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Fronts on the Continental Shelf

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A well-defined class of fronts occurring in the shelf seas around the United Kingdom during the summer months marks the boundary between stratified and vertically mixed regimes. The occurrence of these fronts may be interpreted in terms of the distribution of available turbulent kinetic energy from the tidal currents and wind stress and the buoyancy flux input at the surface. The principal parameter controlling stratification is the ratio of the water depth to a Monin-Obukhov length determined by the tidal velocity. A corresponding parameter based on the wind stress is also found to contribute significantly in a combined model of tide and wind mixing. Vertical sections perpendicular to the front by undulating CTD indicate a strongly baroclinic region with horizontal temperature gradients of ~1°C/km, which imply strong flows parallel to the front. Drogue observations show that non-tidal velocities of ~10 cm/s occur in the vicinity of the fronts but the flow regime is apparently complicated by large-scale (~25 km) instabilities which are clearly manifest in satellite infrared imagery and airborne radiation thermometer surveys. There is also indirect evidence for vertical motions suggesting that both upwelling and downwelling occur in the frontal zone.

INTRODUCTION

Recent studies have drawn attention to marked frontal structures occurring on the European continental shelf during the summer months. These fronts are produced by variations in the level of tidal mixing, and it has been shown [Simpson and Hunter, 1974; Simpson et al., 1977] that their location is essentially determined by the parameter $h/a^2$, where $h$ is the water depth and $a$ the amplitude of the tidal stream. If $a$ is taken for mean spring tides, the critical $h/a^2$ is in the range 50–100 m$^{-1}$ for the shelf seas around the United Kingdom.

These fronts therefore form a series of almost fixed geographical boundaries in the summer regime of the tidally energetic seas of the European shelf. At the frontal boundary there is a transition between a vertically mixed system on one side and stratified water on the other. An example of the temperature, salinity, and density fields in a front of this kind, taken from a recent survey by Batfish CTD, is shown in Figure 1. The density here is predominantly controlled by temperature, and this is usually, though not always, the case. The high resolution afforded by the Batfish system reveals a complex detailed structure within the front, but the basic transition between stratified and mixed systems is clearly evident. This basic pattern is also found to be imposed on the distribution of chemical nutrients with important implications for the distribution of biomass [Saunders, 1976; Pingree et al., 1977].

The surface manifestation of these fronts is often sufficiently strong to permit their detection by radiation thermometry. On the small number of cloud-free days in the British summer the fronts may be clearly identified in satellite infrared images. Figure 2 is a selected example showing several of the major frontal systems. The positions of the fronts shown here are all close to the critical contour of $h/a^2$ and in accord with the results of direct observations by ships. Infrared images of this kind, together with measurements of sea surface temperature by airborne radiation thermometry, have extended considerably our ability to study the detailed evolution of these fronts.

In this paper, following a summary of the essential ideas of the mixing model, we make use of recent NOAA 5 images, together with data taken by radio-tracked drogues and conventional ship techniques, to extend our knowledge of the field of motion associated with the fronts.

VERTICAL MIXING MODEL

A previous model of stratification and mixing in shallow seas [Simpson and Hunter, 1974] may be extended to include the effect of wind mixing. As an index of stratification we use the potential energy relative to the mixed state defined by

$$V = \frac{(\rho - \rho_0)g}{h} \int_z^d \rho \, dz = \frac{1}{h} \int_z^d \rho \, dz$$

where $\rho$ is density and $h$ the depth of the water column. For a vertically mixed system, $V = 0$, and for increasingly stable stratification, $V$ becomes negative.

When $V < 0$, tidal stirring and mixing due to wind stress bring about positive changes in $V$.

The available power for these changes $P_A$ is assumed to be a fixed fraction of the rates of working by the two boundary stresses and may be written as

$$P_A = \varepsilon k_p \rho(u_s^2) + \delta k_p (W_s^2)$$

where $\rho$ and $\rho_0$ are water and air densities, $u_s$ is the near-bottom velocity, $W_s$ is the wind speed near the sea surface, $k_p$ and $k_s$ are drag coefficients, and $\varepsilon$ and $\delta$ are the efficiencies of tide and wind mixing.

Surface heating, on the other hand, brings about increasingly negative $V$ at a rate given by [Simpson and Hunter, 1974]

$$(\frac{dV}{dt})_{\text{heat}} = -\alpha \frac{\bar{Q}h}{2c}$$

where $\bar{Q}$ is the rate of heat input, $c$ is the specific heat, and $\alpha$ is the volume expansion coefficient.

