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A mechanism for local upwelling along the European continental slope

Upwelling
European continental slope
Wind-driven circulation
Mathematical model

Upwelling
Pente continentale européenne
Circulation engendrée par le vent
Modèle mathématique

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ABSTRACT

A simple theory is given, formulating an upwelling mechanism to account for the band of relatively cold surface water observed during summer months along the South-West European continental slope.
A radical alteration in turbulence conditions at thermocline level, on passing from shelf to ocean over the upper part of the continental slope, is inferred from observed isotherm patterns. Under such circumstances, it is shown that local upwelling occurs in the surface layer on the slope under the influence of an onshore wind.
The prevailing winds of the South-Western shelf area are correspondingly onshore and therefore it seems likely that such upwelling happens in reality. Thereby, colder water from below the thermocline could be brought up to the surface to produce the observed cooling.

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RÉSUMÉ

Un mécanisme possible pour l'upwelling local le long de la pente continentale européenne.

L'article propose une théorie simple suggérant un mécanisme d'upwelling pour rendre compte de la présence d'une bande d'eau de surface relativement froide, observée en été le long de la pente continentale au sud-ouest de l'Europe.
De l'observation des configurations d'isothermes, on conclut à un changement radical des conditions de turbulence au niveau de la thermocline, à la transition du talus continental. On démontre que dans ces circonstances se produit un upwelling local dans la couche de surface au-dessus de la pente, sous l'influence d'un vent dirigé vers le rivage.
Les vents prédominants au-dessus du plateau Sud-Ouest sont effectivement dirigés vers le rivage, et il est donc probable qu'un tel upwelling se produit en réalité. De ce fait, des eaux plus froides, provenant d'en dessous la thermocline, pourraient être ramenées à la surface et entraîner le refroidissement observé.

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INTRODUCTION

Infra-red images of sea surface temperature during the Summer (Pingree, 1979; Dickson, Gurbutt, Pillai, 1980) have shown a narrow band of relatively cold surface water following the edge of the European continental shelf from south-west of Ireland to the Bay of Biscay (Fig. 1). A physical explanation of this phenomenon is now being sought and, towards an understanding of the processes involved, the present paper gives a very simple first model, demonstrating the possibility of an upwelling mechanism which might account for the observed surface cooling.

The section of shelf edge under consideration (bounding the Celtic Sea in the North) has a Northwestward orientation and therefore, following conventional arguments, one might expect winds with a pronounced northerly component to produce any upwelling which

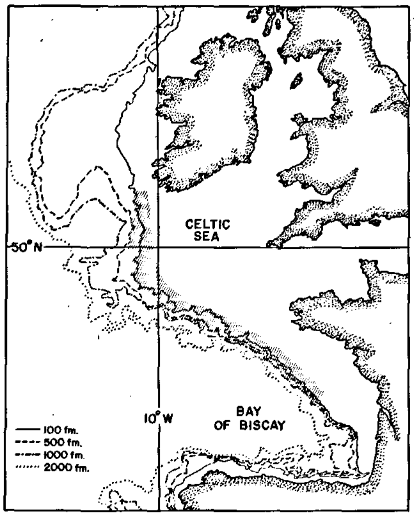

Figure 1
South-West part of the European shelf and the Bay of Biscay. The band of relatively cold surface water, observed along the shelf break during the summer, is shaded (*Pingree*, 1979).

might be present. However, while Northerly winds do occur in the region, the prevailing winds are from between West and South, and hence tend to blow normally across the line of the shelf break in an onshore direction. Thus, usual conditions (the so-called coastal Ekman divergence) for upwelling do not appear to be generally satisfied.

The main assumption in the following work is that turbulence conditions at thermocline level change quite radically on passing across the upper part of the continental slope in this area. Consideration is given to the Summer situation of thermal stratification, since the satellite observations mentioned above were taken during Summer months. An XBT section across the North Atlantic from the Scilly Isles to the Grand Banks, taken by Wegner (1979) in September 1978, clearly shows a fairly sharp thermocline in the Celtic Sea which fans out downwards quite rapidly on passing over the top of the shelf edge into the open ocean. All this happens below what is essentially an isothermal surface layer of uniform thickness 40 m (Fig. 2). Directly underneath and at the bottom of this layer, it therefore seems reasonable to conclude that there is a rapid decrease in vertical stability and a corresponding steep increase in vertical eddy viscosity on passing from shelf to open ocean across the upper part of the shelf slope. Such changes might be attributed mainly to the change in tidal conditions between shelf and ocean, with vertical shears relatively high on the shelf and relatively low in the ocean. Be this as it may, the main point to be made at present is that, at the base of the surface layer, it appears likely that there is a steep increase in eddy viscosity and hence friction coefficient, on moving from shelf to ocean, when considering wind-driven motion within the layer. It is demonstrated in this paper that such an increase in friction coefficient gives rise to local upwelling in the surface layer under an onshore wind. The upwelling is shown to be concentrated in the region where the horizontal gradient of the coefficient is greatest, supposed here to be at the top of the shelf slope. Thus, a mechanism is provided, by which colder water from below the thermocline, entrained into the bottom of the surface layer, may be brought up to the surface to produce localised cooling along the shelf break as observed.

