Cooling Parsons' Model of the Separated Gulf Stream

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ABSTRACT

The effect of cooling on the separated boundary current predicted by the model of Parsons is studied. The separating current is found to strengthen and to move southwards and eastwards. The model is also robust to limited heating, in which case the separating current weakens and moves northwards.

1. Introduction

Parsons (1969) has elegantly explained the separation of the western boundary current in terms of the transport balances in a subtropical gyre. The position of the boundary current is determined by balancing the transports of the Ekman, geostrophic and boundary current flows for a single layer of light fluid overlying a deep abyssal layer. Veronis (1973) extended this model to include subtropical and subpolar gyres and diagnosed the global circulation by imposing a realistic wind stress field. He further deduced the likely diabatic sources and sinks of light water needed to give the observed separation points of the boundary currents (Veronis 1978). Huang and Flierl (1987) examined the structure of the boundary currents required in an adiabatic model and showed them to be dynamically consistent, while Huang (1986) confirmed the theory using a numerical model. However, Pedlosky (1987) has recently cast doubt on the Parsons' mechanism by showing its sensitivity to a special form of diabatic forcing. He incorporates a surface heating that allows the Ekman flux to cross the outcrop line and finds that the boundary current disappears.

In this work, we consider further how robust Parsons' model is to diabatic forcing over the separated jet. We are particularly interested in the effect of surface cooling concentrated along the outcrop, as observations show intense cooling of the Gulf Stream (Bunker 1976). We also consider the influence of an artificial surface heating for comparison with Pedlosky's work.

We review Parsons' separation mechanism in section 2 and concentrate on the dynamical regime in which separation of the boundary current occurs in a subtropical/subpolar double gyre: Huang and Flierl describe this as the supercritical state I. Then, we examine the influence of diabatic forcing on the separated current in section 3 and present conclusions in section 4.

2. Parsons' Mechanism

a. Adiabatic transport balances

The separation mechanism in a subtropical/subpolar double gyre is examined using a model in which the stratified ocean is represented by a single dynamically active layer of density ρ overlying a deep abyssal layer of density $\rho + \Delta\rho$. The model is forced by a kinematic wind stress that is zonal and varies as $\tau = -\tau_m \times \cos(2\pi y/y_n)$ in a rectangular basin on a β-plane with $x = 0$, x_e on the western and eastern boundaries and $y = 0$, y_n on the southern and northern boundaries (Fig. 1).

The western boundary current is forced to leave the west coast and to follow a path $X_p(y)$ in order to satisfy the transport balances for the light water. If the separated jet extends into the subpolar gyre, the circulation must be completed by a southward return flow, T_{iso}, in an isolated western boundary current and/or by incorporating a diabatic source and sink circulation (Veronis 1973, 1978).

We define the northward transport, T_{north}, of light water to be made up of the Ekman flux, T_e, the interior geostrophic flux, T_i, and the western boundary current flux T_w:

$$T_{north} = T_e + T_i + T_w. \quad (1)$$

This is the total transport over the light water domain between the outcrop line, $X_p(y)$, and the eastern wall, x_e, and does not include any return flow, T_{iso}, carried in the isolated western boundary current. The Ekman flux over the light water is given by

$$T_e = -(x_e - X_p)\frac{\tau}{f}. \quad (2)$$

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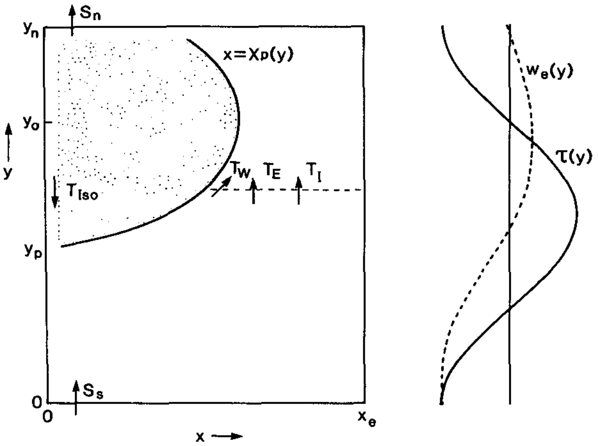

FIG. 1. A schematic plan view of a subtropical/subpolar double gyre showing the outcrop line $x = X_p(y)$ separating the light and abyssal water (stippled). Over the light water domain, there is an Ekman transport T_e, an interior geostrophic transport T_i and a western boundary transport T_w. In order to complete the circulation, there also has to be a southward return transport, T_{iso}, in an isolated western boundary current and/or a source of light water, S_s, on the southern boundary and a sink, S_n, on the northern boundary. The zonal wind stress, $\tau(y)$, and the Ekman suction velocity, $w_e(y)$, are shown on the right.