On the assumption that only these local processes are important, the overall potential energy balance may then be written

$$\frac{dV}{dt} = \frac{-\alpha \bar{Q}h}{2c} + \varepsilon k_p \rho(u_s^2) + \delta k_p (W_s^2)$$

Dividing by the heating term we have the nondimensional form

$$R = \frac{-dV}{dt} \frac{\bar{Q}h}{2c} = 1 - \frac{2\varepsilon k_p \rho(u_s^2)}{\alpha \bar{Q}h} - \frac{2\delta k_p (W_s^2)}{\alpha \bar{Q}h}$$

$$= 1 - a \left( \frac{u_s^2}{\bar{Q}h} \right) - b \left( \frac{W_s^2}{\bar{Q}h} \right)$$

(1)
where $a$ and $b$ may be taken as positive constants. $R$ represents the fraction of the (negative) potential energy input by heating which is retained in the water column. When $R > 0$, stratification will increase with time, while $R = 0$ is the critical condition which specifies the shelf sea fronts. Note that this model does not describe potential energy changes for $R < 0$, since positive $V$ corresponds to unstable stratification.

The two nondimensional terms on the right of (1) have the form of the Monin-Obukhov lengths divided by the geometrical depth. To test the relevance of these parameters in determining $R$, we need time series data on the density $\rho(t)$ at a number of points on the shelf. Presently available data do not permit satisfactory estimates over short time scales, but it is possible to estimate an average $R$ over the period March–June from the available $V$ data.

In March, $V \approx 0$ throughout the area, so that the mean $R$ for the period March–June may be obtained from June data for $V$ according to

$$R = -\frac{dV}{dt} \left/ \frac{\alpha gQh}{2c} \right. = -V \left/ \frac{\alpha gQh}{2c} \right. = \frac{(V)}{(V_m)}$$

where $(V_m) = -\alpha gQ/2c$ is the limiting potential energy density

which would be the case of no $R$ time, $Q$, or $a$.

Values of $E$ are shown in the Celtic Sea in Figure 3. $E$, taken for $\mu$ (which is usual), and for $\mu$.

$E = E_0$

For a dat $Q$ will not function of...
which would occur for the March–June heat input $Q$ in the case of no mixing. For a limited geographical area and a given time, $Q$, and hence $(V_w)$, may be regarded as constants.

Values of $R$ obtained in this way and based on 146 stations in the Celtic and Irish seas in June 1973 are plotted versus $u^2/h$ in Figure 3. Here the surface tidal stream amplitude at springs $u_s$ taken from tidal atlas data, has been used in place of $u_0$ (which is unknown in most areas) on the assumption that $u_s \approx u$. The parameter $E$ is a nondimensional form of $u^2/h$ defined by $E = c_{pT}^2/\rho g Q h$.

For a data set from a single year for a limited area, $(W^2)$ and $Q$ will not vary greatly, so (1) would indicate that $(V)$ is a function of $u^2/h$ (tidal mixing) and $1/h$ (wind mixing). A regression analysis was therefore performed on the data in the range $u^2/h < 0.01$, where $R > 0$, yielding

$$|\langle V \rangle| = 138 - 8438 \frac{u^2}{h} - 3889(1/h)$$

with a multiple correlation coefficient of 0.79 and both $u^2/h$ ($t = 10.4$) and $1/h$ ($t = 5.4$) making significant contributions at the 0.1% level. Excluding the $1/h$ term gives a less satisfactory fit with a correlation coefficient of 0.73. The $|\langle V \rangle|$ intercept is 138 J/m$^2$, which may be compared to the value of $|\langle V_w \rangle| = 160$ J/m$^2$ estimated from the net heat input before the end of June.

Setting $u^2/h = 0$ we find that for $h \leq 28$ m there is complete vertical mixing ($\langle V \rangle = 0$). This, of course, does not preclude the occurrence of transient stratification in such areas when the
surface heating rate is high or the mixing weak [Simpson et al., 1977].

The relative importance of the tide and wind contributions for the period studied (March–June 1973) may be assessed from the ratio of the regression coefficients. This indicates that the tide dominates when \( u > 0.75 \) m/s, which is satisfied over most of the Irish Sea but excludes large areas of the Celtic Sea.

Much of the scatter in Figure 3 probably arises from our poor knowledge of tidal stream amplitudes, which may be in error by \(-30\% \) or more in places with consequences indicated by the error bars in Figure 3. This uncertainty should be substantially reduced by \( u^2/h \) estimated from numerical models of the shelf seas.

