Observational Evidence for the Influence of Surface Heat Fluxes on Rapid Maritime Cyclogenesis

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ABSTRACT

We present an observational study of the possible effects of sea surface fluxes of latent and sensible heat on rapidly deepening cyclones over the western Atlantic Ocean. Based on the recognition that conventional operational models (specifically the LFM), until very recently, routinely underpredicted the intensity of such cyclogenesis events, we form the hypothesis that this failure may be systematically related to the amount of warming that can occur from the ocean, through sensible heating and the condensation of evaporated sea water aloft. We test this idea for a sample of 29 explosive cases occurring between 1981 and 1985. Results indicate a strong correlation between the underestimate of 24 hour central pressure fall by the model and the increase in 1000–500 mb thickness that would occur if that layer were to become neutrally stable to surface parcels saturated at sea surface temperature. We find the correlation particularly robust for the more rapid deepeners of the sample.

In the second portion of this study, the climatology of the difference between this "potential saturation thickness" and the actual 1000–500 mb thickness is calculated. From twice daily data extending over 15 years (1963–77), the mean is computed for positive values of thickness difference only. Regions of large mean values correspond quite well with regions of frequent explosive cyclogenesis.

1. Introduction

In recent years, considerable attention has been devoted to the phenomenon of explosive cyclogenesis, the extraordinarily rapid development of midlatitude surface cyclones. Typically, explosive developments are singled out from what we might term "ordinary" by employing a cutoff criterion on the rate of central pressure fall. We will use the "bomb criterion" introduced by Sanders and Gyakum (1980, hereafter SG) which considers a cyclone explosive if its maximum 24-hour central pressure fall is $\geqslant 1$ Bergeron (B), or 24 mb $\times \sin\phi/\sin 60°$, where ϕ is the latitude position of the cyclone center midway through the 24-hour period. We note that many cyclones undergo brief periods of deepening where dp/dt is greater than 1 mb h^{-1}, but are excluded from explosive status. This criterion also assures that "bombs" are cyclones of exceptionally large amplitude.

At the heart of this study is the notion that bombs are indeed themselves a phenomenon, i.e., that they differ importantly from the majority of cyclones which are believed to arise almost exclusively from baroclinic instability. It is quite likely that baroclinic processes are the most important single factor in explosive developments (Sanders 1986a, hereafter S1; Rogers and Bosart 1986) but there is some evidence that physical processes of minor consequence in typical cases become quite important in bombs. The climatology of SG shows a strong preference for bombs to occur over western ocean regions and in the cold season of the year. These are the regions with maximum sea surface temperatures at a given latitude, suggesting the possible importance of diabatic processes.

It is also true, however, that these regions have stronger mean baroclinicity than other regions of the world, are typically found slightly downstream of planetary trough axes and are regions of frequent mobile trough passage, all conditions favorable for baroclinic development. It is exceedingly difficult to separate the relative contributions of baroclinic dynamics, diabatic processes and other physical processes (e.g., friction) from routine observations. Indeed, it is likely that they are interdependent.

Better evidence for viewing explosive cyclogenesis as dynamically somewhat different from ordinary cyclogenesis resulted from the work of Roebber (1984). Examining all Northern Hemisphere extratropical cyclones for a one-year period, he simply recorded the number of times a given maximum deepening rate occurred. The number of events versus deepening rate distribution showed a statistically significant surplus of cases in the rapid deepening regime, relative to a normal distribution. This suggests that another physical mechanism may be operating in addition to baroclinic

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instability, but, of course, does not address the identity of that mechanism. Many efforts to explain the essential physics have utilized numerical models, where it is possible to include or exclude various physical processes and examine the effect on cyclogenesis. Most of these studies focussed on single cases (Anthes et al. 1983, Atlas 1987) and their results are not necessarily generalizable to the majority of explosive cyclogenesis events. The results from these case studies do suggest a strong importance of baroclinic processes but also a large growth rate enhancement subsequent to the inclusion of fluxes of moisture from the ocean surface.

The failure of a numerical model to adequately simulate a bomb event may be at least as useful as a success, however. If the failure is systematic over a large sample of cases, we may gain insight through the knowledge of model deficiencies. Until recently, the primary National Meteorological Center (NMC) forecast model was the Limited-area Fine-mesh Model (LFM II) and as discussed in Sanders (1986b, hereafter S2), it had a strong tendency to underforecast explosive maritime developments. On the other hand, Silberberg and Bosart (1982) noted that the same model was reasonably successful in continental cases of cyclogenesis, even for fairly rapid deepeners. There was a slight bias in favor of overdevelopment of cyclones over land, probably tied to orographic effects.

This behavior suggests that either some physical process peculiar to oceanic cyclogenesis may be poorly handled by the LFM or the poor quality of data over the ocean makes accurate forecasts unlikely. Recent success with the Nested Grid Model (NGM) in forecasting bombs over the Western Atlantic (Sanders 1987) indicates that the small number of observations is not detrimental there since both models are initialized with available data over land and the 12 hour global forecast (none of the previous LFM or NGM forecasts appears in their subsequent initializations, Gerrity 1977 and personal communication).

We are thus left to consider some physical process or processes peculiar to oceanic cyclogenesis. Perhaps the most plausible factor is the influence of latent and sensible heat derived from the sea surface. Emanuel (1986) in an analytical study, showed that an intense steady state tropical cyclone can be maintained against frictional dissipation solely by a moist entropy flux from the sea surface. In addition, Rotunno and Emanuel (1987) showed that this evaporation feedback can develop a finite amplitude initial vortex into a mature hurricane on a time scale consistent with observations. No conditional instability was present in the environment, which, combined with the need for a finite amplitude vortex, eliminates the possibility of conditional instability of the second kind (CISK, Charney and Eliassen 1964). We feel that a similar mechanism may be operating to enhance some instances of extratropical oceanic cyclogenesis.