The interior geostrophic flux is

$$T_i = \frac{g'}{2f}[h_e^2 - h_i^2(X_p)] \tag{3}$$

where the thickness of the light water in the interior, h_i, predicted from the Sverdrup balance satisfies

$$h_i^2(x) = h_e^2 - \frac{2f^2}{\beta g'} w_e(x_e - x). \tag{4}$$

Here, h_e is the constant thickness of the light water on the eastern boundary, $g' = g\Delta\rho/\rho$ is the reduced gravity, f is the planetary vorticity and β is its meridional gradient. Presuming that the downstream velocity is geostrophic, then the western boundary current flux is

$$T_w = \frac{g'}{2f}(h_i^2(X_p) - h_w^2). \tag{5}$$

Here, h_w is the thickness of the light water on the western boundary.

While the boundary current is attached to the coast, with $X_p(y) = 0$, the transport balance for the light water is given by combining (1) to (5):

$$\frac{g'}{2f}(h_e^2 - h_w^2) - x_e\frac{\tau}{f} = T_{\rm north}(y). \tag{6}$$

Assuming that $T_{\rm north}(y)$ is constant here, then the thickness of the light water, h_w, on the western boundary decreases as the eastward wind stress increases with latitude in the subtropical gyre.

If the wind stress becomes sufficiently large, then the light water outcrops, $h_w = 0$, and the western boundary current leaves the coast at a latitude y_p given by

$$\tau(y_p) = \frac{g'h_e^2}{2x_e} - \frac{f(y_p)}{x_e} T_{\rm north}(y_p). \tag{7}$$

The separated boundary current subsequently swings eastward so as to reduce the southward Ekman flux by narrowing the width of the light water, $x_e - X_p(y)$, and hence provides the required northward transport, $T_{\rm north}(y)$:

$$\frac{g'}{2f}h_e^2 - (x_e - X_p)\frac{\tau}{f} = T_{\rm north}(y). \tag{8}$$

If the separated jet crosses the zero wind latitude in the subpolar gyre (ZWL, at $y = y_0$, $f = f_0$), then since the Ekman flux disappears there, the northward transport is set simply by the interior geostrophic flow:

$$T_{\rm north}(y_0) = \frac{g'}{2f}h_e^2. \tag{9}$$

This result is crucial for the theory of separation in a subtropical/subpolar double gyre. It still holds even if diabatic forcing is incorporated.

In an adiabatic model, light water is conserved so that the northward transport everywhere in the separated regime has to be constant and equal to the transport across the ZWL:

$$T_{\rm north}(y) = T_{\rm north}(y_0) \quad y_p \leqslant y \leqslant y_n. \tag{10}$$

This northward transport, $T_{\rm north}(y_0)$, in the separated domain, has to be supplied somehow. In a purely adiabatic model, no net flow is allowed across the southern boundary,

$$T_{\rm north}(y) = 0 \quad 0 \leqslant y < y_p \tag{11}$$

Therefore, the northward transport in the separated region (10) has to be balanced by a southward transport, $T_{\rm iso}$, in an isolated western boundary current (Fig.1):

$$T_{\rm iso} = T_{\rm north}(y_0). \tag{12}$$

This isolated boundary current might represent the southward flows in the Labrador or Oyashio currents. The isolated boundary current returns the light fluid back into the rest of the basin at the separation point, y_p, which leads to a jump in the northward transport, $T_{\rm north}(y)$; separation occurs suddenly with an abrupt decrease to zero in the thickness of light water on the coast.

