Comments on "The Response of Intense Ocean Current Systems Entering Regions of Strong Cooling"

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ABSTRACT

In this brief note it is demonstrated that the question of what is the mechanism(s) responsible for the southern migration of the Gulf Stream front during winter—is still open.

1. Introduction

In an interesting paper, Adamec and Elsberry (1985b, hereafter referred to as AE) present a series of numerical simulations which examine the response of an oceanic jet to both atmospheric cooling and wind forcing. Their main aim was to determine whether the southward shift of the Gulf Stream front, which occurs in late winter (Fuglister, 1963; Worthington, 1976), is a result of atmospheric cooling or a result of wind forcing. The relationship between the Gulf Stream and winter cooling was previously examined by Worthington (e.g., see Worthington, 1972), Csanady (1982) and Nof (1983, hereafter referred to as N) whereas the influence of increased wind was first examined by Iselin (1940; see also Monin et al., 1977, p. 185) and recently reconsidered by Adamec and Elsberry (1985a). The basic ideas behind the processes in question are as follows.

(i) *Atmospheric Cooling:* The possible association of atmospheric cooling with the strength and position of the Gulf Stream was first suggested qualitatively by Worthington (e.g., see Worthington, 1972). His idea was that, with the aid of convection, winter cooling causes deepening of the thermocline south of the Stream and this causes an intensification of the Stream. Further developments of this idea were made by Csanady (1982) who showed analytically that a large cross-stream cooling gradient does indeed cause an intensification of the Stream which, in turn, causes a steeper interface and a southward displacement of the front. He found that due to this mechanism the Gulf Stream front may be displaced by ~35 km to the south during winter.

Nof (1983) being, at the time, unaware of Csanady's efforts (which had not appeared in press yet), has independently examined the effect of a different kind of cooling. Instead of focusing on the changes that take place due to large cross-stream cooling gradients, considered by Worthington and Csanady, he has looked at the effect of *long-stream* cooling gradients. Specifically, using a two-layer nonlinear analytical model, he has shown that an imposed along-stream atmospheric cooling forces cross-stream velocities which are in "thermal wind" balance. With the aid of a uniformly valid perturbation scheme, N has also showed that although cross-stream velocities are generated, the *main speed and transport remain unaltered.* As a result, the front migrates toward the south while the interface pivots around the curve representing the intersection of the lower interface with the level of the undisturbed depth. He estimated that, during winter, this produces a southern migration of ~90 km of the Gulf Stream front.

(ii) *Increased wind stress:* Iselin (1940) has introduced the concept that an increase of the wind over the North Atlantic circulation system causes an intensification of the currents. He argues that, because the system is roughly in geostrophic balance, such an intensification causes an increase in the thermocline depth (in center of the gyre) and a shrinkage of the gyre outer edge. This implies that during winter (when the wind is strong) the position of the Stream will be farther to the south than it is during summer.

With the aid of a ten-level primitive equation numerical model, AE have shown that an increase in the eastward wind stress causes a shift of the Stream toward the south—as expected from simple Ekman layer theory and the fact that the Stream intensifies. In addition, AE found that atmospheric cooling produces southward motions which are in thermal wind balance as predicted by N.

In this note we shall mainly be concerned with a comparison between the AE numerical study and N's analytical model. There are no fundamental differences between the main cooling processes in the two models in question although there are some technical differences as discussed by AE. Despite the fact that, dynamically, both models (i.e., the AE and the N model) produce the same process, AE argue that cooling alone cannot produce the observed southward shift of the Gulf Stream front (~100 km) as suggested by N. Their main argument is that N used cooling rates which are too large and that when "correct" values are chosen the southward shift is about one-fifth of the observed values. In what follows it will be shown that, although the values used by N are indeed high relative to many parts of the ocean, they are in agreement with heat loss measurements made in the North Atlantic and with the values used by Csanady (1982). Furthermore, it will be shown that the values used by AE are *unusually low* for the Gulf Stream in the late winter.

