Preliminary Estimates of the Role of Mesosynoptic Scale Sea Surface Temperature Features in Fostering Explosive Midlatitude Cyclogenesis

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1. Introduction

An intriguing problem for current meteorological research is an explanation of the tendency for explosive atmospheric cyclogenesis (with central pressure falls $\geqslant 1$ mb h^{-1} for 24 h) to occur preferentially in marine environments in regions of large sea surface temperature (SST) gradients ($\sim$10°C per 180 km) (Sanders and Gyakum, 1980; hereafter, SG). These SST gradients are found near the mean winter positions of the Gulf Stream and Kuroshio, and their extension currents. In addition to the possibility that explosive cyclogenesis may be a necessary component in realistic model simulations of the atmospheric general circulation (SG), it is an important problem in operational forecasting. These potentially destructive storms are typically underforecast by operational forecast models (SG; Anthes *et al.*, 1983). Therefore, further study of explosive extratropical cyclogenesis is needed.

In addition, it has been noted that two periods of especially intense cyclone-scale baroclinic development over the North Atlantic were followed by a breakdown of the zonal flow, and by the onset of typical blocking patterns over Europe and the North Atlantic (SG; Hansen and Chen, 1982). If this last association does not turn out to be coincidental, increased understanding of the "bomb" (in SG's parlance) phenomenon may also contribute to the eventual improved prediction of the transition between cold season high- and low-index regimes on the planetary scale, in certain cases.

The answer to the important question of why explosive cyclogenesis occurs primarily over oceans (SG; Mullen, 1982, 1983) must lie in the differences between the marine and continental environments. Factors that distinguish the oceanic environment from the continental include: 1) the presence of potentially large surface fluxes of sensible and latent heat over the oceans; 2) the resultant low static stability and high water-vapor mixing ratios at low levels due to those fluxes; and 3) the relatively low surface drag over water as compared to land. The smaller surface drag over the ocean would imply reduced frictional dissipation in the atmospheric boundary layer.

This note presents some order-of-magnitude estimates of potential vorticity generation rates in the lowest hundred meters of atmosphere over schematic SST features typical of the Atlantic region of maximum frequency of bomb occurrence (SG, Figure 1). These estimates are certainly not intended to suggest that the lower boundary conditions are the only factor influencing bomb inception and development. Mid- and upper-tropospheric flow features and processes identified, e.g., by SG, Gyakum (1983a), and Mullen (1983), are also likely to be important. Furthermore, the subsequent motion of flow features developing in the lower troposphere across the relatively steady and locally fixed SST patterns cannot be taken into account in the crude estimates presented here. Determination of the effects of such motion would, of course, require the full four-dimensional complexity of numerical models, combining, perhaps, the mesoscale horizontal resolution of Anthes *et al.* (1983) and the detailed vertical resolution of the coupled air–ocean surface boundary layers of Brown *et al.* (1982). Nevertheless, the estimates will provide some guidance as to surface conditions and SST patterns which would tend to favor bomb occurrence.

2. Estimates of possible atmospheric lower boundary conditions on potential vorticity implied by schematic SST patterns

We use the potential vorticity equation of Staley (1960), in his form (17), in which no restricting adiabatic or quasi-geostrophic assumptions have been applied, *viz.*

$$\frac{dP}{dt} = \frac{10^3}{\rho g}\left(\eta_z \frac{\partial H}{\partial z} + \eta_x \frac{\partial H}{\partial x} + \eta_y \frac{\partial H}{\partial y}\right) - \frac{\partial \theta}{\partial p} R_N, \quad (1)$$

where

η = absolute vorticity vector with components,
$\eta_z = (\partial v/\partial x) - (\partial u/\partial y) + f,$
$\eta_y = (\partial u/\partial z) - (\partial w/\partial x),$
$\eta_x = (\partial w/\partial y) - (\partial v/\partial z).$

The rate of generation of potential vorticity is, obviously, a good measure of the intensity of cyclogenesis (see, e.g., Gyakum 1983b). The other symbols are:

H $(d\theta/dt)$, i.e., the diabatic heating rate (K s^{-1})
ρ density of air (gm cm^{-3})
θ potential temperature (K)
g specific gravitational force (dynes gm^{-1})
t time (s)
P potential vorticity $\equiv -(\partial\theta/\partial p)\eta_z$ (K mb^{-1} s^{-1})
f Coriolis parameter (s^{-1})
p pressure (mb).