Table 1. Residual Currents From Drogue Tracks

<table>
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<tr>
<th>Drogue</th>
<th>Date (Launch)</th>
<th>Duration, hours</th>
<th>Depth, m</th>
<th>Stratification</th>
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*Two identical buoys.
Fig. 5. Sequence of infrared images from NOAA 5 on August 18–20, 1976, showing the evolution of a large-scale (~25 km) frontal instability. Note also the mixed plume produced by the Scilly Isles advecting northward in the mean current.
is small in relation to the June values for the stratified areas \((V_2 > 100 \, \text{J/m}^2)\), the importance of such intermittent marginal stratification in biological processes could be significant [Pin jersey et al., 1977].

On the other hand, the constant efficiency model may not adequately describe the situation near the transition point, where \(\varepsilon\) is probably dependent on existing stratification (positive feedback), so that the front position would be effectively controlled by the lowest level of tidal mixing experienced. If this were the case, only limited adjustment would be possible.

MODELS OF THE VELOCITY FIELD

In view of the known persistence of these features it is tempting to try to interpret them as approximately two-dimensional density and velocity patterns oscillating back and forth at the tidal frequency. This tidal advection, typically over an excursion of \(\approx 10-20\) km, may be ignored to a first approximation by taking coordinates in the moving front. In this reference frame the density gradients imposed by variations in mixing set up pressure forces which will cause accelerations resulting eventually in a velocity field in which pressure gradients and frictional forces are in balance with the geostrophic accelerations.

In the crudest model we may simply interpret the density field in terms of geostrophically balanced velocities. This approach [Simpson, 1976] implies the existence of a high-velocity jet (up to 30 cm/s) in the most strongly baroclinic regions of the fronts. The neglect of friction, which must play some part, means that the longitudinal velocities will be exaggerated. A two-dimensional model, which allows for both horizontal and vertical frictional stresses, has recently been developed by James [1977]. The coordinate frame is again chosen in the front, and the tidal velocity distribution enters only insofar as it controls the effective eddy diffusion coefficients. The resulting flow parallel to the front is similar to that of the geostrophic case, but the frictional damping of the baroclinic jet produces an unbalanced component of the pressure field which drives a circulation in the vertical plane transverse to the front with strong upwelling in the mixed water close to the front. A balancing downwelling motion occurs on the other side of the baroclinic zone with an associated convergence at the surface.

Observations

Radio Drogues

A number of direct observations have been made of the residual current field in the vicinity of these fronts. Because of the large tidal excursions involved, drogue buoys offer the best prospect of a satisfactory measurement. We have used a radio tracked system, which in the latest version uses a fixed geometry (cruising) drogue mounted below a very low windage surface element. The buoys are fixed at intervals (ideally a multiple of the semidiurnal tidal period) by the ship homing on their radio beacons, and velocities are deduced after correction for wind drag and tidal motion.
When only a single ship is used, it is difficult to make satisfactory observations of the density and velocity fields simultaneously. There is a particular difficulty in the accurate deployment of the drogues in the moving density field. The results of a number of such drogue tracking experiments made during the last few years are shown in Table 1. Generally, they were not amenable to interpretation in terms of approximately geostrophic flows parallel to the front. In several cases the observed residuals were nearly normal to the estimated direction of the front. The speeds vary over a wide range (2–27 cm/s) with no discernible tendency for large values to occur in the stratified, transitional, or unstratified regimes.

The implication seems to be that the flow is not organized into a quasi-two-dimensional pattern. This is perhaps not surprising, as theoretical models of this type of flow suggest that it may be unstable [e.g., Harri, 1974]. Further evidence for the unstable character of the flow comes from remote sensing of the sea surface, to which we now turn.

**Surface Isotherms**

Even from early surveys by ship and aircraft it was evident that the fronts were frequently convoluted. Figure 4 is an example of an aircraft survey using a radiation thermometer. Flight lines were at intervals of ~2 km, and the temperatures have been smoothed over a 4-km square. The temperature scale has been adjusted to agree with simultaneous ship observations in the area, and a correction for tidal displacement has also been applied. The data presented, which are a composite of two surveys on successive days, indicate a tight meander of the front with a radius of curvature of <10 km in places. Two drogues deployed near the apex of the meander showed motion parallel to the isotherms at 70 m (2.4 cm/s) but perpendicular to them at 15 m (3.2 cm/s).