Separation of the boundary current occurs for a weaker wind stress in (7) than in Parsons' original model of an isolated subtropical gyre, as the separated flow has to carry the northward transport, $T_{\rm north}(y_0)$. Therefore, separation is now more likely and should occur farther south in a subtropical/subpolar double gyre.

b. Consistency condition

These transport balances cannot, in general, be satisfied unless there is a boundary current along the separation line. However, the northward geostrophic flow in the subpolar gyre implies a westward shallowing of the thermocline. This may, itself, lead to outcropping with $h_i = 0$ in (4) at a longitude $X_{geo}(y)$ where

$$w_e(x_e - X_{geo}) = \frac{\beta g' h_e^2}{2f^2}. \qquad (13)$$

There is a danger that this geostrophic surfacing line may occur eastward of the Parsons' surfacing line, in which case no boundary current could exist. Therefore, we require for an adiabatic model [combining (8) and (13)] that

$$X_p - X_{geo} = \frac{g' h_e^2}{2}\left(\frac{\beta}{f^2 w_e} - \frac{1}{\tau}\right) + \frac{f}{\tau} T_{north}(y) \geq 0. \qquad (14)$$

Huang and Flierl show that this outcropping condition (14) will be satisfied by using the appropriate $T_{north}(y)$ given by (9), as long as $\partial^2\tau/\partial y^2$ changes sign from negative just south of the ZWL to positive just north of it. In this case, the boundary current disappears at the zero wind latitude and $X_p(y_0) = X_{geo}(y_0)$. Huang has shown in a numerical model that this rather peculiar restriction upon the form of τ is an artifact arising from the assumption of semigeostrophy in the separated jet, which becomes questionable for the broad shallow flow that develops near the zero wind latitude.

c. *Adiabatic results*

In Fig. 2 the path of the separated current, $X_p(y)$, and the thickness of light water are shown for an outcropping case in this adiabatic theory, together with the geostrophic surfacing line, $X_{geo}(y)$, using reasonable wind stresses and thermocline depths (Table 1). The boundary current separates from the west coast in the northern part of the subtropical gyre and swings eastwards. This decreases the width of the light water domain, $x_e - X_p$, and hence the Ekman flux of light water, so as to give the required northward transport. The separated current leads to a marked asymmetry between the circulation in the subtropical and subpolar gyres.

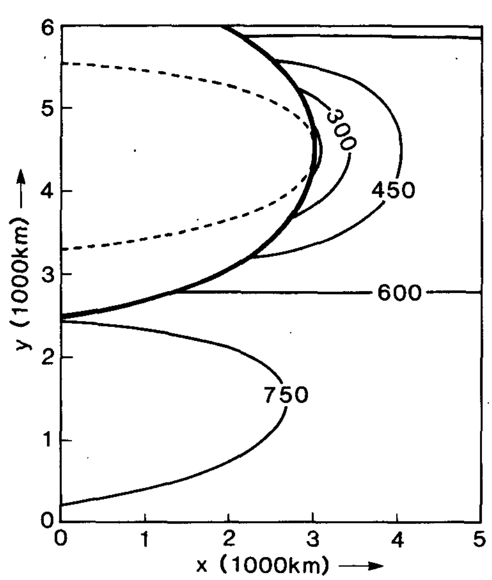

FIG. 2. The path of the separated boundary current, $X_p(y)$, (thick full line) is shown together with the thickness field of the light water (full line) predicted by the Parsons' adiabatic theory in a subtropical/subpolar double gyre. The geostrophic surfacing line, $X_{geo}(y)$, (dashed line) from Ekman suction in the subpolar gyre is also included.

For this model, the western boundary current transport T_w reaches a maximum of 41 Sverdrups (Sv $\equiv 10^6$ m^3 s^{-1}) at the maximum wind stress latitude in the subtropical gyre ($y = y_n/4$) and then decreases to zero at the zero wind stress latitude in the subpolar gyre ($y_o = 3y_n/4$). Here, the northward transport, $T_{north}(y_o)$, is set by the interior geostrophic flow as 17 Sv. In this adiabatic case, this is also the magnitude of the southward return flow in the isolated boundary current, T_{iso}.