Figure 2, illustrating the above argument, gives a diagrammatic representation of isotherms at the edge of the shelf, Celtic Sea, in a section normal to the line of the shelf break. As already mentioned, the pattern shown is based on observations by Wegner (1979). These observations lie approximately along the latitude 50°N in Figure 1.

Citing pertinent literature, theories for wind-induced upwelling over a shelf break have been given by Hill and Johnson (1974), Johnson and Killworth (1975), and Johnson and Manja (1979). These studies determined three-dimensional barotropic flow produced by a steady wind stress, and showed upwelling occurring in a shear layer above the shelf break. The possibility of shelf-break upwelling fostered by longshore currents at the break, has been investigated theoretically by Hsueth and Ou (1975). None of this earlier work has dealt with the upwelling mechanism of the present paper.

For the stratified situation on the European shelf, considered here, other authors (Simpson, Hughes, Morris, 1977; Simpson, Pingree, 1978) have related the stratification to tidal mixing. Their findings show that stratification occurs when a dimensionless parameter ε exceeds a critical value, delineating transition from a stratified to a vertically-mixed regime. Primarily, ε depends on depth and the amplitude of the surface tidal stream at springs; also on surface heat input and the frictional drag at the sea bed. In comparison, enhanced upwelling in the sense of the present theory is related to the horizontal gradient of a friction coefficient at thermocline level.

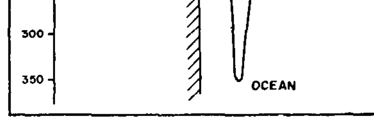

Figure 2
Diagrammatic representation of isotherms at the edge of the shelf, Celtic Sea, based on observations by Wegner (1979) *in September* 1978.

CIRCULATION MODEL

Consider a model of shelf and ocean with a straight vertical coast, longshore uniformity, and a cross-section normal to the shoreline as shown in Figure 3. Taking origin O at the coast in the undisturbed sea surface, let x denote the horizontal coordinate measured offshore, and z the vertical coordinate measured downwards. Here, the sign convention for coordinates is that employed by Proudman (1953). The problem is to determine the steady motion induced in a homogeneous surface layer of uniform depth h by an offshore wind stress F_s, applied uniformly over the water surface. Ignoring the effects of the Earth's rotation, the motion is confined to the (x, z)-plane with current components u, w in the x, z directions. Level $z=0$ represents the undisturbed sea surface, and level $z=h$ the top of the thermocline, supposed to remain horizontal. The wind-driven motion is restricted to the surface layer, making the assumption of no wind effects below $z=h$.

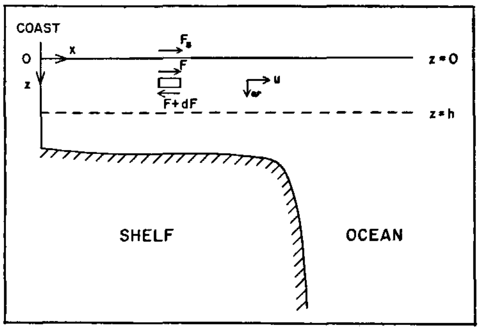

Figure 3
Section of shelf and ocean with a homogeneous surface layer of depth h.

