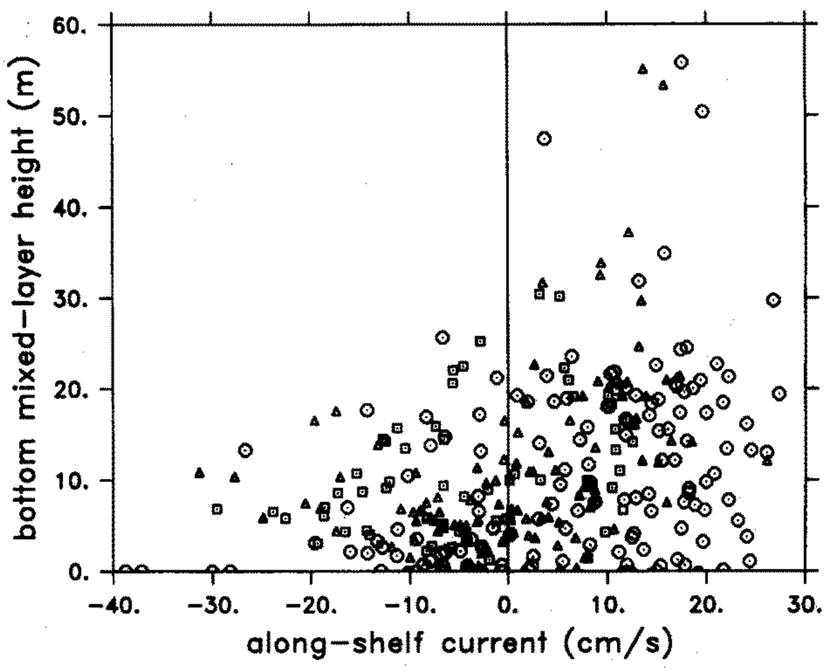

Fig. 3: Scatterplot of observed bottom boundary layer thickness vs. coincident alongshore velocity for the central continental shelf off California. The considerable scatter is associated with variables besides alongshore velocity, such as water column density stratification, that are important for determining bottom boundary layer thickness. From Lentz and Trowbridge (1991).

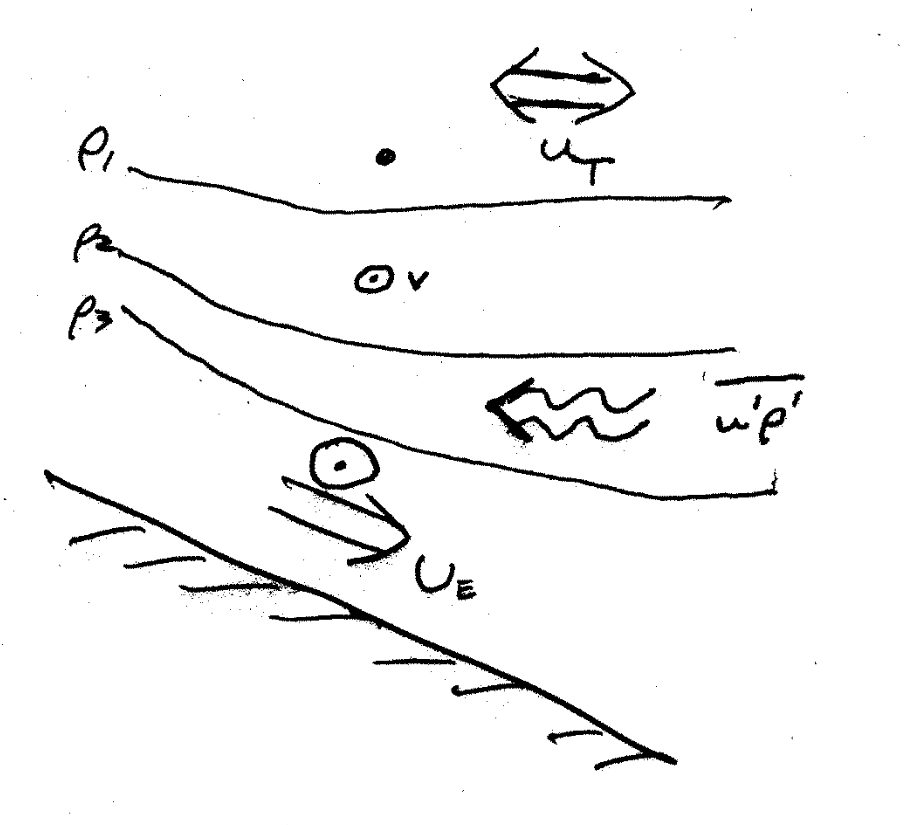

Figure 4: Schematic of density structure and flow features resulting from topographic rectification of an oscillating cross shelf flow (u_T) in an alongshore-uniform system. Even though the Ekman transport U_E is downslope, the isopycnal slope is made possible by a horizontal eddy density flux $\overline{u'\rho'}$.