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ON THE DISSIPATION OF AN UPPER OCEANIC FRONT

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ABSTRACT

This study seeks to gain some insight into the role played by horizontal friction in the dissipation of an oceanic front. For this purpose, a two layer model is used. The barotropic, or external, mode is filtered out. To simplify the equations, the wind stress terms are omitted. Two cases are considered: the coastal front and the upper ocean front. In the first case, it is shown that the larger the Ekman number, the lesser time it would take for the front to dissipate. In the second case, it is demonstrated that the average time scale of dissipation, due to horizontal friction, of an upper oceanic front is in the order of decades. Oceanic fronts do represent areas of horizontal convergence of different water masses. This convergence is due to the change of the curl of the wind stress pattern. Because of the fact that the wind stress is not considered in this study, it can be concluded that upper ocean fronts should be a permanent feature. This is the case of the Gulf Stream and the Kuroshio front.

RESUMEN

En el presente trabajo se estudia el papel jugado por la fricción horizontal en la disipación de un frente oceánico. A tales efectos, se ha utilizado un modelo de dos capas, en el cual el modo externo, o barotrópico, ha sido eliminado. A fin de simplificar las ecuaciones, los términos que representan la forzante del viento, no han sido incluídos en este trabajo. Se consideran dos casos: los frentes costeros y los de alta mar. En el primer caso, se demuestra que cuanto mayor es el número de Ekman, menor es el tiempo de disipación del frente costero. En el segundo caso, se demuestra que el tiempo promedio de disipación de un frente en alta mar es del orden de décadas. Los frentes de alta mar representan zonas de convergencia de masas de agua de distintas características. Dicha convergencia está asociada a las variaciones del rotor de la forzante del viento. Dado que la forzante de viento no ha sido considerada en este trabajo, se deduce que los frentes de alta mar deberán ser una característica permanente del océano. Tal es el caso de la corriente del Golfo y del frente de Kuroshio.

1. INTRODUCTION

Oceanic fronts occur on areas of increasing turbulence and convection. Their main characteristic are sharp gradients of salinity and/or temperature. It is a region of increasing motion (Roden, 1976). Usually oceanic fronts represent the discontinuity between two different water masses where one or more of the following properties: temperature, salinity, density and velocity, are different.

Oceanic fronts occur at all depths. Fronts are also variable in time. They intensify and decay, in response to changes in the atmospheric and oceanic flow patterns. Examples of permanent oceanic fronts are the Antarctic Circumpolar Current, the Subarctic and the Subtropical oceanic fronts in the North Pacific Ocean.

Oceanics fronts differ from their atmospheric counterpart in many ways. In the atmosphere, both the temperature and the density fronts coincide. Thus, atmospheric frontal zones are extremely baroclinic (Palmen & Newton, 1969). The first attempts in the understanding and the physical interpretation of atmospheric fronts started early in this century. The classical Norwegian school was the first one which attempted to understand this meteorological phenomenon. The studies of Bjerknes (1919) and Bergeron (1928) are classical in this matter. In the case of oceanic fronts, if the temperature and the salinity fronts do have the same geographical location, the density front is weak, or nonexistent. The baroclinicity is small. Such is the case of the subtropical Pacific Ocean Front (Roden, 1974).

Frontogenesis in the North Pacific is due to differential vertical and horizontal advection of the Ekman type. A given configuration of the wind stress field may lead to frontogenesis in some regions, and to frontolysis in other areas. Furthermore, in the upper layer of the ocean, the formation and maintenance of oceanic fronts depends of the horizontal shear of the wind stress (Roden, 1977).