2. Discussion

Before presenting the observations in question, it is appropriate to point out that, in the late seventies and early eighties, there was no consensus on the appropriate cooling rate for the Gulf Stream region. Even today there is no agreement on the various constants that should be used in the bulk formulae for sensible and latent heat flux (e.g., see Woods 1984). For this reason, N has taken the observed density gradient (Fig. 1) instead of an estimated heat loss. The disadvantage of this method is that, in addition to pure atmospheric cooling which increases the density by both reducing the temperature and increasing the salinity via evaporation, other processes may also be represented in the observed density increase. For example, cooling by entrainment and heat exchange with the Slope Water could, in principle, also produce a long-stream density gradient. Unfortunately, it is difficult to estimate the heat exchange with the neighboring waters. We shall see shortly, however, that since the effect of the Slope Water can only be that of cooling and not of heating, we shall be able to make our case without estimating this lateral heat loss. Note that for Gulf Stream rings, Joyce and Stalcup (1985) have found that the lateral heat loss was comparable to the surface heat loss.

As far as the effects of evaporation are concerned, Csanady (1982) estimates that the effect of salt residues (due to evaporation) can be accounted for by increasing the thermal expansion coefficient by 25%. This appears to be consistent with the high values for the latent heat flux in the region. An alternative way to consider the effect of salt residues in computations which use direct cooling instead of density gradients, is simply to increase the forcing (i.e., the surface cooling) by 25%.

With the aid of this information, we shall now return to the main point of our note, i.e., demonstrating that the Gulf Stream heat losses to the atmosphere have important effects on the current dynamics. As can be seen from Fig. 1, the density of the Gulf Stream increases downstream; the increase is much larger during winter (roughly 1×10^{-3} per 1200 km) than it is during the summer ($\sim 0.5 \times 10^{-3}$ per 1200 km). Note that

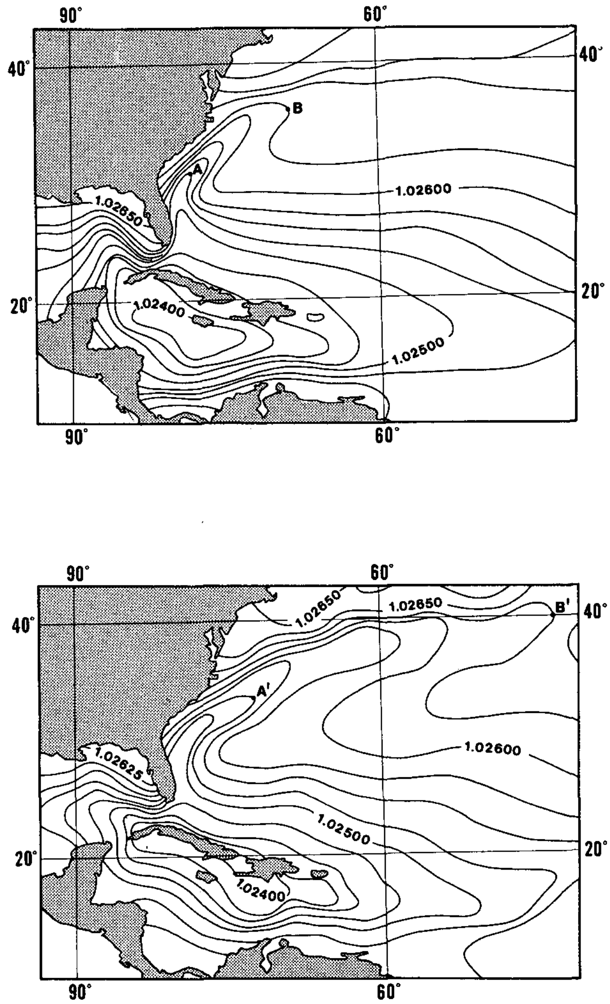

FIG. 1. The density at 100 m depth during February (upper panel) and August (lower panel). The maps have been adapted from Gorshkov (1978). Points A and B (and A′ and B′) represent the approximate intersection of the Gulf Stream axis with the curves corresponding to a density of 1.02525 and 1.02625. Note that the Stream's density increases downstream. In addition, note that during the winter (upper panel) the density gradient is at least twice as large as the one during the summer (lower panel).