There is growing evidence (Emanuel 1988) that the environments of midlatitude cyclones may be characterized by zero moist potential vorticity (q_e). Again, there seems to be no potential energy for CISK, but this condition represents a large susceptibility for strong slantwise ascent in the presence of environmental forcing. Such motions presumably must exist if the energy extracted from the sea surface is to become realized in the interior. The observation of $q_e = 0$ has many implications for forecast models as well. Models such as the LFM, with a crude, modified convective adjustment routine and coarse vertical resolution (three interior tropospheric levels and a boundary layer of fixed 50 mb depth) will have difficulty distributing evaporated water from the ocean surface aloft. The condition $q_e = 0$ implies a stable lapse rate in the vertical for a saturated atmosphere, hence the convective parameterization would not be implemented. Emanuel et al. (1987) have presented theoretical evidence that the sloping updraft under these conditions may become very narrow after a time, hence it may not be captured by the LFM due to the coarse vertical resolution and modest horizontal resolution of the model (180 km at 60°N).

Apparently, the LFM is very deficient in its upward transport of the evaporated water, resulting in a boundary layer which quickly saturates, thereby cutting off the evaporation (Gerrity, personal communication; note that NMC only allows the boundary layer to get within a few percent of saturation). There is also *no wind speed dependence* in the parameterization of surface fluxes. Thus, even if communication of these fluxes to the interior was occurring, the magnitude of the fluxes would be too small and there would be none of the important feedback that exists in the model of tropical cyclone development.

We hypothesize then that the principle reason for the poor performance of the LFM in forecasting bombs is an inadequate representation of the effects of evaporation and sensible heating from the sea surface. In this paper, we will test this hypothesis by investigating the degree to which the forecast deficiency is related to the amount of heating obtainable from the ocean surface for a number of cases. Furthermore, we will perform a climatological calculation which suggests that regions having a large susceptibility for heating from the sea surface coincide well with regions where explosive cyclogenesis occurs with appreciable frequency.

2. The LFM Experiment

In considering the possible effects of heating from the ocean surface, we will not explicitly calculate the heating from sea surface fluxes, rather, we will use as a measure of *potential* heating the maximum amount of warming that could occur from such fluxes. Specifically, the maximum warming that can result from this heating from below will leave the atmosphere saturated, with a moist adiabatic lapse rate and the surface temperature equal to the sea surface temperature (SST). If we consider the surface pressure to be 1000 mb for the moment, we can define a 1000–500 mb potential saturation thickness

$$\mathrm{PT} = -\frac{R}{g}\int_{1000\ \mathrm{mb}}^{500\ \mathrm{mb}} T_{vs} d(\ln p)$$
$$\approx 5126.5 + 27.3 \times \mathrm{SST}\ (^\circ\mathrm{C})\ [\mathrm{meters}] \quad (1)$$

where T_{vs} is the saturated virtual temperature along a moist adiabat, $R = 287\ \mathrm{J\ kg^{-1}\ K^{-1}}$ and $g = 9.8\ \mathrm{m\ s^{-2}}$. As it turns out, the linear fit to this integral of the hydrostatic equation has a correlation coefficient of 0.998. Hereafter, then, we will use the numerical equivalent for PT, thus, PT is a linear function of SST alone.

To calculate the amount of warming that is possible from the sea surface at a given location, we simply subtract the actual 1000–500 mb thickness from PT to obtain the maximum thickness change possible, denoted Δ. Hence, Δ must be >0 for warming to occur. The choice of the 1000–500 mb layer is strictly motivated by the convenient availability of 1000–500 mb thickness analyses and forecasts over the ocean.

Observations of GALE (Genesis of Atlantic Lows Experiment) cyclones suggest that the atmosphere in the vicinity of cyclones over the east coast of the United States may be characterized by zero moist potential vorticity (Emanuel 1988). To put this in another context, consider that for two dimensional flow, the quantity $M = v + fx$ is conserved following parcels, where x is the distance orthogonal to the thermal wind increasing toward warmer air, and v is the velocity component parallel to the thermal wind. As shown in Emanuel (1983) the presence of shear provides a centrifugal force in addition to the buoyant force, hence, in an environment stable to upright convection, there can still exist an instability to parcels with appropriately sloped trajectories. The amount of potential energy that exists can be assessed by lifting a parcel along an M surface as defined above and comparing its virtual temperature to that of the environment along the same surface. For a saturated atmosphere, the statement that $q_e = 0$ is equivalent to the statement that a parcel lifted along such a surface has zero buoyancy (and is stable to upright displacements). The existence of such a neutral state is not likely due to chance, but rather is probably the result of slantwise motions that have neutralized a presumably unstable state. The apparent tendency for the atmosphere to adjust in this manner suggests that we redefine our measure of potential warming using the 1000–500 mb thickness along an M surface to represent the actual thickness (note that PT is independent of this choice). This thickness difference will be denoted Δ_M and that using the conventional upright thickness will be Δ_U.