Camerlengo (1982) studied the large scale response of the Pacific Ocean Subarctic Front to different forms of atmospheric forcing, i.e., atmospheric fronts, extratropical cyclones, wind stress curl, etc. His results agree with Roden's (1972) observations, in the sense that the dynamic response of the subarctic front to momentum transfer is limited to the layer between the sea surface and the high stability layer. The only exception was represented by the passage of a strong extratropical cyclone, where an upwelling of the order of 20 m is observed at the wake of the cyclone. Due to the fact that the time scale of dissipation of an upper oceanic front is an integral part of ocean dynamics, a series of theoretical experiments are conducted. This is the first attempt in this regard.

2. MODEL FORMULATION

2.1. Statement of the problem

A two layer model is considered in this study. The vertically averaged equations of conservation of mass and continuity (barotropic mode) are filtered out. For this purpose, the lower layer is chosen to be infinitely deep. Therefore, the currents, in that particular layer, are set to be zero. The model equations for this study are similar to MacVean &

Woods (1980) and Camerlengo (1982).

The analysis was limited to the first layer, or the first baroclinic mode. This type of model is, sometimes, referred as the one a half layer model (Busalacchi & O'Brien, 1980). The east-west and north-south velocity components, u and v, correspond to the horizontal coordinates x and y (positive in the east and north direction, respectively).

To simplify the equations to be used, the following hypotheses are made: 1) the time scale of dissipation of the oceanic front is larger that the time scale of the perturbation produced by the wind stress; 2) the longitudinal frontal length scale is smaller than the latitudinal length scale. Thus, the derivatives with respect to x are neglected as compared to the derivatives with respect to y; 3) the oceanic front is geostrophically balanced; 4) since the meridional length scale of the oceanic front in the upper layer has an order of magnitude of ten kilometers, the f-plane approximation is used; 5) the zonal wave number, k, is much smaller than 1/L. Therefore, only long waves are considered; and, 6) the vorticity at the front, represented by $\partial u/\partial y$, is much smaller than the planetary vorticity, f.

The pressure in the upper layer, layer 1, will have the form

$$P = P_\alpha + \rho_1 g (h_1 - z) + \rho_2 g (H - h_1) \qquad (1)$$

where P_α represents the atmospheric pressure; h_1, the depth of the upper layer; ρ the water's density; H, the total depth; g, the acceleration due to the earth's gravity; and z, the vertical coordinate, which is measured from the bottom. The gradient of P_1 yields

$$\nabla P_1 = - \rho_1 g^* \nabla h_1 \qquad (2)$$

where $g^* = g (\rho_1 - \rho_2)/\rho_1$ represents the reduced gravity. As usual, the horizontal fluctuations of the atmospheric pressure are neglected. For convenience, the subindex 1, representing the upper layer, will be omitted.

2.2. Equations of motion

With the above assumptions, the equations of motion for the upper layer are:

$$\partial u/\partial t + v \partial u/\partial y - f v = A \partial^2 u/\partial y^2 \qquad (3)$$
$$\partial v/\partial t + v \partial v/\partial y + f u = - g^* \partial h/\partial t + A \partial^2 v/\partial y^2 \qquad (4)$$
$$\partial h/\partial t + \partial (vh)/\partial y = 0 \qquad (5)$$

where f, is the Coriolis parameter; A, the horizontal diffusion coefficient of momentum; h, upper layer thickness and t, time. In our study, the forcing due to the wind stress terms are neglected in order to retain only the forcing due to the frictional terms.

This non-linear set of equations are difficult to solve. Therefore, it is assumed that the frictional effect will slowly change a steady, inviscid initial front, and a perturbation solution is presented by assuming that friction has a weak effect. For this steady oceanic front to persist, the interface, h , must be balanced initially by a zonal velocity of the form

$$u_o = - (g^*/f)\, \partial h_o / \partial y \tag{6}$$

where the subscript "o" stands for the basic state.

The height of the upper (first) layer, h , has an hyperbolic profile, centered in the middle of the north-south extent of the basin, W, of the form:

$$h = h \quad - \Delta h \tanh ((y - W/2)/L) \tag{7}$$

where h represents the mean value depth; L, the longitudinal frontal length scale; Δh, the maximum amplitude of the interface perturbation. By continuity, the meridional velocity, v , is initially set to zero.