these long-stream density gradients are consistent with the frequently used sea surface temperature maps which illustrate that the Stream loses a few degrees as it flows toward the northeast. When the above density gradients are combined with appropriate values for the current's uniform velocity (say, 0.4 m s^{-1}) and average depth (~400 m), and the dynamical relationships used by the N model [i.e., equations (4.9) and (2.10) in N], one finds that cooling rates of 1000–2000 W m^2 are acting on the Stream during winter. In other words, in order to produce the long-stream density gradients (given by Gorshkov) and obtain the N's cross-stream velocities (~3 cm s^{-1}), it is necessary to apply a cooling rate of 1000–2000 W m^{-2}.

This can be easily demonstrated by considering the following dimensional variables predicted by N,

$$v = -\frac{g'}{f}\frac{\partial\delta}{\partial x}z \qquad (1)$$

$$\frac{\partial\delta}{\partial x}U_0(H+\xi) = \frac{Q}{C_p(T_0 - T_l)}. \qquad (2)$$

Here, v is the predicted "thermal wind" velocities across the Stream, g' the "reduced gravity" based on the uncooled upstream state [i.e., $g(\Delta\rho_0/\rho)$, where $\Delta\rho_0$ is the original upstream density difference between the layers], f the Coriolis parameter, and δ corresponds to the local temperature [i.e., $\delta = (T_0 - T)/(T_0 - T_l)$, where T is the upper layer temperature at some point x, T_0 and T_l are the uncooled upper and lower layer temperatures]. In (2) U_0 is the uniform upstream flow, H the upper layer undisturbed depth, ξ the interface displacement, Q the heat loss, and C_p is the water heat capacity. The x coordinate is directed downstream and z is measured positively upward. Elimination of $\partial\delta/\partial x$ between (1) and (2) and taking $g' = g\alpha(T_0 - T_l)$ (where α is the coefficient of thermal expansion) gives

$$Q = -C_p f U_0 v(1 + \xi/H)/g\alpha\,. \qquad (3)$$

For a thermal expansion coefficient of 2.5×10^{-4} K^{-1}, water heat capacity of 4.18×10^6 W s/m^3 °K, mean speed of 0.4 m s^{-1}, predicted cross-stream speed of 0.03 m s^{-1}, and a Coriolis parameter of 10^{-4} sec^{-1}, one finds a cooling rate of about 2000 W m^{-2} *for the undisturbed depth of the Stream* (~400 m). At 200 m depth (i.e., the approximate core of the current where $\xi/H = -0.5$) the required cooling is smaller, about 1000 W m^{-2}. These cooling rates are required only during the coldest weeks of the winter when the southern migration is maximum (~90 km).

In contrast to the estimates mentioned above, AE have chosen to use an unusually low value for the average winter cooling (160 W m^{-2}). Their value is an order of magnitude smaller than the amount required to produce the results given by N so that it is not surprising that their north–south displacements are also an order of magnitude smaller. It is not clear why such a low value was chosen by AE. Even the average winter cooling estimate of Bunker (1976) [which is not limited to the Stream's central core] is about *four* times larger (560 W m^{-2}). Also, to account for the salt residue due to evaporation, the chosen value should be increased; Csanady (1982) estimates that an increase of 25% is appropriate. This was neglected by AE.

In addition, note that AE have used a jet with a maximum speed which is somewhat high—close to 2 m s^{-1}. Since the density gradients which result from the cooling are directly proportional to the jet's flux, a reduction in the jet speed would probably increase the cross-stream speed and the north–south displacements. For example, a reduction of 50% in the jet speed would probably double the long stream density gradients.