d. *A source/sink interpretation*

The conservation of light water over the whole domain can be alternatively achieved by incorporating sources and sinks of light water (Veronis 1978). A source of light water, S_s, can be permitted on the southern boundary together with a sink S_n on the northern boundary of the basin (Fig. 1):

$$S_s = T_{north}(y = 0)$$
$$S_n = T_{north}(y_n) - T_{iso}. \qquad (15)$$

If there is no diabatic forcing within the interior of the model, then light water simply flows from the southern source to the northern sink with conservation of light fluid requiring that $S_s = S_n$. The return flow, T_{iso}, in the isolated boundary current is weakened by cooling on the northern boundary that converts the light fluid into abyssal fluid: the isolated boundary current, T_{iso}, then vanishes in the limit of $S_s = S_n = T_{north}(y_0)$.

This is perhaps a more realistic picture of the circulation in the North Atlantic than the purely adiabatic circulation composed of a Gulf Stream and Labrador Current. The value of, $T_{north}(y_n)$, is 17 Sverdrups, which is of the same order as the 10–15 Sv estimated to make up the thermohaline circulation (McCartney and Talley 1984).

3. Diabatic forcing of Parsons' model

a. *The heat balance*

The circulation model has so far been completely adiabatic and no net flow has been allowed to cross the outcrop line. This assumption may be too restrictive because the separated Gulf Stream is in fact observed to experience strong cooling. We are therefore interested in how diabatic forcing modifies the results of section 2.

TABLE 1. Model parameters.

Variable	Symbol	Value
Maximum kinematic wind stress	τ_m	1.4×10^{-4} m^2 s^{-2}
Maximum Ekman upwelling	w_m	30 m yr^{-1}
Zonal extent	x_e	5000 km
Meridional extent	y_n	6000 km
Density difference	$\Delta\rho$	1.02 kg m^{-3}
Reduced gravity	g'	0.010 m s^{-2}
Thickness of light water on the eastern boundary	h_e	600 m

Suppose a surface heat flux, $Q_{\rm in}$, is applied to a surface boundary with a vertically homogeneous density, ρ_m, and thickness, h_m (Fig. 3). We choose to separate the velocity field in this boundary layer into an Ekman flow, $\mathbf{v}_e$, (which balances the wind stress) and a mixed-layer flow, $\mathbf{v}_m$, (which balances the pressure gradient and any remaining inertial and frictional terms). For a steady state, the surface heating balances the horizontal advection by both these flows in the surface boundary layer:

$$\int_{-h_m}^{o} (\mathbf{v}_e + \mathbf{v}_m)\cdot\nabla\rho_m dz = -\frac{\alpha}{C_w} Q_{\rm in}. \quad (16)$$

Here, $\alpha = 2.2 \times 10^{-4}$ K^{-1} is the expansion coefficient and $C_w = 4000$ J kg^{-1} K^{-1} is the heat capacity of seawater.

In Parsons' adiabatic model, the surface heating is set to zero, which implies that the Ekman flow, $\mathbf{v}_e$, across the outcrop exactly balances an opposing mixed-layer flow, $\mathbf{v}_m$, in the surface boundary.

In the Pedlosky model, a particular surface heating field is chosen in order to allow an Ekman flow, $\mathbf{v}_e$, across the outcrop line, but any residual mixed-layer flow, $\mathbf{v}_m$, is prevented from crossing the outcrop: this significantly modifies the Parsons' solutions for the separated domain.

We instead prefer to let a surface cooling or heating allow a net flow across the outcrop, which is due to a combination of both the Ekman, $\mathbf{v}_e$, and mixed-layer $\mathbf{v}_m$ flows. The outcrop line is assumed here to only be a barrier to the geostrophic flow in the underlying adiabatic interior.