Let F denote shear stress in the vertical (the only component of internal stress considered), p the pressure and ρ the density of the water, N a coefficient of vertical eddy viscosity regarded as constant and uniform, and p_a atmospheric pressure over the sea surface: uniformly distributed and invariable. Then adopting the usual hydrostatic approximation, the linear equations of steady wind-induced motion in the surface layer are:

$$\frac{\partial p}{\partial x}+\frac{\partial F}{\partial z}=0, \tag{1}$$

$$\frac{\partial p}{\partial z}=\rho g, \tag{2}$$

$$F=-\rho N\frac{\partial u}{\partial z}, \tag{3}$$

and equations of continuity are:

$$\frac{\partial u}{\partial x}+\frac{\partial w}{\partial z}=0, \tag{4}$$

$$\int_{-\zeta}^{h} u\,dz=0, \tag{5}$$

where ζ denotes the elevation of the sea surface above its undisturbed level. Equation (5) incorporates the condition of zero net horizontal flow at the vertical coast. Formulating conditions at the sea surface:

$$p=p_a \quad \text{at} \quad z=-\zeta, \tag{6}$$

$$F=F_s \quad \text{at} \quad z=-\zeta, \tag{7}$$

and at the bottom of the surface layer:

$$F=k\rho u \quad \text{at} \quad z=h, \tag{8}$$

$$w=0 \quad \text{at} \quad z=h. \tag{9}$$

The friction coefficient k at the base of the layer is taken to be some prescribed function of x, namely

$$k=k(x). \tag{10}$$

The main proposal made in this paper is that, due to changing circumstances of turbulence in the thermocline region immediately below the surface layer, k increases rapidly as x increases (oceanwards) over the edge of the continental shelf. The field of turbulence and hence k, may well depend importantly on the tidal flow and the changes which take place in it when passing over the shelf edge. Thus, on the shelf, it is envisaged that relatively high tidal velocities and the associated vertical shearing due to bed stress produce strong mixing in the lower layer-tending to sharpen the thermocline from below, thereby increasing vertical stability and suppressing k.

In the subsequent analysis, the above equations are solved for the wind-induced circulation in the surface layer. The effect on the circulation produced by $k(x)$-variation is determined.

ANALYSIS

Combining (2) and (6) yields the hydrostatic law:

$$p=p_a+\rho g(z+\zeta), \tag{11}$$

so that from (1):

$$\frac{\partial F}{\partial z}=-\rho g\frac{\partial \zeta}{\partial x}. \tag{12}$$

Integrating and satisfying (7) yields

$$F=F_s-\rho g\frac{\partial \zeta}{\partial x}(z+\zeta). \tag{13}$$

Therefore, from (3),

$$\frac{\partial u}{\partial z}=-\frac{F_s}{\rho N}+\frac{g}{N}\frac{\partial \zeta}{\partial x}(z+\zeta). \tag{14}$$

Further integration, satisfying (8), gives

$$u=\frac{F_s}{\rho k}\{1+\varkappa(1-\eta)\}-\frac{gH}{2k}\frac{\partial \zeta}{\partial x}\{2+\varkappa(1-\eta^2)\}, \tag{15}$$

where

$$H=h+\zeta, \qquad Z=z+\zeta \tag{16}$$

and

$$\eta=\frac{Z}{H}, \qquad \varkappa=\frac{kH}{N}. \tag{17}$$

Substituting (15) into (5) leads to the result

$$\frac{\partial \zeta}{\partial x}=\frac{3F_s}{2\rho g H}\left(\frac{2+\varkappa}{3+\varkappa}\right). \tag{18}$$

Then (18) in (15) gives

$$u=\frac{HF_s}{4\rho N}\left\{\frac{6(1-\eta)^2-2+\varkappa(1-\eta)(1-3\eta)}{3+\varkappa}\right\}. \tag{19}$$

To find w, from (4) and (9) we have

$$w=H\int_\eta^1 \frac{\partial u}{\partial x}\,d\eta. \tag{20}$$

Substituting u from (19) into this expression, regarding H and Z as independent of x since

$$H\sim h, \qquad Z\sim z, \tag{21}$$

it follows that

$$w=\frac{H^3 F_s}{4\rho N^2(3+\varkappa)^2}\left(\frac{dk}{dx}\right)\eta(1-\eta^2). \tag{22}$$

The wind-driven circulation in the surface layer ($0\leqq\eta\leqq1$) is defined by (19) and (22), which give, respectively, the horizontal (offshore) and vertical (down) components of current. The system is maintained by an offshore wind stress F_s.

Note that without the approximation (21), i. e. with H and Z given by (16), w comes out as in (22) but with the function $\eta(1-\eta^2)$ incremented by

$$\frac{3NF_s}{2\rho g H^3(dk/dx)}\left(\frac{2+\varkappa}{3+\varkappa}\right)(1-\eta)\{\varkappa\eta(1+\eta) \\ -(2+\varkappa)(3+\varkappa)(1-\eta)^2-2(3+\varkappa)\}.$$

Except in the immediate vicinity of the surface ($\eta=0$), this additional term may be shown, at least for values considered in this paper, to be negligibly small in comparison with $\eta(1-\eta^2)$. In this respect, the validity of (22) is confirmed.

