EVOLUTION OF SHALLOW BAROCLINIC DISTURBANCES IN SATURATED ENVIRONMENT

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1. Introduction

The work by Emanuel and Rotunno (1989) suggested that the rapidly deepening, small scale, 'polar lows' frequently observed in the high latitudes of both hemispheres could be maintained in the mature stage by the collection of latent and sensible heat at the air-sea interface and its redistribution in mid-troposphere performed by convergence of cross-isobar flow into a central updraft, in the same manner invoked previously by Emanuel (1986), Rotunno and Emanuel (1987) for the steady-state of tropical storms, the main difference between the two situations being the prevalence of latent heating in the latter versus the near equality of the two forms of energy exchange in the former. The suggested mechanism would only provide maintenance of a steady-state, and would only be effective at finite amplitude. Therefore some other means of initiating the process was required, and baroclinic instability was suggested as an obvious candidate.

This suggestion was pursued by Fantini and Buzzi (1993; 1994; 1997) by studying the small and finite amplitude characteristics of two-dimensional baroclinic waves confined to a shallow saturated layer, and showing that their evolution displays spatial and temporal scales compatible with a role of 'seeds' of polar lows. For clear geometric reasons the actual formation of the latter could not be studied with the slab-symmetric model used in those papers.

The present paper presents preliminary steps in the direction of an extension to three dimensions of the previously mentioned work. A primitive equation model with moisture representation was written by the author, and will be briefly described in the next Section, and experiments in an environment saturated only in the lowest levels of the troposphere, with or without the presence of heat fluxes at the air-sea interface, were performed to provide a reference life-cycle for a baroclinic/polar-low type of storm.

Further work will be needed to translate the present result in a less idealized context before actual comparison can be made with observed events. The present study is only intended to identify qualitatively the behaviour of atmospheric disturbances in the environment mentioned above. The limitations of the model, as described in the next section are too obvious to be further remarked upon.

2. Model

The model integrates the primitive equations on a B-grid in a doubly periodic f-plane. The environment is defined as in Eady model, *i.e.* uniform shear and dry static stability, Boussinesq approximation, rigid lid at the top of the troposphere. Moisture representation follows in part Emanuel *et al.* (1987), in that a small value of static stability is applied in points where the vertical velocity is directed upward and the air is saturated. The same model, with the exception of the latter condition, was presented in Fantini (1998) with a study of deep normal-mode moist baroclinic waves. In order to represent a more structured moisture distribution an equation for specific humidity was added to the model so that the small static stability, which accounts for the enhanced buoyancy of air parcels when release of latent heat is occurring, can be applied just at saturated grid points rather than everywhere there is ascent, as was done in previous papers. This feature will also allow us to include a moisture flux at the air-sea interface and let the model redistribute it consistently with its dynamics.

The surface temperature is chosen at 285 K at the southern boundary and decreases northward in geostrophic balance with the zonal wind, which is 0 m s^{-1} at the ground and increases uniformly by 30 m s^{-1} over the 10 km height of the model. All experiments are run in a 3,000 x 6,000 km domain represented in 64 x 64 grid points, and 11 vertical levels, with vertical velocity levels half a grid point above and below the θ and wind levels. The heating due to surface fluxes is included at the lowest model level as

$$\dot{Q} = \frac{c_D}{h}|\mathbf{u}_{z=0}|\,(\theta_{sea} - \theta_{z=0})$$

where h is the height of the lowest model level (450 m), having assumed that heat received from the sea is uniformly distributed over that height. The potential temperature of the sea surface θ_{sea} is defined as the base potential temperature of surface air plus a $\Delta\theta_{sea}$ which varies in the experiments presented below. The drag coefficient is $c_D = .001$, chosen for a smooth surface and with no wind dependence. Moisture flux is treated in a similar manner, considering 'sea' humidity to be the saturation value at the sea surface