3. ANALYSIS OF THE PROBLEM

The viscous perturbation solutions are obtained by assuming the value of A to be very small. Let

$$\begin{aligned} u &= u \ + u' \\ v &= v \ + v' \\ h &= h \ + h' \end{aligned} \tag{8}$$

where the perturbation variables are denoted by primes. The perturbation variables are assumed to be of order A (Pedlosky, 1979). The equations for the upper layer then become

$$\partial u'/\partial t - (v' g^*/f)\, \partial^2 h_o/\partial y^2 + v' \partial u'/\partial y - f v' = - (A g^*/f)\, \partial^3 h_o/\partial y^3 + A\, \partial^2 u'/\partial y^2 \tag{9}$$

$$\partial v'/\partial t + v' \partial v'/\partial y + f u_o + f u' = - g^* \{ \partial h_o/\partial y + \partial h'/\partial y \} + A\, \partial^2 v'/\partial y^2 \tag{10}$$

$$\partial h'/\partial t + \partial (h_o v')/\partial y + \partial (h' v')/\partial y = 0 \tag{11}$$

The terms $v' \partial u'/\partial y$, $v' \partial v'/\partial y$, $\partial (h' v')/\partial y$, $A\, \partial^2 u'/\partial y^2$ and $A\, \partial^2 v'/\partial y^2$ are of second order. Thus, they are considered negligible. The terms fu and -g*dh /dy represent the basic state balance. In order to filter out the gravity waves, the term $\partial v'/\partial t$ is omitted (Pedlosky, 1979). With these simplifications, the above equations are reduced to linear perturbation equations of the form

$$\partial u'/\partial t - (v' g^*/f)\ \{\partial^2 h_o/\partial y^2 + (f^2/g^*) = -(Ag^*/f)\partial^3 h_o/\partial y^3 \tag{12}$$

$$u' = -(g^*/f)\ \partial h'/\partial y \tag{13}$$

$$\partial h'/\partial t + \partial (h_o v')/\partial y = 0 \tag{14}$$

To solve this system, a single equation for v' is first needed. Replacing the value of u' in (12) yields

$$-\ (g^*/f)\{(\partial^2 h'/\partial y \partial t) - (v'\ g^*/f)\{\partial^2 h_o/\partial y^2 + (f^2/g^*)\} = -\ (A\ g^*/f)\ \partial^3 h_o/\partial y^3 \tag{15}$$

Applying the operator $(g^*/f)\ \partial(\ )/\partial y$ to (14), it follows that

$$(g^*/f)\{\ (\partial^2 h'/\partial t\ \partial y) + \ \partial^2\ (h_o v')/\partial y^2\ \} = 0 \tag{16}$$

The addition of (15) and (16) yields

$$\partial^2 V/\partial y^2 - \{\partial^2 h_o/\partial y^2 + (f^2/g^*)\}/h_o\ V = -\ A\ \partial^3 h_o/\partial y^3 \tag{17}$$

where V (=v'h) is the mass transport in the upper layer. The boundary conditions are such that the meridional mass transport, V, is set to be zero at both the northern and southern boundaries. It is of interest to note that V is independent of time.

As h (y) is a known function, the coefficients of (17) are known. Therefore, this equation can be rewritten as

$$\partial^2 V/\partial y^2 - F(y)\ V = -\ G(y) \tag{18}$$

where

$$F(y) = \{\partial^2 h_o/\partial y^2 + (f^2/g^*)\}/h_o \tag{19a}$$

and

$$G(y) = A\ \partial^3 h_o/\partial y^3 \tag{19b}$$

In order to cover a wider range of oceanic fronts, two different meridional length scales, L, are considered. The chosen dimensions are 1 and 50 kilometers, respectively. In the first case, an approximate value of F(y) is:

$$F(y) = (f^2/g^* h_o) \tag{20}$$

while in the second case, F(y) has an approximate value of:

$$F(y) = \{\partial^2 h_o/\partial y^2\}/h_o \tag{21}$$

Using the first approximation, (18) may be rewritten as:

$$\partial^2 V/\partial y^2 - \alpha^2 V = - A\, \partial^3 h_o/\partial y^3 \tag{22}$$

where $\alpha^2 = f^2 (g^* h_o)^{-1}$ is the inverse of the squared Rossby radius of deformation. For this particular case, an analytical solution can be obtained. If the Ekman number is set to be equal to 0.1, the resulting value of A is then 10 m sec. The scaling of (22) shows that the first and third terms are the largest ones, by two orders of magnitude. Therefore, (22) has the form

$$\partial^2 V/\partial y^2 = - (2 A \Delta h/L^3)\{[2 \sinh^2((y - 0.5 W)/L) - 1]/\cosh^4((y - 0.5 W)/L)\} \tag{23}$$

The analytical solution of this expression is

$$V(y) = - (A \Delta h/L) \operatorname{sech}^2((y - 0.5 W)/L) \tag{24}$$

The integration, with respect to time, of the continuity equation yields a perturbation expression for the interface, h', of the form

$$h' = (\partial V/\partial y)\ t \tag{25}$$

Considering the maximum absolute value of the y-derivative of the meridional mass transport, the time scale of dissipation, T, is defined in such a way that

$$T = h' \,|\partial V/\partial y|^{-1}_M \tag{26}$$

where $|\partial/\partial y|_M$ represents the maximum absolute of such derivative. The value of h' is arbitrarily set up to be equal to 10 meters. For different values of A and L, different values of $|\partial V/\partial y|_M$ may be obtained (Table 1).

However, this solution is restricted to oceanic fronts of the order of one kilometer.

As stated previously, the upper oceanic fronts have a meridional length scale of the order of ten kilometers. A new scaling of (22), using this same length scale, shows that the three terms of the equation are comparable. Thus, the full equation has to be solved.

A (m sec^{-1})	L (km)	$\lvert \partial V/\partial y \rvert_M$ (m sec^{-1})	T (sec)
10	2	$2\ 10^{-5}$	$5\ 10^{5}$
10	1	$8\ 10^{-5}$	$1.25\ 10^{5}$
1	2	$2\ 10^{-6}$	$5\ 10^{6}$
1	1	$8\ 10^{-6}$	$1.25\ 10^{6}$

Table 1. Time scale of dissipation, T, for different values of A, L,$\lvert \partial V/\partial y \rvert_M$, using equation (22).

In using a meridional length scale of 50 kilometers, (22) has the form:

$$\partial^2 V/\partial y^2 \ - \ (\partial^2 h_o/\partial y^2)\ (V/h_o) = -A\, \partial^3 h_o/\partial y^3 \tag{27}$$

The general solution of the corresponding homogeneous equation (27), Vc(y), is

$$Vc(y) = C_1 V_1 + C_2 V_2 \tag{28}$$

where V = h , V = h Ii h dy and C and C are arbitrary constants. The method of variation of parameters is used to solve the nonhomogeneous equation (22). This method requires the replacement of the constants C and C by two arbitrary functions, c(y) and z(y). These arbitrary functions will be determined in such a way that the particular solution, V (y), has the form

$$V(y) = \chi(y)\, V_1(y) + \zeta(y)\, V_2(y) \tag{29}$$

The arbitrary function, c and x, must satisfy certain conditions in order to satisfy the nonhomogeneous ordinary differential equation. These conditions are

$$\chi_{,y} V_1 + \zeta_y V_2 = 0 \tag{30a}$$

$$\chi_{,y} V_{1y} + \zeta_y V_{2y} = \psi(y) \tag{30b}$$

where χ_y, ζ_y, V_{1y} and V_{2y} represent the meridional derivatives of the functions χ , ζ,