A cooling rate of 1000–2000 W m^{-2} (which is required to produce the deflections discussed by N) appears, at first, to be very large because it is considerably larger than the cooling rates observed elsewhere (e.g., see Gill, 1982; Huh et al., 1984; Kondo, 1976). While it is certainly true that such values are high, Fig. 2 shows that even higher rates are possible in the region in question. These estimates, which have recently been made by Schmitt and Olson (1985), are based on measurements made in warm-core Gulf Stream rings. They showed that, during winter storms, the heat loss to the atmosphere can peak to values of about 2000 W m^{-2}

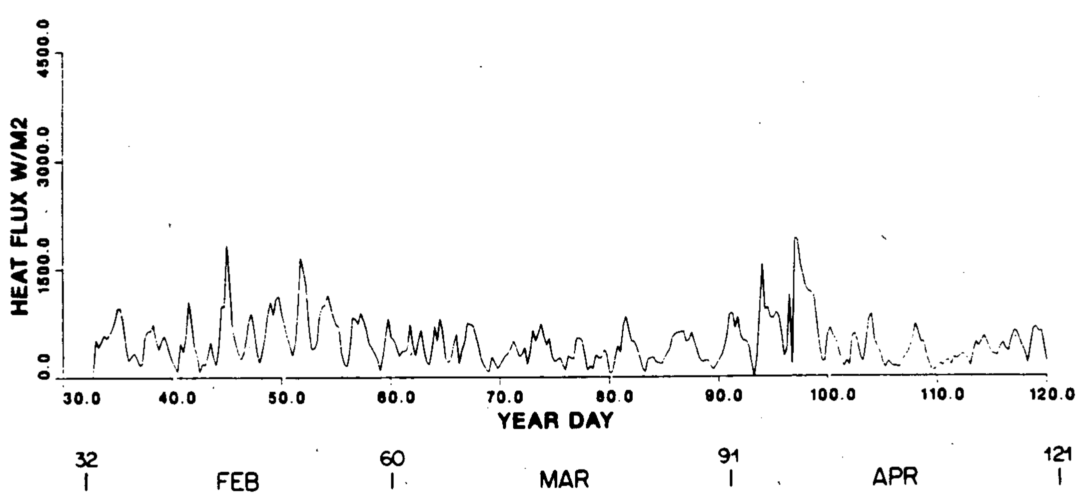

FIG. 2. The total heat loss (to the atmosphere) of a Gulf Stream warm-core ring (reproduced from Schmitt and Olson, 1985). Note that, occasionally, the heat loss peaks to 2000 W m^{-2} and that the average heat loss during the coldest months of the winter is roughly 700–800 W m^{-2}. This is consistent with Csanady (1982) estimate of 800 W m^{-2} for the averaged winter cooling for the Gulf Stream central core.

and that the *average heat loss* during the coldest months in the winter is roughly 700–800 W m^{-2}. In his study of the Gulf Stream intensification in winter, Csanady (1982) used a very similar value (800 W m^{-2}) for the *averaged* cooling rate in the central core of the Stream. It is recalled now that, in order to incorporate the effect of salt residue due to evaporation, it is necessary to increase this amount by ~25% (Csanady 1982). This brings the total surface cooling rate which one should use (i.e., sensible, latent, and the "equivalent salt" cooling) during the winter to about 1000 W m^{-2}. The latter is about *six times* larger than the average winter cooling rate used by AE.