To calculate the thickness along an M surface originating from some 1000 mb point, we first need to find where this surface intersects the 500 mb level. If $dM = 0$, then for f = const, $dx = -dv/f$. Integrating this relation along the M surface and letting

$$(v_{500\ \mathrm{mb}} - v_{1000\ \mathrm{mb}})\,|_M \approx \frac{g}{f}\left.\overline{\frac{\partial(\mathrm{THK})}{\partial x}}\right|_M \quad (2)$$

we obtain for the M surface displacement

$$\delta x \approx -\frac{g}{f^2}\left.\overline{\frac{\partial(\mathrm{THK})}{\partial x}}\right|_M \quad (3)$$

where the overbar denotes the average along the M surface. As shown in Fig. 1, having moved a distance δx at 500 mb relative to our origin at 1000 mb, we

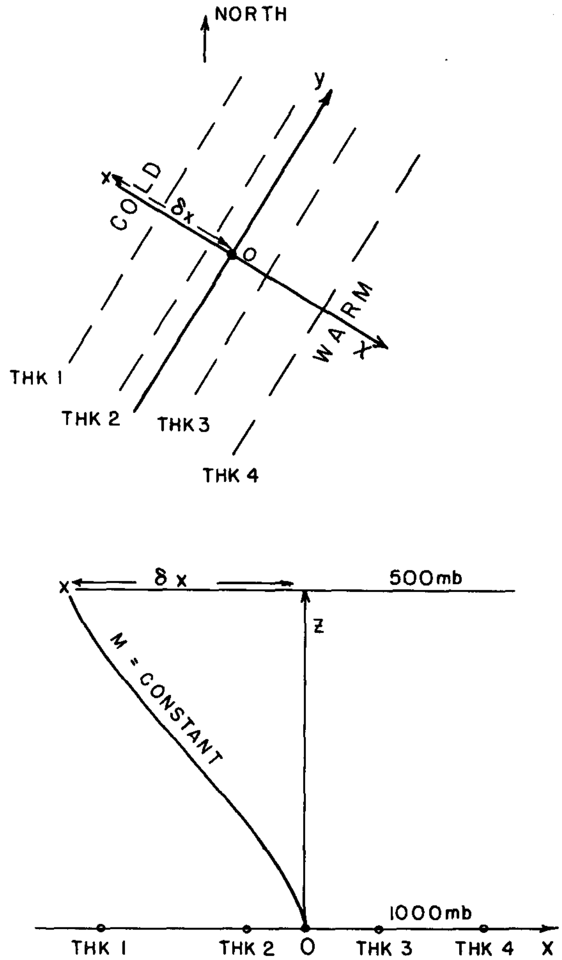

FIG. 1. Schematic of M surface geometry for two dimensional flow. Top figure is a top view; dashed are lines of constant 1000–500 mb thickness with values increasing toward $+x$. The "O" can be taken as the location of the cyclone center; the cross represents the projection of the intersection of the M surface originating from O with the 500 mb level. Bottom figure is a vertical cross section along the x axis. The curved solid line is a projection of the M surface originating from O whose intersection with the 500 mb surface is marked by the cross in the upper left.

subtract the new value of 500 mb height (at the location of the cross in the figure) from the 1000 mb height at the origin, O. It should be remarked that the utility of M surfaces as we have defined them depends on the validity of the geostrophic momentum approximation locally. Conservation of this quantity requires a small pressure gradient in y or, equivalently, small geostrophic temperature advection along M. By choosing our origin, O, at the cyclone center, the approximation is valid at low levels by definition. The M surfaces will slope toward $-x$ with height and if the shear vector does not rotate, $\partial p/\partial y$ should remain small at all points along M.

In the latter stages of explosive development our assumptions may break down. As it turns out, however, most of the potential for heating occurs in the initial stages of growth in our region of study. For the cases examined (see also S2), the mean position at the end of the 24-hour maximum deepening was near Sable Island, Nova Scotia (44°N, 60°W) where SSTs average 3°–5°C in wintertime, and the potential for warming is usually small (see also section 3 and Fig. 5). Thus, consequences of the breakdown of our assumptions are likely not severe.

Having our measures of potential warming, Δ_M and Δ_U, we now seek to determine whether or not they relate to the underprediction of the LFM for a number of cases of Western Atlantic explosive cyclogenesis. The cases we will examine compose the majority of the sample studied in S2, occurring over a 4-year period from January 1981 to January 1985 (see Table 1). After this time, a change was made in the LFM which involved the modification of the surface drag formulation. The stress was forced to go to zero at the top of the second model layer (about 700 mb) instead of the boundary layer (950 mb). The intent was to improve forecasts of lee troughs and cyclones, but it may have had an effect on oceanic cyclones as well (NWS Tech Bulletin 348, 1985). Due to the small sample of available cases after January 1985, we will focus only on prior events.

In addition to satisfying the bomb criterion, each storm had to have the majority of its deepening occur over water and remain within the LFM forecast grid field for the entire 24-hour period. All LFM 4-panel outputs and NMC analyses were taken from the MIT archive. We used the LFM forecast initialized at the start of the 24-hour period of most rapid deepening (time 0) as an estimate of the cyclone growth that would have occurred with minimal influence of sea surface fluxes. We calculated the quantities Δ_M and Δ_U at times 0, 12 and 24 hours with our origin following the NMC *analyzed* positions of the cyclone center and set negative values equal to zero. For comparison with the model underprediction, we used the average of these quantities over the three times ($\bar{\Delta}_M$ and $\bar{\Delta}_U$). The sea surface temperature under the cyclone center was estimated from the most recent regional SST analysis

TABLE 1. Cases used in LFM study.

Category	Time of initialization YY/MM/DD/HH (UTC)	Maximum bergerons
Strong	81/01/10/00	1.8
	81/03/03/00	2.0
	81/03/16/12	2.0
	81/12/05/12	1.8
	82/02/13/00	2.5
	83/01/06/00	1.8
	85/01/04/12	1.9
	85/01/11/00	2.0
Moderate	81/01/17/12	1.6
	81/03/05/12	1.4
	81/03/14/12	1.4
	82/02/19/12	1.6
	82/04/06/00	1.7
	82/11/15/12	1.3
	83/02/12/00	1.4
	83/03/11/12	1.4
	84/01/27/12	1.7
	84/01/31/00	1.4
Weak	81/01/12/00	1.2
	81/01/29/00	1.2
	81/02/08/00	1.0
	82/02/21/12	1.1
	82/04/09/12	1.1
	83/01/15/12	1.1
	83/03/24/12	1.2
	83/03/31/12	1.2
	84/02/06/12	1.0
	84/03/09/00	1.0
	84/06/03/12	1.0

available before the event with a probable error of ±0.5°C. The analyzed cyclone center positions were used since we wanted to calculate PT based on the SST "experienced" by the real storm.