The factor 10^3 gives the units of the first (generation) term in K mb^{-1} s^{-2} for ready comparison with the diagnostic results of Gyakum (1983b).

The term with R_N as a factor represents the frictional effects on P in Staley's equation. We shall, in these crude order of magnitude estimates, be concerned only with the other terms. However, at the end of this note, we will outline some interesting qualitative aspects of this term.

We shall only consider the lowest 100 m of the atmosphere, in order to obtain estimates of the lower boundary conditions that might be encountered by numerical models of the atmosphere used to study and predict cyclogenesis.

We will first obtain order of magnitude estimates for the major synoptic scale SST feature of SG (their Figure 12a). As may be seen in their figure, the SST gradient is of the order of 10 K (180 N $mi)^{-1}$ across both the eastward flowing Gulf Stream branch and northward flowing North Atlantic current branch of, e.g., Fig. 2 of Menard (1983). We will estimate the upward flux of sensible heat into a cold surface boundary layer coming into adjustment on the warm side of the SST boundary by the conventional formula,

$$F_0 = C_D|V|(\theta_0 - \theta_A), \quad (2)$$

where

C_D $\approx 1.5 \times 10^{-3}$
$|V|$ ≈ 1 m s^{-1}
$(\theta_0 - \theta_A)$ ≈ 10 K
θ_0 SST
θ_A the surface temperature of the unadjusted air layer.

The vertical flux at the top of this layer (say at 100 m) F_A, is assumed to be negligible,

$$|F_A| \ll |F_0|,$$

as is the flux on the opposite side of the SST boundary.

The diabatic heating rate is estimated, therefore, as

$$H \sim \frac{\partial F}{\partial z} \sim 13 \text{ K day}^{-1}$$

on the unadjusted side of the boundary.

We note here that the same heating rate would be estimated for a surface boundary layer in quasi-adjustment, in which the heat flux at the top (F_A)

$$F_A \approx 0.9F_0,$$

in which the wind speed is 10 m s^{-1}, and for a full range of physically reasonable conditions between these two speeds. We also note that it is consistent in magnitude with that calculated by the numerical model of Anthes *et al.* (1983, Fig. 13) over a region of strong cyclogenesis, and with the tropospheric heating rate estimated by Gyakum (1983b, Table 1) at a time immediately preceding bomb development.

In the following discussion, estimates are made for each of the current branches. It is to be understood that the two cases are being treated separately. It is not necessary, for our purposes, that the conditions specified (separately for each branch) occur at the same time.

Therefore, based on SG's (Figure 12a) maximum SST gradient,

$$\frac{\partial H}{\partial y} \sim -4.5 \times 10^{-12} \text{ K s}^{-1} \text{ cm}^{-1}, \quad \left|\frac{\partial H}{\partial x}\right| < \left|\frac{\partial H}{\partial y}\right|, \quad (3a)$$

across the Gulf Stream branch, and similarly

$$\frac{\partial H}{\partial x} \sim 4.5 \times 10^{-12} \text{ K s}^{-1} \text{ cm}^{-1}, \quad \left|\frac{\partial H}{\partial y}\right| < \left|\frac{\partial H}{\partial x}\right|, \quad (3b)$$

across the North Atlantic branch.