That such meanders represent just one stage in an evolving instability process is strongly supported by a fortunate sequence of satellite infrared images acquired during the sustained period of cloud-free weather which the United Kingdom enjoyed in August 1976. Figure 5 shows the Celtic Sea front on 3 successive days at intervals requiring minimal corrections for tidal motions (maximum error of ~0.8 km). A marked meandering of the front with a wavelength of ~25 km is already evident in the west on the 18th. This meandering further increases on the 19th, and by the time of the image on the 20th the 'roll-up' is almost complete, isolating two cells of cold mixed water in the stratified region.

The velocities normal to the front may be deduced from comparison of successive images, as in Figure 6, which indicates maximum velocities of ~12 cm/s normal to the front. Such values are not inconsistent with the estimates from the drogue measurements.

Another interesting feature apparent in these pictures is the 'comet-tail' shaped patch of cold water surrounding the Scilly Isles. This is the result of local breakdown of stratification caused by the stirring rod effect of the islands which, relative to the water, are oscillating at tidal frequency. The asymmetry of the mixed regime is not due to the topography and so presumably reflects northward advection by the mean current. By measuring the mean velocity and the structure of this turbulent plume, it should be possible to estimate the time scales involved in the establishment of stratification. This feature may also be useful in studying the time scales associated with biological processes at shelf sea fronts.

**Vertical Motions**

The complexities of the horizontal velocity field described above suggest that the identification of a two-dimensional circulation in the plane perpendicular to the front is unlikely. Moreover, the small magnitude of the velocities (maximum of <1 mm/s) would seem to preclude their direct measurement.

There are, however, two observed features of these fronts which are clearly indicative of vertical motions. First, in near calm surface conditions, a region of surface convergence is frequently observed in the fronts. From the air a more or less continuous slick may be seen along the length of the front, while observers at the sea surface report significant accumulations of seaweed, jellyfish, etc. Such convergences are formed at or close to the region of maximum horizontal temperature gradient. A second indication of circulation in the vertical plane is the frequent (though not invariable) observation of a minimum in sea surface temperature occurring in the mixed water just before crossing the front. Examples of this phenomenon are shown in Figure 7a, where this minimum temperature is seen to be ~0.1°C below that of the rest of the mixed water. An associated upwelling of isotherms close to the maximum surface gradient in the front is also apparent at times in vertical section data as, for example, in Figure 1a.

These indications of upwelling combined with the convergence suggest the circulations shown in Figure 7b, which is qualitatively consistent with the vertical plane flow predicted in the James model. This picture should, however, be treated with caution, in view of the obvious complexity of the horizontal flows. It has been suggested [Woods et al., 1977] that such meandering flows will themselves force strong upwelling and downwelling along the front, and these may predominate over any steady state motions. A pronounced patch of cold water (<10.75°C) evident in Figure 4 may be a manifestation of this type of localized upwelling.

**Discussion**

The model of potential energy production by mixing gives a satisfactory qualitative account of the distribution of stratification on the shelf and the occurrence of associated fronts. It suggests the general predominance of tidal mixing in controlling stratification in the Irish and Celtic seas. Wind stress does make a significant contribution to the total mixing, but because it is more spatially uniform than tidal stirring, it is of lesser importance in determining the location of fronts.

While the mean positions of the fronts are predicted by the mixing model, the detailed structure of a front is not. Clearly, the fronts are complex and variable and are controlled by poorly understood dynamical processes. The results presented above indicate that the nontidal motions in the vicinity of fronts may be relatively energetic (>10 cm/s) which is large in comparison with available estimates by drogues for nonfrontal regions of the Irish Sea (~2 cm/s).

These high speeds may occur in flow parallel to the front with a jet in the baroclinic region, or in the breakdown of such geostrophically balanced flows by a large-scale instability process which has been observed in the IR images of the Celtic Sea front. This front appears to be quite variable and convoluted in most of the available IR images. By contrast, the Islay front off the west coast of Scotland is found to be remarkably consistent in form, suggesting an orderly flow pattern.

Just why one front should be stable when the other is not is as yet unclear, although the difference could be due to the enhanced gradient of the parameter $h/a^2$, which probably ex-
ists in the vicinity of the Islay front. The detail of the $h/u^2$ distribution is still to be evaluated from recent current meter observations in the area, but in view of the marked change in depth across the frontal region and the tendency for $u$ to be inversely related to $h$, a rapid change in $h/u^2$ is to be anticipated. Such a high gradient would be expected to enhance the damping of any initial perturbation of the front.

An extended investigation of the IR imagery of these and other fronts seems desirable in order to establish the frequency and nature of these instabilities which, in bringing about crossfrontal transfer, may be one of the agencies affecting productivity changes at fronts.

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**References**


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