For the amount of underprediction by the LFM, we used the difference between the forecasted central pressure fall and the observed pressure fall taken from NMC analyses, thus accounting for initial discrepancies in pressure (which we will consider errors). Cases having an absolute value of initialization error >4 mb were excluded. This difference was denoted δp such that $\delta p > 0$ denotes an underprediction. The use of central pressure fall to measure intensification may have some drawbacks, but for our purposes, it is both convenient and appropriate since unforecasted net interior heating from below will result in a positive δp from simple hydrostatics.

One last important consideration involves the assumption we have made that the LFM will correctly forecast the development that would have occurred with a minimal influence of sea surface fluxes of moist entropy. Under circumstances such as a poor initialization, either at the surface or aloft, or a poor treatment of the evolution of the upper level field early in the development, this assumption may be invalid. We have removed cases with large initialization errors in cyclone central pressure, but potential problems at upper levels remain. As implied by Silberberg and Bosart, a realistic representation of the evolution of mobile upper-level features is important for accurate forecasts of continental cyclogenesis. Since we had no reliable way to correct for mistakes already made, we simply compared the strength of the upper level forcing in the LFM forecasts and subsequent analyses, which is typically governed by the advection of absolute vorticity aloft (S1). Cases in which the forecasted vorticity advection, averaged over the 24-hour period, was not within 50% of the observed were excluded. Only two events were eliminated by this requirement. One was an instance of extraordinary *overprediction* of intensification (discussed in S2), the other a case of virtually no forecasted development (considered in S1).

Figures 2a and 2b are scatter diagrams of δp versus $\bar{\Delta}_M$ and $\bar{\Delta}_U$ for a total of 29 bombs. The cases were stratified into three arbitrary categories of deepening rates analogous to those in S1 and S2; weak (1.0B–1.2B), moderate (1.3B–1.7B) and strong ($\geq$1.8B). The correlation coefficient for the entire sample is 0.50 for cases in Fig. 2a and 0.41 for those in Fig. 2b. However, if we choose to divide the total sample in Fig. 2a into a subset of weak cases and one of moderate and strong cases, we get substantially higher correlations for each subset than we do for the total sample. Examining δp versus $\bar{\Delta}_M$, a least squares fit yields the relation

$$\delta p = 0.48\bar{\Delta}_M + 4.78 \text{ mb}$$

$$\text{correlation coeff.} = 0.81 \quad (4)$$

for moderate and strong bombs, while weak bombs fit to

$$\delta p = 0.20\bar{\Delta}_M + 3.05 \text{ mb}$$

$$\text{correlation coeff.} = 0.63. \quad (5)$$

For Δp versus Δ_U, the same partitioning yields a less remarkable difference in correlation from the sample as a whole. The moderate and strong cases fit to

$$\delta p = 0.51\bar{\Delta}_U + 7.33 \text{ mb}$$

$$\text{correlation coeff.} = 0.65 \quad (6)$$

and the weak cases to

$$\delta p = 0.28\bar{\Delta}_U + 3.80 \text{ mb}$$

$$\text{correlation coeff.} = 0.47. \quad (7)$$

By performing statistical tests, (generally described in Freund and Walpole 1980), we deduced that all correlations except (7) were significant at the 95% level of confidence or better. Differences in the correlations were less significant. We did not test subsets against the entire sample because these are not independent but we have compared subset correlations. Of interest

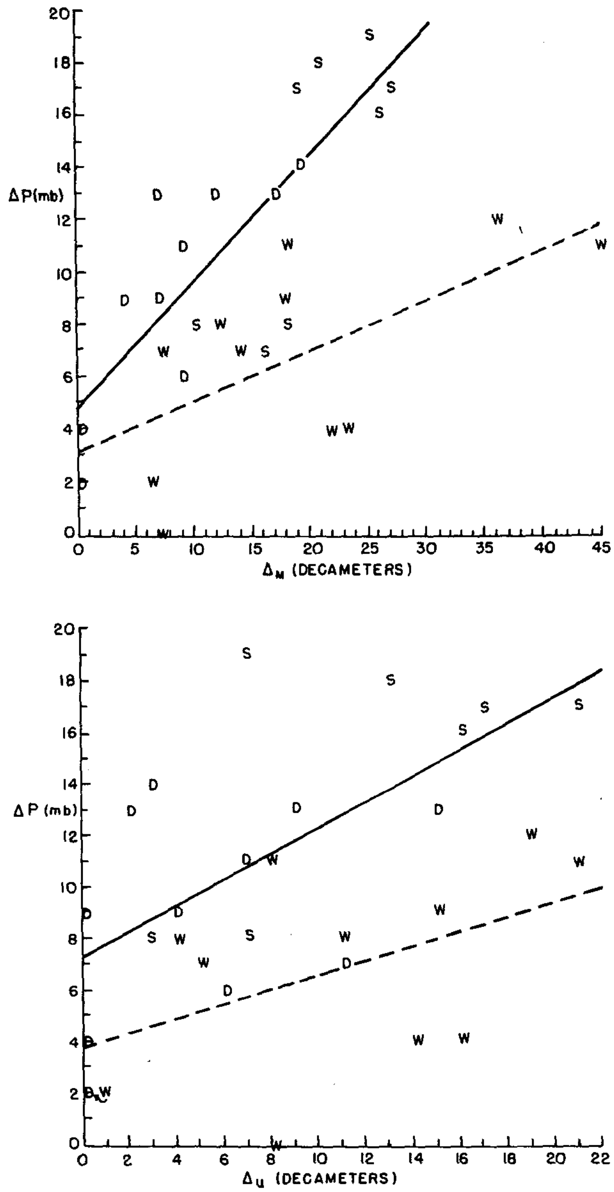

FIG. 2. Scatter plots of δp (mb) versus $\bar{\Delta}_M$ and $\bar{\Delta}_U$ (decameters); S: strong bombs, D: moderate and W: weak. (a) δp vs $\bar{\Delta}_M$ and (b) δp vs $\bar{\Delta}_U$ include all 29 cases; solid lines are best fits for cases $\geq$ 1.3B, dashed are for cases < 1.3B.