In this layered model, the cooling or heating is concentrated at the outcrop separating the two layers, and the density gradient is simply taken as the density difference $\Delta\rho$ across the separated current of width Δl. The mixed layer is assumed to have a thickness, h_m, that is relatively small compared with the thickness of the light water on the eastern wall, h_e. Surface cooling then gives a conversion of light to abyssal water in the mixed layer with the interfacial flux/unit length of outcrop given from (16) by

$$U_c = \left(-\mathbf{k}\times\frac{\boldsymbol{\tau}}{f} + h_m\mathbf{v}_m\right)\cdot\mathbf{n} = \frac{-\alpha Q_{\rm in}}{C_w\Delta\rho/\Delta l}. \quad (17)$$

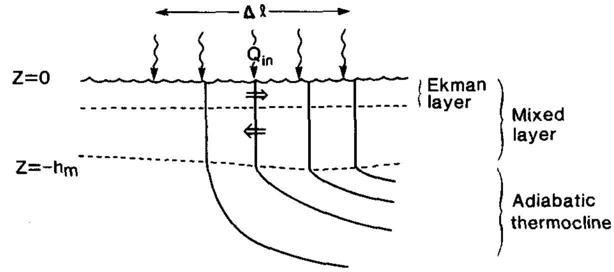

FIG. 3. A schematic section showing isopycnals outcropping into a mixed-layer of thickness, h_m, from an adiabatic thermocline. The surface heat flux, $Q_{\rm in}$, applied over a width Δl allows a cross-isopycnal flow in the Ekman layer and the mixed layer.

Here, $\mathbf{n}$ is the unit normal out of the light water and $\mathbf{k}$ is the unit normal upward.

b. Diabatic transport balances

Cooling or warming of the separated Gulf Stream produces an interfacial flux across the outcrop (Fig. 4). This implies that the northward transport, $T_{\rm north}$, is no longer constant after separation, but rather varies with the distance s along the outcrop according to

$$\frac{d}{ds} T_{\rm north} = -U_c. \quad (18)$$

If a steady state is to be maintained, there can be no net conversion of light to dense water over the whole basin. We assume, for simplicity, that the flux of light to dense water along the outcrop is compensated by an equatorial source of light water, S_s, of greater strength than the polar sink, S_n, so

$$S_s - S_n = \int_{y_p}^{y_n} U_c ds. \quad (19)$$

At the zero wind latitude, the northward transport is still set by the interior geostrophic flow and is the

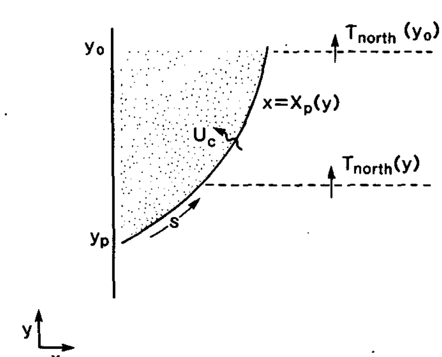

FIG. 4. Cooling along the outcrop line, $x = X_p(y)$, leads to an interfacial flux/unit length U_c from the light to abyssal water (stippled). This results in the northward transport, $T_{\rm north}(y)$, varying from its fixed value, $T_{\rm north}(y_0)$, at the zero wind latitude, y_0, in the subpolar gyre.

same as in the adiabatic case (9) denoted here by $T_{\mathrm{adiabatic}} = T_{\mathrm{north}}(y_0)$. The northward transport, T_{north}, is therefore

(i) a constant value, S_s, south of separation,

$$T_{\mathrm{north}}(y) = S_s \quad 0 \leqslant y < y_p,$$

(ii) a variable value elsewhere, differing from the constant adiabatic value according to the diabatic forcing (integrating (18) from the ZWL) as

$$T_{\mathrm{north}}(y) = T_{\mathrm{adiabatic}} + \int_y^{y_0} U_c ds, \quad y_p \leqslant y \leqslant y_n. \quad (20)$$

Suppose there is a net cooling and a conversion of light to dense water between a given latitude and the ZWL. If that latitude is *south* of the ZWL, then the northward transport, $T_{\mathrm{north}}(y)$ there must be *greater* than at the ZWL (and, hence, greater than the adiabatic value). Conversely, cooling north of the ZWL implies that the northward transport there is less than at the ZWL.

c. *Cooling the Gulf Stream*

Climatological observations show that the Gulf Stream suffers annual mean heat losses of typically 200 W m^{-2} (Bunker 1976). In the model, this surface cooling is applied along the first 2000 km of the separated jet over a width Δl of 180 km; other parameters remain the same as in the adiabatic case. This gives an interfacial flux/unit length along the outcrop of $U_c = 1.95$ m^2 s^{-1}, i.e., 1.95 Sv/1000 km, which requires a compensating net source of light water, $S_s - S_n$, of 3.9 Sv. Elsewhere, the model is adiabatic.