DISCUSSION

For an onshore wind ($F_s<0$), it is clear from (22) that there is upwelling ($w<0$) throughout the depth of the surface layer where $dk/dx>0$. The vertical current attains a maximum

$$w_m=\frac{\sqrt{3}\,H^3 F_s}{18\rho N^2(3+\varkappa)^2}\left(\frac{dk}{dx}\right) \tag{23}$$

at depth $\eta=1/\sqrt{3}$. If therefore k increases rapidly with x over the upper part of the shelf edge, being sensibly uniform elsewhere, then according to present theory there is local upwelling on that upper part under an onshore wind.

As already argued, Figure 2 would appear to imply that k does increase in this way across the South-West European continental slope, where the prevailing winds are onshore. Hence, the mechanism of upwelling presented here might operate along that slope and account for the band of relatively cold surface water which has been observed there.

Consider the horizontal component of current u, given by (19). From a value u_0 at the sea surface ($\eta=0$) given by

$$u_0=\frac{HF_s}{4\rho N}\left(\frac{4+\varkappa}{3+\varkappa}\right), \tag{24}$$

it decreases with increasing depth to become zero at $\eta=\eta_a$, where

$$\eta_a=\frac{6+2\varkappa-(\varkappa^2+6\varkappa+12)^{1/2}}{6+3\varkappa}. \tag{25}$$

Above this level, the current is with the wind; below the level it is of opposite sign and against the wind reaching a maximum at $\eta=\eta_b$, where

$$\eta_b=\frac{6+2\varkappa}{6+3\varkappa}. \tag{26}$$

At the bottom of the surface layer, $u=u_1$ where

$$u_1=-\frac{HF_s}{4\rho N}\left(\frac{2}{3+\varkappa}\right). \tag{27}$$

If on passing from ocean to shelf across the upwelling region, k decreases from k_D to k_s say, and $\varkappa$ decreases correspondingly from $\varkappa_D$ to $\varkappa_s$, then from (24) and (27) the surface current increases by

$$\Delta u_0=\frac{HF_s}{4\rho N}\frac{\varkappa_D-\varkappa_s}{(3+\varkappa_D)(3+\varkappa_s)}, \tag{28}$$

and the bottom current by

$$\Delta u_1=-2\Delta u_0. \tag{29}$$

A diagrammatic representation of the wind circulation illustrating this feature is shown in Figure 4.

Speculating on representative numerical values relevant to conditions of upwelling at the oceanic edge of the Celtic Sea, we take

$H=40\,\text{m}$, $\quad N=400\,\text{cm}^2\times\text{sec}^{-1}$,

$F_s=-5\,\text{dyn.cm}^{-2}$ (corresponding to an onshore wind of 14 to 15 m.sec^{-1}),

$\rho=1.025\,\text{g.cm}^{-3}$,

$k_D=0.20\,\text{cm.sec}^{-1}$, $\qquad k_s=0.02\,\text{cm.sec}^{-1}$.

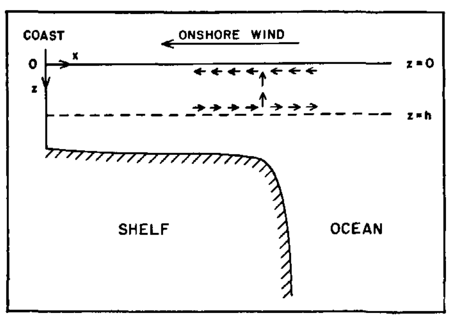

Figure 4

Diagrammatic representation of the circulation in the surface layer in the vicinity of the shelf edge due to an onshore wind. Thick arrows indicate the stronger currents.

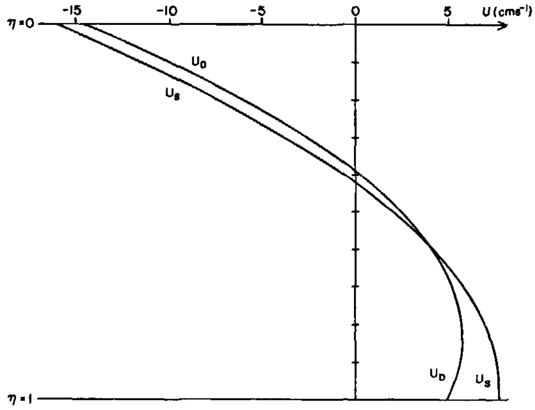

Figure 5
Depth profiles of horizontal current through the surface layer ($0<\eta<1$) *in the upwelling region:* $u=u_D$ *on ocean side,* $u=u_s$ *on shelf side.*