perhaps is the confidence level of the difference in correlation coefficient between δp versus $\bar{\Delta}_M$ and δp versus $\bar{\Delta}_U$ for moderate and strong bombs. This is about 80%, a rather modest value, but since we had reason to *expect* on physical grounds that the distribution of heating due to fluxes of moist entropy from the sea surface should occur along M surfaces, the higher correlation for δp versus $\bar{\Delta}_M$ is at least consistent. In addition, the robustness of the linear fit is in qualitative agreement with theoretical results obtained by M. Fantini (personal communication). In a semigeostrophic, analytical model, he found that the growth of the most unstable Eady mode increased almost linearly with the value of a parameter directly proportional to $\bar{\Delta}_M$. The correlations cited above are consistent with our original hypothesis that the failure of the LFM was systematically related to the amount of warming that could occur from moist entropy fluxes from the ocean surface. The slopes of the linear regression are also of the order that one might expect from hydrostatic arguments, all of which implies that neglect of this physical process may be a serious deficiency in a model attempting to forecast rapid oceanic cyclogenesis.

As a final note to this section, we turn to Fig. 3 which shows the correlation coefficient as a function of the minimum deepening rate allowed in a subsample of the total sample; the δp versus $\bar{\Delta}_M$ coefficients make up the solid line, those from δp versus $\bar{\Delta}_U$ are the dashed line. Results show that the improved correlations for δp versus $\bar{\Delta}_M$ do not depend on the deepening rate chosen to partition the total sample. The figure shows that the 1.3B value is near the separation between two regimes of deepening rate with different correlation properties. A test for the significance of the difference in slope between relations (4) and (5) yields a confidence level higher than 99%. Such behavior implies a marked transition in the sensitivity of the model and may suggest the presence of a finite amplitude mechanism. The word "amplitude" may seem incorrect here, but note that we are considering cases which are all initially weak perturbations, deepening over a fixed interval of time, implying a strong relation between amplitude and deepening rate. We note that a

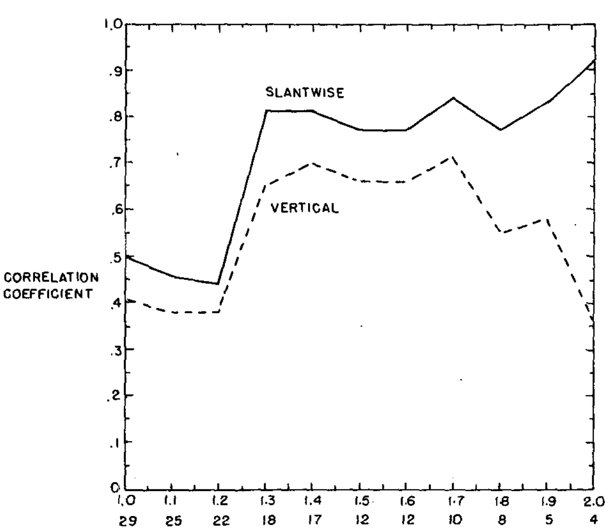

FIG. 3. Correlation coefficient (in the linear regression of δp vs $\bar{\Delta}_M$ and $\bar{\Delta}_U$) as a function of the minimum deepening rate allowed in a subsample. Correlation coefficients for δp versus $\bar{\Delta}_M$ form the solid line; those for δp versus $\bar{\Delta}_U$ form the dashed line. The bottom set of numbers below the abscissa denotes the number of cases in that subsample.

finite amplitude effect, which is the nature of the air-sea interaction process, could offer an explanation of the non-normal tail on Roebber's distribution of deepening rates. Our apparent "threshold" near 1.3B corresponds roughly to the point of maximum departure of his distribution from normal. Differences could be due to the fact that Roebber looked at all cyclones over the Northern Hemisphere whereas we restricted our attention to bombs over the western Atlantic.

3. Climatology of Δ_M

One question that arises from the preceding section pertains to the importance of Δ_M or Δ_U in explosive cyclogenesis in general, i.e., whether regions of appreciable bomb frequency are also regions where the potential heating from the sea surface is great in a climatological sense.

To address this issue, we have calculated the climatological mean of Δ_M (results for Δ_U, not shown, are similar) with an important deviation from a typical climatological calculation. We only averaged over times when $\Delta_M > 0$, since only then can sea surface fluxes warm the atmosphere.

Data for this calculation consisted of 15 years (1963–77) of twice daily data on a 4° latitude by 5° longitude mesh obtained on tape from the National Center for Atmospheric Research (NCAR) as well as monthly mean sea surface temperature analyses obtained from the National Oceanic and Atmospheric Administration (NOAA, NWS Technical Report 31, 1982) and from data available at MIT. These were gridded manually to the same grid spacing as the above tape data. Contained on the tapes were final NMC analyses of 500 mb heights and sea level pressures, from which we calculated the 1000–500 mb M surface thickness at each 12 hour interval, at each point. This was subtracted from the value of potential thickness obtained from the SST field and the *positive* differences averaged together to form the climatology.[1]

To calculate M surface thicknesses, a method different from that discussed previously for manual analysis is desirable. It proves convenient to make a transformation of coordinates in which M surfaces become vertical. These new coordinates are termed "geostrophic coordinates" and represent the coordinates a parcel would have if it moved at all times with the geostrophic wind. Geostrophic coordinates are described in detail in Hoskins and Bretherton (1972) and Hoskins (1975). They are