Cooling shifts the boundary current by changing the northward transport of the light water. The boundary current again separates from the western wall when the wind stress satisfies (7):

$$\tau(y_p) = \frac{g' h_e^2}{2x_e} - \frac{f(y_p)}{x_e} T_{\mathrm{north}}(y_p),$$

but the northwards transport at the point of separation, $T_{\mathrm{north}}(y_p)$, is increased from the adiabatic value in (20) by having to supply the light water that is cooled along the outcrop south of the ZWL:

$$T_{\mathrm{north}}(y_p) = T_{\mathrm{adiabatic}} + \int_{y_p}^{y_0} U_c ds. \quad (21)$$

Therefore, the boundary current has to leave the coast farther south, at a latitude where there is a smaller wind stress than in the adiabatic case.

With cooling, the separation point is shifted 320 km farther south than in the adiabatic case (Fig. 5). The cooling strengthens the western boundary current transport, $T_w(y_p)$, at the separation point by shifting the separation point farther south and by requiring it to carry this additional light water transport, $S_s - S_n$: the transport $T_w(y_p)$ is increased to 53 Sv whereas in the adiabatic case it is 39 Sv.

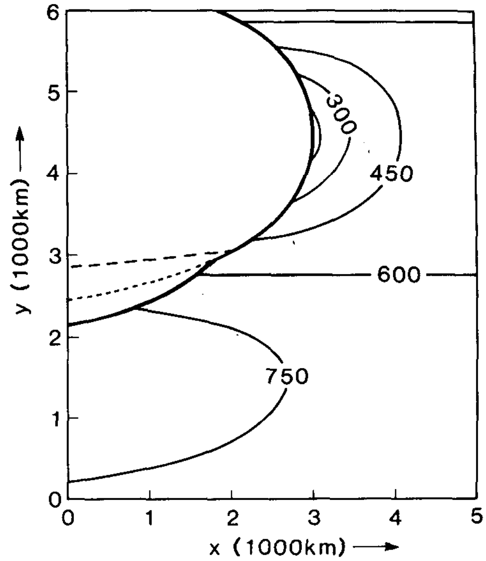

FIG. 5. The path of the separated boundary current $X_p(y)$ predicted for realistic surface cooling (thick full line) is shown together with the thickness field of the light water (full line) in a subtropical/subpolar double gyre. The paths of the separated current for the adiabatic case (short dashed line) and an artificial surface heating (long dashed line) are also included.

The separated boundary current swings eastwards in order to satisfy the transport balance for the light water [combining (8) and (20)]:

$$\frac{g'}{2f} h_c^2 - (x_e - X_p)\frac{\tau}{f} = T_{\mathrm{north}}(y) = T_{\mathrm{adiabatic}} + \int_y^{y_0} U_c ds. \quad (22)$$

Comparing this with the adiabatic expression (8) gives

$$X_p(y)_{\mathrm{diabatic}} - X_p(y)_{\mathrm{adiabatic}} = \frac{f}{\tau}\int_y^{y_0} U_c ds. \quad (23)$$

South of the zero wind latitude, the northward transport, $T_{\mathrm{north}}(y)$, is increased by cooling in order to supply light water for conversion to abyssal water. Hence, the outcrop line shifts eastwards of its adiabatic position so as to decrease the southwards Ekman flux. Here, the outcrop line is pushed by up to 1000 km farther east of the adiabatic line (Fig. 5). North of the ZWL, cooling would reduce the northward transport, but the outcrop line would still be pushed eastward as the Ekman flux is northwards here.

d. *Warming the Gulf Stream*

An artificial warming of 150 W m^{-2} is now applied along the first 2000 km of the separated jet over a width of 180 km. This implies that the interfacial flux/unit length out of the light water $U_c = -1.45$ m^2 s^{-1} and the compensating net source of light water, $S_s - S_n$, is -2.9 Sv.