In summary, I am not convinced by AE arguments regarding the smallness of the cooling-induced drift because they have used surface cooling rates which are $^1/_5$–$^1/_6$ of both the rates measured by Schmitt and Olson (1985) and the estimate of Csanady (1982). Their value is also much smaller than Bunker (1986) estimates and they ignore the effect of salt residues due to evaporation. It is expected that, if AE were to force their model with the appropriate estimates, then displacements and cross-stream velocities similar to those discussed by N would have been obtained. This should be the case unless there are some important differences between the N and AE models. The only clear dynamical differences between the two models is the way that the cross-stream cooling is imposed. It appears, however, that this cannot account for the lack of southern migration in the AE study because, according to Csanady's model, such cooling is also causing a southward migration. While I do not believe that AE arguments are false, the preceding considerations imply that the cooling-induced drift of the Gulf Stream front is of the same order as the observed shift. Note, however, that, as stated by N, this does not imply that the wind's effects are unimportant nor does it suggest that long-stream cooling is the only process responsible for the southern migration of the Stream.

Overall, it can be said that there are three possible candidates for the southern migration of the Gulf Stream front during winter—the wind (i.e., the AE model), the cross-stream heat loss gradient (i.e., the Worthington-Csanady mechanism), and the long-Stream cooling (i.e., the process suggested by N). It is quite likely that all of these are important contributors to the seasonal displacement of the Stream.

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REFERENCES

Adamec, D., and R. L. Elsberry, 1985a: Response of an intense oceanic current system to cross-stream cooling events. *J. Phys. Oceanogr.*, **15**, 273–287.

——, and ——, 1985b: The response of intense oceanic current systems entering regions of strong cooling. *J. Phys. Oceanogr.*, **15**, 1284–1295.

Bunker, A. F., 1976: Computations of surface energy flux and annual air–sea interaction cycles of the North Atlantic Ocean. *Mon. Wea. Rev.*, **104**, 1122–1140.

Csanady, G. T., 1982: The thermocline driving mechanism of oceanic jet streams. *J. Mar. Res.*, **40**(Suppl.), 113–142.

Fuglister, F. C., 1963: Gulf Stream '60. *Progress in Oceanography*, Vol. 1, Pergamon, 263–273.

Gill, A. E., 1982: *Atmosphere Ocean Dynamics.* Academic Press, 662 pp.

Gorshkov, S. G., 1978: *World Ocean Atlas, Vol. 2, Atlantic and Indian Oceans.* Pergamon Press, 306 pp.

Huh, O. K., L. J. Rouse Jr. and N. D. Walker, 1984: Cold-air outbreaks over the northwest Florida continental shelf: Heat flux processes and hydrographic changes. *J. Geophys. Res.*, **89**, 717–726.

Iselin, C. O'D., 1940: Preliminary report on long-period variations in the transport of the Gulf Stream system. *Pap. Phys. Oceanogr. Meteor.*, **8**, 1–40.

Joyce, T. M., and M. C. Stalcup, 1985: Wintertime convection in a Gulf Stream warm core ring. *J. Phys. Oceanogr.*, **15**, 1032–1042.

Kondo, J., 1976: Heat balance of the East China Sea during the air mass transformation experiment. *J. Meteor. Soc. Japan*, **54**, 382–398.

Monin, A. S., V. M. Kamenkovich and V. G. Kort, 1977: *Variability of the Oceans.* Wiley & Sons, 241 pp.

Nof, D., 1983: On the response of ocean currents to atmospheric cooling. *Tellus*, **35A**, 60–72.

Schmitt, R. W., and D. B. Olson, 1985: Wintertime convection in warm-core rings: Thermocline ventilation and the formation of mesoscale lenses. *J. Geophys. Res.*, **90**, 8823–8837.

Woods, J. D., 1984: The upper ocean and air-sea interaction in global climate. *The Global Climate*, J. T. Houghton, Ed., Cambridge University Press, 233 pp.

Worthington, V. L., 1972: Anticyclogenesis in the oceans as a result of outbreaks of continental polar air. *Studies in Physical Oceanography, Vol. 1*, A. L. Gordon, Ed., Gordon and Breach, 194 pp.

——, 1976: *On the North Atlantic circulation.* Johns Hopkins University Press, 110 pp.