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[1] There is a small correction to the potential thickness when the surface pressure is not 1000 mb. This amounts to approximately 1 meter × (1000 mb − SLP)/1 mb or an amount of 2 to 3 decameters at most. It was not included in section 2 since the mean central pressure during the explosive development was only slightly below 1000 mb. We will include it in this section however.

$$X = x + \frac{v_g}{f}$$
$$Y = y - \frac{u_g}{f}$$
$$P = p$$
$$\tau = t \tag{8}$$

and to go along with this we define a modified geopotential

$$\Phi = \phi + \frac{1}{2}(u_g{}^2 + v_g{}^2) \tag{9}$$

from which follow

$$\frac{\partial \Phi}{\partial X} = \frac{\partial \phi}{\partial x}, \quad \frac{\partial \Phi}{\partial Y} = \frac{\partial \phi}{\partial y}, \quad \frac{\partial \Phi}{\partial P} = \frac{\partial \phi}{\partial p}. \tag{10}$$

If we consider two dimensional flow as before, such that the geostrophic momentum approximation is valid, then $M = v_g + fx = fX$, independent of P. Note, too, that with f constant, we are free to orient our horizontal coordinates such that x is always orthogonal to the mean shear.

Thus to calculate Δ_M, we make the transformations in (8) and (9) at both the 500 mb and 1000 mb pressure levels. Then, being in geostrophic coordinates, we simply take the difference between the *modified* geopotential heights at those levels. This value is equal to the thickness along an M surface originating from the 1000 mb point in physical space. In practice, we had to interpolate among the 500 mb modified geopotential heights to obtain a value lying directly above the desired location at 1000 mb. This was done by performing a weighted average among the five nearest points at 500 mb. The weighting factor was chosen to be the reciprocal of the horizontal distance between each point in 500 mb transformed space and the desired position above the 1000 mb point (see Fig. 4).

The climatological values of positive Δ_M in physical space (termed Δ_C) appear in Fig. 5a while the bomb climatology of SG appears in 5b. Of interest are the regions over the Western Atlantic and Western Pacific where Δ_C is large in a mean sense. This is not surprising when one considers the proximity of the warm western boundary ocean currents to the cold continental air. In addition, enhanced regions of Δ_C appear in the Gulf of Alaska, the Barent Sea north of Scandinavia and in the Mediterranean Sea. Of these areas, only the Gulf of Alaska has an appreciable frequency of bombs (Winston 1954; Murty et al. 1983). However, as noted by Businger (1985), the Barent Sea is a site of frequent rapid cyclogenesis, but often over a duration too short to satisfy the criterion of SG. The Mediterranean, while a location of frequent cyclogenesis, has few rapidly deepening events (Gleeson 1954).

We point out that large values of Δ_C would be found in regions of cold air outflow to the west of cyclones

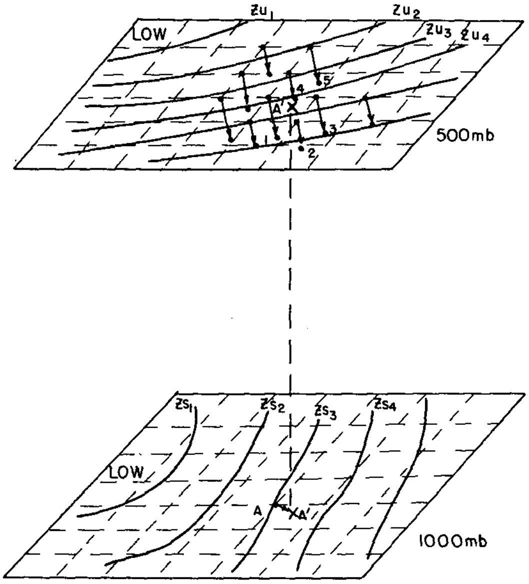


FIG. 4. Schematic of the calculation of 1000–500 mb thickness in geostrophic coordinates; broken lines on each surface represent the data grid, solid lines are hypothetical height contours. Arrows indicate vector displacement via the coordinate transformations in Eq. (8). "A" is the physical space 1000 mb point of interest, "A" prime is that point transformed to geostrophic coordinates; numbers 1 to 5 denote the 5 points at 500 mb nearest "A" prime.

which would actually do little to enhance their development, but could strongly influence our calculation of Δ_C. However, it is possible that this cold air outflow may affect the environment for subsequent cyclogenesis upstream of the original cyclone. Situations where cyclones approach regions of large Δ_M then rapidly develop have been documented (Bosart 1981; Gyakum 1983; Reed and Albright 1986). In a typical synoptic situation such as this, we would see southerly surface winds ahead of the nascent bomb. This and the fact that the surface heat fluxes are proportional to the surface horizonal velocity suggest that a climatology of Δ_M, weighted by positive meridional surface velocity might be appropriate. We therefore recalculated the climatology in the following form:

$$\Delta_{Cv} = \frac{\sum_{i=1}^{n} v_{g_i} \Delta_{M_i}}{\sum_{i=1}^{n} v_{g_i}} \tag{11}$$

where n is the number of times at an individual gridpoint in 15 winters of twice daily analyses (December–February) that Δ_M *and* v_g (the southerly component of surface geostrophic wind) are >0 (typical values of n are a few hundred). The field of Δ_{Cv} for the three months averaged together is shown in Fig. 5c. The important result here is that, having removed most of the