Surface heating weakens the western boundary current, which pushes the separation point farther north and the path of the separated current farther west. In this heating case, the western boundary current transport at the separation point is reduced to 19 Sv. The separation point is shifted 410 km farther north and the outcrop line is pushed up to 1600 km farther west of the adiabatic case (Fig. 5).

There is a limit on the surface heating south of the ZWL. If this is exceeded, the implied northward transport at the point of separation, $T_{\rm north}(y_p)$, from (21) becomes too small for separation to be possible by (7). The minimum possible transport that allows separation is

$$T_{p\min} = \min_{\text{for all } y_p} [T_{\rm north}(y_p)]$$

$$= \min_{\text{for all } y_p}\left[\frac{1}{f(y_p)}\left(\frac{g'h_e^2}{2} - x_e\tau(y_p)\right)\right] \quad (24)$$

and occurs at $y = y_{p\min}$ that satisfies

$$w_e(y_{p\min})x_e = \frac{\beta g' h_e^2}{2f^2(y_{p\min})}. \quad (25)$$

This latitude $y_{p\min}$ is exactly where the geostrophic surfacing line, $X_{\rm geo}$, strikes the coast (13). Therefore, no transport would be carried in a western boundary current that separates here.

For the parameters we have chosen, the minimum possible separating transport, $T_{p\min}$, is 13 Sv; the transport across the ZWL, $T_{\rm adiabatic}$, is 17 Sv, and thus the maximum possible conversion of dense to light water along the outcrop south of the ZWL is 4 Sv. The heating rate chosen in Fig. 5 gives a dense to light water conversion of 2.9 Sv, which is well within this limit.

However, as the imposed wind stress field is strengthened, the minimum separating transport, $T_{p\min}$, decreases and thus the maximum permitted heating increases. If the wind stress becomes sufficiently strong, such that $\tau(y_p) > \frac{1}{2}g'h_e^2/x_e$ so $T_{p\min} < 0$, then it is even possible to have a southwards separating transport given a large enough heating, although this would necessitate an equatorial sink of light water.

Pedlosky has investigated a special case in which surface heating exactly balances the Ekman flow across the outcrop but with no mixed-layer flow $\mathbf{v}_m$ across the outcrop; thus (17) reduces to

$$U_c = \left(-\mathbf{k} \times \frac{\boldsymbol{\tau}}{f}\right) \cdot \mathbf{n} = -\frac{\tau}{f}\frac{dX_p}{ds}. \quad (26)$$

Whatever the form of the heating, the outcrop line, $X_p(y)$, always satisfies the differentiated version of the transport balance (22):

$$-\frac{\beta g' h_e^2}{2f^2} + \frac{\tau}{f}\frac{dX_p}{dy} + w_e(x_e - X_p) = -U_c\frac{ds}{dy}. \quad (27)$$

In the case treated by Pedlosky, this reduces to a balance between Ekman suction and the convergence of the geostrophic flow:

$$w_e(x_e - X_p) = \frac{\beta g' h_e^2}{2f^2}. \quad (28)$$

The separation line now follows the geostrophic surfacing line, $X_{\rm geo}$, and the separating transport, $T_{\rm north}(y_p)$, is the minimum possible, $T_{p\min}$. Therefore, the Pedlosky model corresponds to a special limit of the maximum possible heating in which the Parsons' boundary current disappears and the outcrop line is simply due to the Ekman suction in subpolar gyre.

4. Conclusions

Parsons' adiabatic model elegantly describes the transport balances that cause the Gulf Stream to separate. The inclusion of a subpolar gyre makes separation possible for reasonable wind stresses and thermocline depths. The resulting circulation pattern shows a striking asymmetry between the subtropical and subpolar gyres.

Realistic cooling, consistent with the heat loss suffered by the Gulf Stream, is imposed along the outcrop line of the separated boundary current in an extension of the Parsons' model. This results in a strengthening of the separating boundary current by approximately 15 Sv and a pushing of the separation path southeastward by up to 3000 km.

In contrast, an artifical heating along the outcrop line weakens the separating boundary current and shifts its path northwestward. As the heating is increased, a limit is eventually reached in which the separating boundary current disappears and the outcrop line coincides with the surfacing line due to Ekman suction.

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