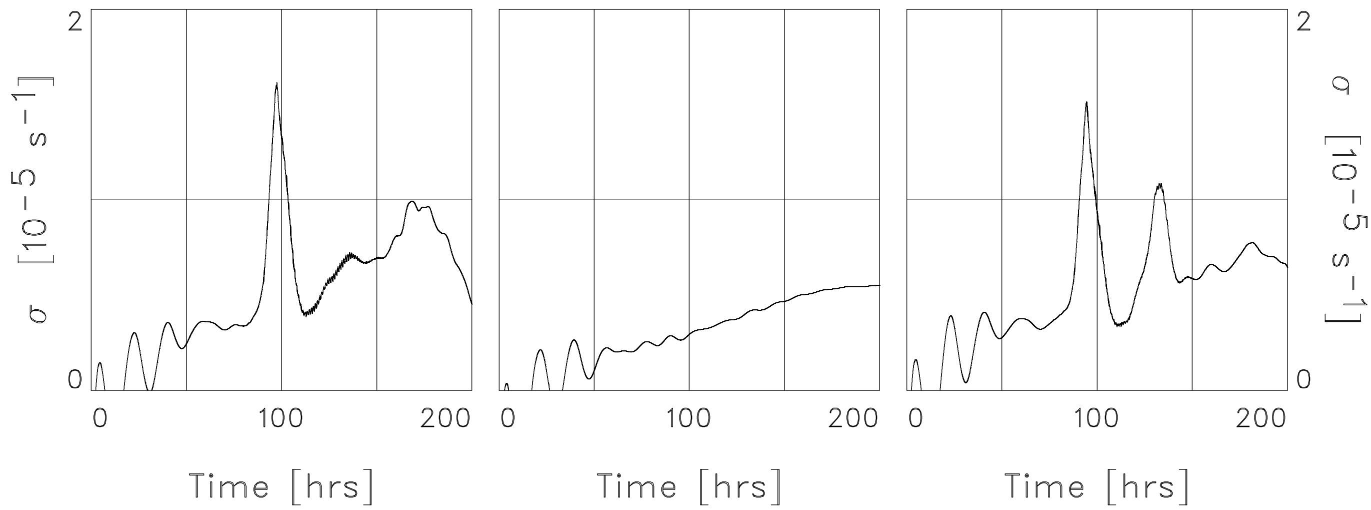

Figure 1. Plot of growth rate vs simulation time in hours for three experiments. Left: Exp. A (no surface fluxes); Center: Exp. B ($\Delta\theta = 0k$); Right: Exp. C ($\Delta\theta = 6k$).

temperature and pressure. When the perturbation potential temperature is large enough to overcome the imposed $\Delta\theta_{sea}$ cooling of surface air can occur, and moisture above the saturation point is then lost.

3. Results

On a basic state saturated only in the lowest 2.5 km of the atmosphere a compact perturbation of pressure is superimposed. The perturbation has an initial amplitude of 1.5 mb at the surface and decays exponentially with height with a vertical scale of 1 km. It extends horizontally with radial symmetry, going to zero at 1000 km from the center. All other initial fields are geostrophically balanced to the pressure perturbation.

Fig.1 shows the evolution of growth rate, computed from perturbation kinetic energy, for three experiments. Experiment A is a reference run with no surface fluxes, *i.e.* $c_D = 0$. After an adjustment lasting about 50 hours the perturbation starts growing on the vertical scale of the saturated layer, and its corresponding Rossby length in the horizontal. The vertical structure remains more upright than one would expect in a baroclinic wave, although the coarse vertical resolution does not allow us to make a strong statement on this point. Around 100 hours a 'mesoscale convective episode' takes place, constituted of a strong mesoscale updraft which extends into the mid-troposphere, carrying along warm and moist air. The tight horizontal structure, and relatively weak thermal perturbation, of the cyclone at this stage is shown in Fig.2. This situation cannot be maintained for more than 20-30 hours, after which a more traditional deep baroclinic mode takes over (Fig.3).

The center panel in Fig.1 refers to a run in which the surface fluxes are active but there is no basic temperature difference between surface air and the sea surface, so that only the thermal perturbation associated with the cyclone drives the heat and moisture exchange. In this situation a warm sector is heated less than a cold one, so that the effect of the surface fluxes is to damp the existing atmospheric perturbation. Moreover, water vapor is lost by the air in the process, to the point that convergence induced by the secondary circulation is not sufficient to destabilize the thermodynamic profile. As a consequence the strong mesoscale updraft denoted by a peak in growth rate in Exp.A does not appear, and the whole experiment displays a smooth evolution from the initial shallow to a final deep baroclinic mode, which takes over because more unstable in this domain.

Finally, the left panel in Fig.1 refers to an experiment in which a 6 K air-sea temperature difference is imposed. The value was chosen so as to reproduce the same timing of Exp.A, while smaller $\Delta\theta_{sea}$'s give rise to a strong mesoscale updraft at later times. The first peak in growth rate reproduces exactly the behaviour of Exp.A, while the second one represents a smaller cyclone forming about 1,000 km to the rear of the first one, but eventually merging with it.

From the results just shown we draw the preliminary conclusion that the development occurring in Exp.A is spurious, in the sense that a marine environment cannot be represented without interfacial energy exchange, and when this aspect is taken into account, as in Exp.B, local excess of heat and moisture is more likely to be disposed of at the lower boundary rather than develop into a recognizable atmospheric disturbance on the mesoscale. Only a non-negligible energy input at the lower boundary is able to reconstitute the conditions needed for the latter development.

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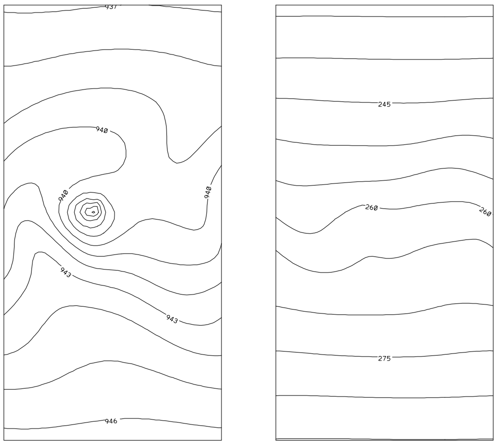

Figure 2. Fields of pressure (left) and potential temperature (right) al lowest internal level for Exp. A at 100 hours

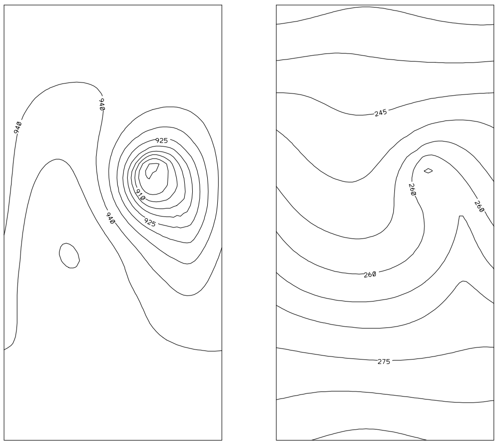

Figure 3. Same as Fig.2 but at 185 hours

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