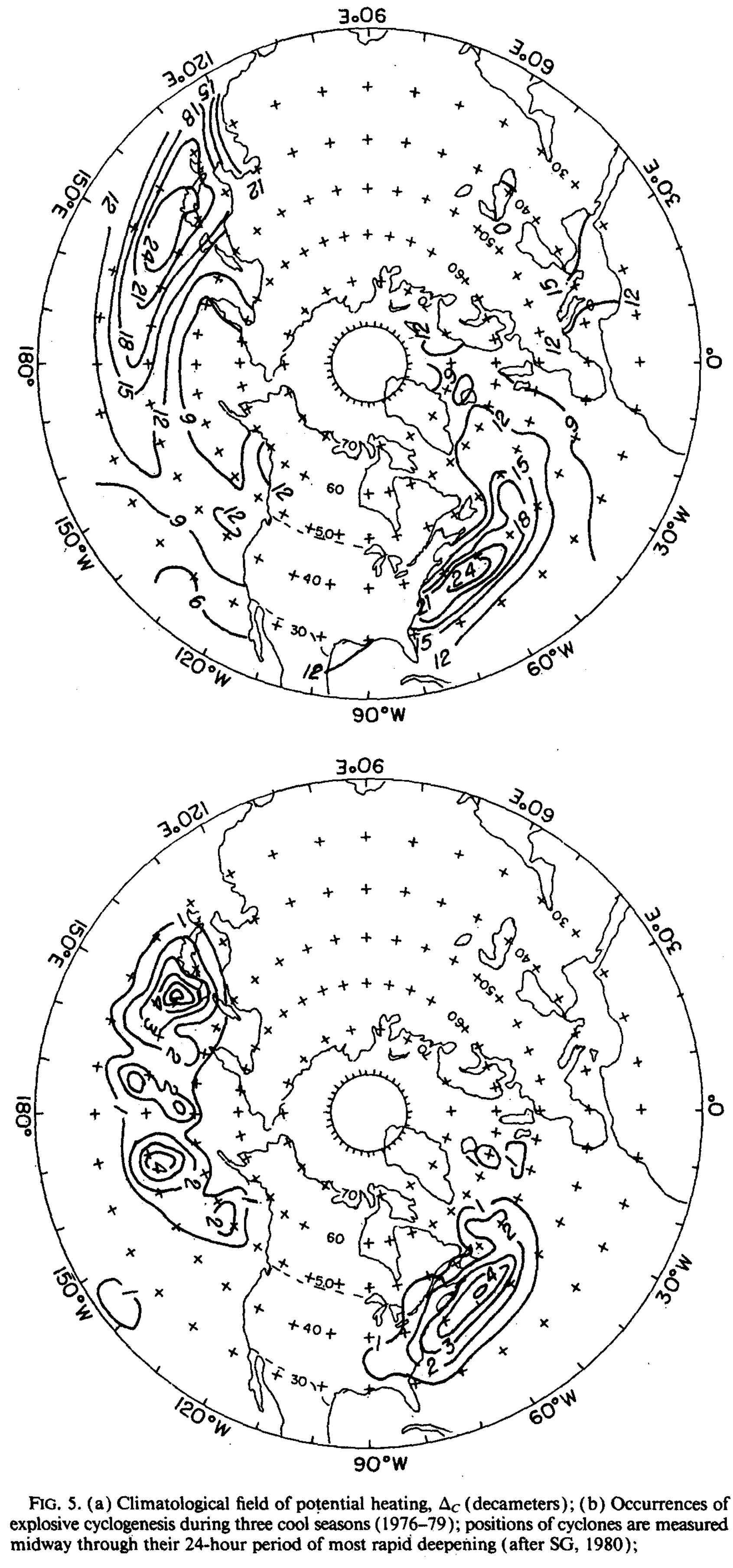
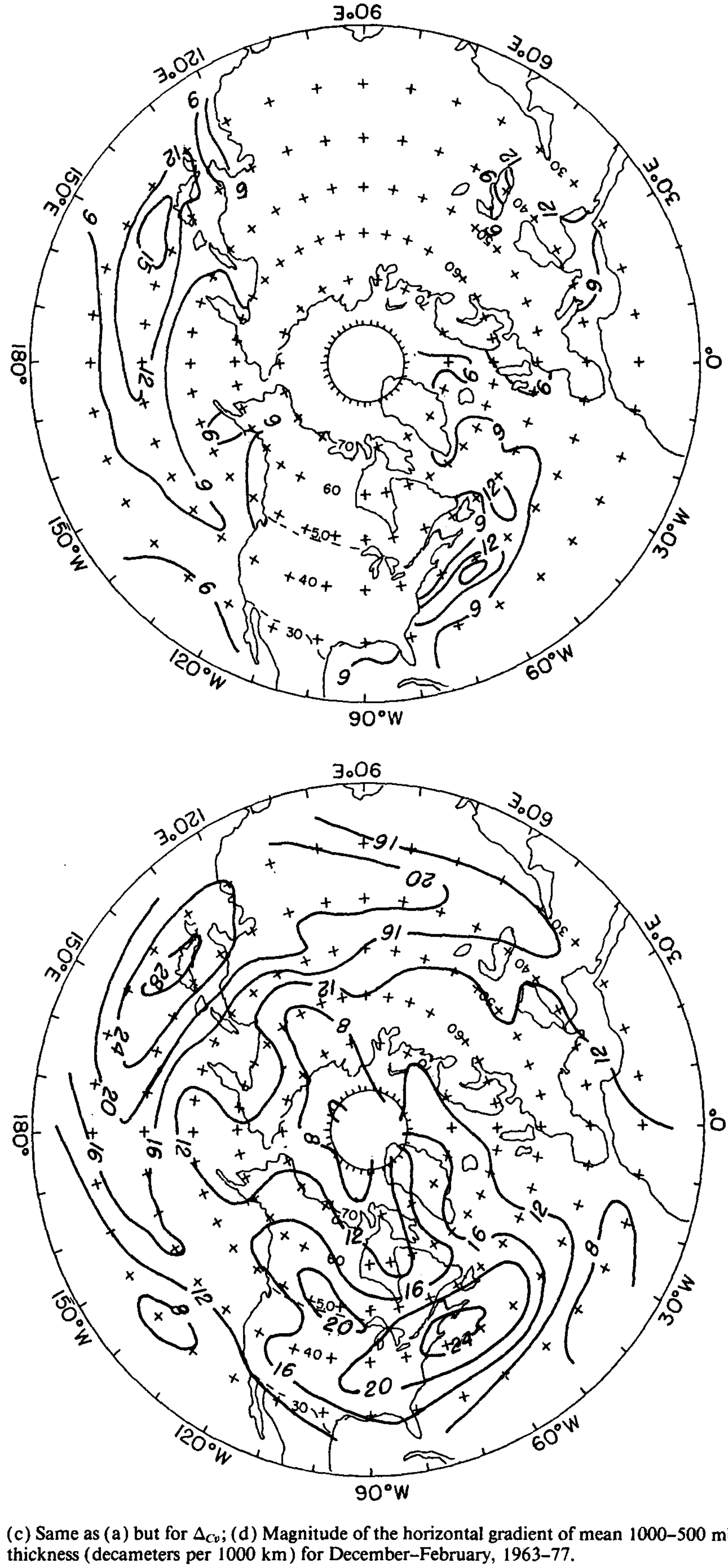