A horizontally uniform surface wind on the synoptic scale with an easterly component of 1 m s^{-1} at 100 m elevation would yield, flowing over a westerly surface current of 1.5 m s^{-1},

$$\eta_y \sim -2.5 \times 10^{-2} \text{ s}^{-1}, \quad (4a)$$

along the Gulf Stream branch (assuming $|\partial w/\partial x| < |\partial u/\partial z|$). Similarly, a northerly surface wind would yield

$$\eta_x \sim 2.5 \times 10^{-2} \text{ s}^{-1}, \quad (4b)$$

along the North Atlantic Current branch.

Staley's equation, then, provides an estimate of the rate of potential vorticity generation at the lower atmospheric boundary of either

$$\frac{10^3}{\rho g}\eta_x\frac{\partial H}{\partial x} \quad \text{(along the North Atlantic branch)},$$

or

$$\frac{10^3}{\rho g}\eta_y\frac{\partial H}{\partial y} \quad \text{(along the Gulf Stream branch)}.$$

In both cases, the relevant term has order of magnitude

$$\sim 1.125 \times 10^{-10} \text{ K mb}^{-1} \text{ sec}^{-2},$$

or

$$\sim(10^{-5}\ \mathrm{K\ mb^{-1}\ sec^{-1}})\ \mathrm{day}^{-1},$$

under these surface layer wind conditions.

We note first, that such setups are consistent with the ageostrophic flow associated with surface isobaric patterns in which low pressure lies either to the west (which favors Gulf Stream branch cyclogenesis), or to the south (which favors North Atlantic branch cyclogenesis) of the belt of maximum SST (SG's Fig. 12a); and second, that the estimated one-day increase in potential vorticity is of the order of magnitude of the value estimated by Gyakum (1983b, Fig. 6) at a time preceding the "*QEII* bomb" inception.

Next, we obtain order of magnitude estimates for the types of mesoscale SST features that might be superimposed on the broad scale SST pattern. We refer to Richardson (1980, Fig. 12) to illustrate some general properties of these features, designated as "rings" and "meanders."

We base our estimates on the following ring or meander properties. The feature's horizontal scale is of order 100 km for both (cold and warm) rings and meanders. The central SST is of order 10 K warmer (or colder) than that of the surrounding water (in the first weeks after ring formation). For this estimate we shall assume that the surface air layer is being rapidly warmed (or cooled) by the central SST while the air above 100 m has not yet been affected. The cold (warm) features have cyclonic (anticyclonic) relative vorticity consistent with tangential current speed of 1.5 m sec^{-1} at the feature's periphery.

Thus, for a cold (cyclonic) feature, the *absolute* vorticity in Eq. (1) is

$$\eta_z \sim 1.5 \times 10^{-4}\ \mathrm{s}^{-1}, \tag{5a}$$

$$\frac{\partial H}{\partial z} \sim 10^{-8}\ \mathrm{K\ cm^{-1}\ s^{-1}}, \tag{5b}$$

$$\frac{10^3}{\rho g}\eta_z\frac{\partial H}{\partial z} \sim 1.5 \times 10^{-9}\ \mathrm{K\ mb^{-1}\ s^{-2}}, \tag{5c}$$

or

$$\sim 12 \times 10^{-5}\quad (\mathrm{K\ mb^{-1}\ s^{-1}})\ \mathrm{day}^{-1}.$$

It is the interaction of the *planetary* vorticity f with the rapid, persistent, mesoscale surface layer cooling that, in this case, makes the dominant contribution to the intense generation of cyclonic potential vorticity. This interaction term could also be large for a warm *atmospheric* intrusion over the cold water side of the straight, synoptic scale oceanic current, but only during a transient period during which the air mass modification is confined to a shallow surface layer.

We note, first, that this contribution to the rate of generation is significantly larger than that estimated for the $\eta_x(\partial H/\partial x)$, $\eta_y(\partial H/\partial y)$ terms. Second, we note that it is of the order of magnitude of the one-day change in potential vorticity estimated by Gyakum (1983b) during the explosive growth phase of the *QEII* bomb (compare his Figs. 6 and 7).