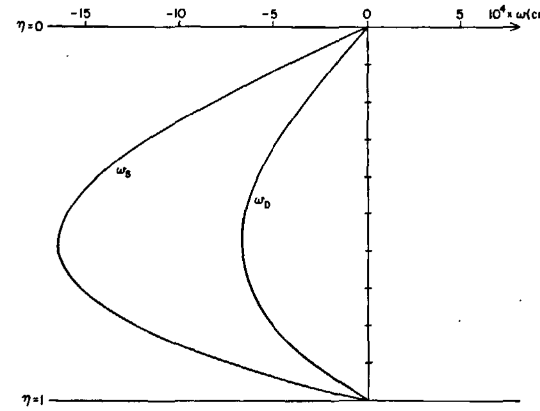

Figure 6
Depth profiles of vertical current through the surface layer ($0<\eta<1$) *in the upwelling region:* $w=w_D$ *on ocean side,* $w=w_s$ *on shelf side.*

If k increases linearly from k_s to k_D over a distance of say 20 km spanning the region of upwelling, then

$$\frac{dk}{dx}=\left(\frac{0.20-0.02}{20}\right)\times 10^{-5}=0.9\times 10^{-7}\,\mathrm{sec}^{-1}.$$

From (17):

$$\varkappa_D=2.0 \qquad \varkappa_s=0.2.$$

Therefore, using the above values in (19) and (22), on the ocean side of the upwelling region:

$$u=-2.44\{6(1-\eta)^2-2+2(1-\eta)(1-3\eta)\}, \quad (30)$$

$$w=-1.76\eta(1-\eta^2)\times 10^{-3} \quad (31)$$

and on the shelf side:

$$u=-3.81\{6(1-\eta)^2-2+0.2(1-\eta)(1-3\eta)\}, \quad (32)$$

$$w=-4.29\eta(1-\eta^2)\times 10^{-3}, \quad (33)$$

where currents are in centimetres per second. The depth profiles of u and w given by these equations, are plotted in Figures 5 and 6. Note the larger upwelling on the shelf side. Maximum values of u and $w\times 10^4$ are of order $15\,\mathrm{cm.sec}^{-1}$. Even when the winds are light, these currents may not be insignificant, being factored on present values by 1/5 for a wind of about $8.5\,\mathrm{m.sec}^{-1}$, and by 1/10 for a wind of about $7\,\mathrm{m.sec}^{-1}$.

CONCLUDING REMARKS

It is proposed that a mechanism of wind-induced upwelling operates on the south-west European continental slope, in the season of thermal stratification, due basically to a mismatch of turbulence conditions at thermocline level between shelf and ocean. The proposal originates from a physical interpretation of isotherm patterns at the shelf break. The mechanism is suggested in a first effort to explain the surface cooling observed along the line of this shelf edge.

Clearly there is a need for more realistic modelling than that given here. In this respect, further theoretical developments are envisaged including the effects of the Earth's rotation, considering a two-layered dynamical system, and introducing a better representation of coastal effects. The neglect of the Coriolis force is obviously serious for any practical application or for a full treatment of the theory. Hence, while not presenting a conclusive case for upwelling, the present paper nevertheless points to a possible cause of it.

In a general review of waves and currents near the continental shelf edge, Huthnance (1980) has suggested other reasons for enhanced vertical mixing near the shelf edge (more energetic internal motions, longshelf currents) which might be responsible for the observed surface cooling along the European continental slope. Largely, these ideas have still to be worked out. Also, it has been shown by Dickson, Gurbutt and Pillai (1980), that the cooling along the slope is consistent with a theory by Killworth (1978), concerning the enhancement of upwelling by interaction between slope topography and Kelvin (or other) waves propagating along the slope. Thus, the present research phase is one of initial exploration into the possibilities of motion along the slope, hopefully to be followed by more comprehensive programmes of observation and theory.

Concerning future experiments, long series of direct temperature measurements at the shelf edge and their comparison with the changing wind conditions should indicate the extent to which cooling of the surface water is associated with wind action. Simultaneous measurements of current would provide information on circulation and flow variations, with which the cooling might be connected. All this suggests that a long-term oceanographical experiment is needed to investigate, overall, the hydrography and dynamics at the shelf break.

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