FIG. 5. (a) Climatological field of potential heating, Δ_C (decameters); (b) Occurrences of explosive cyclogenesis during three cool seasons (1976–79); positions of cyclones are measured midway through their 24-hour period of most rapid deepening (after SG, 1980);



(c) Same as (a) but for Δ_{Cv}; (d) Magnitude of the horizontal gradient of mean 1000–500 mb thickness (decameters per 1000 km) for December–February, 1963–77.

effects from cold surges, the pattern of potential heating is qualitatively the same as in Fig. 5a.

We should point out that neither Fig. 5a nor Fig. 5c show anomalously large values of potential heating near the bomb maximum found in the east central Pacific. However, it is apparent from the study done by Murty et al. (1983) that peak bomb frequency occurs in October and November here, trailing off substantially by January and February. In addition, there is an uncertain contribution to the robustness of this maximum by the surface data distribution in the area as mentioned by Roebber. Cyclones entering a region of improved data coverage may appear to deepen owing to an increased ability to resolve the strength of their circulations. It is therefore not clear whether our results in this region conflict with our stated hypothesis.

The existence of a large potential for heating by the ocean in a mean sense over the western ocean regions and near other regions of frequent rapid cyclogenesis is consistent with our view of the importance of this effect. Nonetheless, the proximity of these regions to areas of strong climatological horizontal temperature gradient (Fig. 5d) does not permit a clear separation between our mechanism and baroclinic processes that typically influence rapid cyclogenesis.

5. Summary and conclusion

The principle aim of this work was to test the notion that the fluxes of latent and sensible heat from the ocean surface are important in the phenomenon of explosive cyclogenesis. Our hypothesis was that these fluxes would enhance already robust cyclone development to the status of "explosive" development. Operational NMC forecast models, in particular the LFM, until very recently, regularly underforecasted such intense developments. There was reason to believe that the process of evaporation of moisture from the sea surface and subsequent distribution of latent heating in the interior was not handled properly by the LFM. We thus postulated that the shortcoming of LFM predicted deepening was systematically related to the effect of these fluxes. The study we conducted to support this hypothesis involved 29 cases of explosive cyclogenesis over the Western Atlantic Ocean. For each case we calculated the maximum amount of 1000–500 mb thickness increase that could occur due to warming from the sea surface. Especially for the more rapid deepeners of the sample, a rather robust correlation between the underprediction of storm central pressure fall and the magnitude of this potential warming was found, consistent with our hypothesis. We found the correlation was somewhat stronger when the potential for warming was measured on surfaces of constant pseudomomentum rather than in the vertical. This was consistent with the admittedly small number of careful observations of cyclones during GALE which indicated a tendency for their environments to have neutral moist stability along such surfaces, but pronounced stability in the vertical.

In addition, there was a significant increase in the magnitude of underprediction for a given potential warming as observed cyclone deepening rates exceeded 1.2 bergerons. This may have several implications. First, use of the linear relations in forecasting is limited because one must know something about the *real* deepening rate of an event in advance. Second, such a change in slope suggests that we are dealing with a finite amplitude mechanism. This is consistent with our hypothesis that a poor treatment of air–sea interaction is at the heart of the LFM shortcoming, since this process is finite amplitude in nature. Third, the presence of such a mechanism would imply a tendency for bimodality in the distribution of deepening rates, perhaps analogous to Roebber's findings. We are able to offer a physical explanation for neither the apparent sharpness in transition from one correlation regime to another, nor the location of the behavior change in the deepening rate spectrum.

The second portion of this study focussed on the relationship of regions where the potential for heating from below is large to regions of frequent explosive cyclogenesis. Overall, the areas coincided quite well. We were not able to separate cleanly the influence of ordinary baroclinic processes, but this is neither completely necessary nor feasible. Since the potential for warming along M surfaces for a given sea surface temperature increases with increasing shear and decreasing static stability, the two are strongly coupled. Furthermore, baroclinic processes are perhaps the most likely source of initial circulations possessing sufficient strength for an air-sea interaction to operate efficiently.

From the results of this paper, one can only say that we have offered more evidence that effects of the underlying ocean may be crucial, and not merely circumstantial, for bomb development. We have no direct measurements that these fluxes are actually being realized. In fact, conventional observations are probably inadequate to resolve such an issue, which leaves open the possibility of other interpretations of our results. The need for further numerical experiments is clear, but the real problem is to explain the non-Gaussian character of the intensification spectrum rather than to explain why individual events intensify at rates exceeding an arbitrary threshold. The mere fact that a model simulates an explosive event with impressive accuracy means very little if we do not understand the underlying reasons for the success and have confidence that those reasons are extendable to rapid oceanic cyclogenesis in general.

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