Some qualitative aspects of the frictional term may be elucidated if we expand the term following Staley [1960, Eq. (24)],

$$-\frac{\partial\theta}{\partial p}R_N \approx \frac{10^3}{g}\frac{\partial\theta}{\partial z}\left[\frac{\partial}{\partial x}\left(\frac{\partial}{\partial z}K\frac{\partial v}{\partial z}\right) - \frac{\partial}{\partial y}\left(\frac{\partial}{\partial z}K\frac{\partial u}{\partial z}\right)\right], \tag{6}$$

where K is a positive exchange coefficient. In the first part of this section, we were concerned primarily with the SST fields in the region; now we will consider further some schematic properties of the surface current fields. The Gulf Stream (North Atlantic) system may be described as a westerly (southerly) jet flowing on the warm side of the sharpest SST gradients. Thus the surface current field exhibits cyclonic vorticity on the left side of the current and anticyclonic vorticity on the right side. An opposing, horizontally uniform, surface layer atmospheric wind over this surface current system will, for example, result in a positive sign for the bracketed factor in the equation above on the left side of the current, and a negative sign on the right side. If these flow fields are accompanied by stable surface layer stratification [i.e., $(\partial\theta/\partial z) > 0$; warm surface air overlying cold surface water], on the cyclonic side of the current system, or unstable surface layer stratification on the anticyclonic side, this frictional term will make a *positive* contribution to the surface layer potential vorticity generation. Gyakum (1983b) has shown that "cumulus" friction can make such a positive contribution throughout the deeper lower troposphere under certain circumstances. We believe that the argument outlined above demonstrates the possibility of a similar effect due to surface layer friction.

We note that, in general, the occurrence of unstable surface layer stratification $[(\partial\theta/\partial z) < 0]$ in cold air overlying warm water probably is a shorter-lived condition than the reverse. This suggests a possible reason for the pattern of bomb frequency of occurrence evident in SG's Fig. 1, in which the regions of most frequent occurrence tend to be on the cyclonic side of the major current system.

A qualitative analysis of the frictional effects in the atmospheric surface layer overlying mesoscale meanders and rings yields similar results. We need merely to note that cyclonic surface current features tend to be associated with warm air overlying cold water and anticyclonic features with cold air overlying warm water.

3. Summary and conclusions

Rates of generation of cyclonic potential vorticity in the atmospheric surface boundary layer of the order of magnitude that was diagnosed prior to a bomb inception (Gyakum, 1983b) are found to be possible over the eastward flowing Gulf Stream with an easterly component of surface layer wind with magnitude as small as 1 m s^{-1}. They are also possible over the northward flowing North Atlantic Current with a northerly component of surface layer wind of similar magnitude.

In both cases, these rates occur with occasionally observed values of surface layer heating, and reasonable ranges of variation of this heating in the horizontal. The analysis outlines a mechanism (which should be investigated in more detail) which may serve to explain the centers of maximum bomb frequency in SG Fig. 1.

Rates of generation of cyclonic potential vorticity an order of magnitude larger than those estimated above are found to be possible over cyclonic (cold center) mesoscale meanders or eddies in these ocean currents. The dominant contribution to these more intense rates results from the interaction of the planetary vorticity and the strong surface layer cooling that can occur when there is persistent warm air flow over such cold mesoscale surface features. Similarly intense rates of generation of anticyclonic potential vorticity are possible over anticyclonic (warm center) meanders or eddies.

In all four cases listed, the frictional terms of Eq. (1) make a positive contribution to the growth of cyclonic potential vorticity.

It was pointed out in Section 1 that previous studies (see, e.g., SG) indicate that it is unlikely that such surface layer and SST features are sufficient to induce bomb inception. The question raised by this analysis is whether they are necessary.

The midtropospheric features identified in SG's synoptic climatology are conducive to cloudiness over the regions in question. This precludes easy documentation of mesosynoptic scale SST fields, which require favorable conditions for satellite-based IR sensing of the sea surface. The more difficult task of associating SST fields inferred from archived conventional data with bomb inception remains to be carried out.

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