V_1 and V_2, respectively; and $\psi(y)$ represents the forcing function defined by

$$\psi(y) = -(2 A \Delta h / L^3) \{[2 \sinh^2((y - 0.5 W)/L) - 1] / \cosh^4((y - 0.5 W)/L)\} \qquad (31)$$

The system of equation (30) gives us a solution for the functions χ_y and ζ_y.. These values are:

$$\chi_y = - V_2(y)\ \psi(y) / W(V1, V_2) \qquad (32a)$$

$$\zeta_y = - V_1(y)\ \psi(y) / W(V1, V_2) \qquad (32b)$$

where the wroskian, $W(V1, V_2)$, is equal to one.

The values of χ and ζ, at each grid point, are determined by numerical integration. To achieve this purpose, Simpson's rule is used. Knowing the values of χ and ζ, a final expression for the meridional mass transport, V, can be numerically evaluated at each grid point. Such an expression has the form

$$V(y) = (C_1 + \chi) V_1(y) + (C_2 + \zeta) V_2(y) \qquad (33)$$

Again, by plotting the function $\partial V / \partial y$ versus y, a maximum absolute value of the former function may be determined for different values of A and L (Table 2).

A (m sec)	L (km)	$\lvert \partial V / \partial y \rvert_M$ (m sec^{-1})	T (sec)
1	10	5 10	10^9
1	50	7 10	10^{11}
10	10	5 10	10^8
10	50	7 10	10^{10}
100	10	5 10	10^7
100	50	7 10	10^9

Table 2. Same as table 1, but using equation (27).

3. CONCLUSIONS

Kao (1980) showed that the structure of an oceanic front in a quasi-steady state depends on the Ekman number, E. Furthermore, in that same study, it is shown that for E << 1, the structure of the front is insensitive to changes in the value of E.

A study dealing with the dissipation of an oceanic front due to horizontal friction has not yet been conducted. This is a first such an attempt. In our study, the value of E was allowed to vary from 10 to 10. In choosing this particular range, the cases of coastal front and upper ocean front are addressed.

Due to the fact that oceanic fronts occur in all horizontal space scales, the value of L is allowed to vary between 1 and 50 km. In the first case, we attempt to study the dissipation of coastal fronts. Therefore, the chosen value of L is in the order of 10 m (Table 1). On the other hand, in the second case, we attempt to study the dissipation of an upper ocean front (Table 2). In this case, L is chosen to be on the order of 10 m (Roden, 1976).

In the first case, A is chosen to vary between 1 and 10 m sec. Due to the almost constancy in the (chosen) value of L, the value of the Ekman number (E = A/(f L)) highly depends on the value of A. In our case, the value of E varies from 10 to 10. An uppermost value of A = 10 m sec is chosen such that E ¾ 10. It is shown that the larger the value of E, the smaller the value of T (Table 1). It is concluded that the larger the value of horizontal friction, the lesser time it would take for the coastal oceanic front to dissipate.

In the second case, we attempt to study the dissipation of an upper ocean front (Table 2). In this case, the values of A are chosen to vary between 1 and 100 m s. The resulting Ekman number varies between 10 to 10. This range in the value of E corresponds to the large scale ocean circulation (Pond & Pickard, 1978). For the smallest value of E, T would be in the order of one year.

Forcing due to the wind stress terms are neglected in this study. However, the existence of the Pacific Ocean subarctic front is a natural consequence of the convergence of two different water masses. This convergence is triggered by the wind stress pattern. The average time scale of dissipation (due to horizontal friction) of upper ocean fronts is in the order of decades (Table 2). It can concluded that upon consideration of the wind stress terms, upper ocean fronts should be a permanent feature in the ocean. This is precisely what happens in the real world. Examples of such permanent fronts are the Kuroshio front, the Gulf Stream, the subtropical and the doldrums fronts.

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