A note on boundary-layer friction in baroclinic cyclones

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Abstract: The interaction between extratropical cyclones and the underlying boundary layer has been a topic of recent discussion in papers by [Adamson et al. (2006) and Beare (2007)]. Their results emphasise different mechanisms through which the boundary layer dynamics may modify the growth of a baroclinic cyclone. By using different sea-surface temperature distributions and comparing the low-level winds, the differences are exposed and both of the proposed mechanisms appear to be acting within a single simulation.

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1 Introduction

Recent papers by [Adamson et al. (2006) and Beare (2007)] have provided two differing explanations for the way in which the boundary layer affects dry, baroclinic cyclones. Adamson et al. (2006) provided a new mechanism for the frictional spin-down of an extratropical cyclone, based on baroclinic potential vorticity (PV) generation. PV generated in the boundary layer (due to a component of stress anti-parallel to the tropospheric thermal wind) is advected along the warm conveyor belt and ventilated from the boundary layer. It turns cyclonically and accumulates above the low centre as a positive PV anomaly of large horizontal extent, but trapped between isentropes in the vertical. This shape is associated mainly with increased static stability, inhibiting communication between the upper- and lower-level features of the developing wave, and so reducing growth.

However, Beare (2007) showed that the dominant low-level PV anomaly in his model was not associated with the spin-down of the cyclone, and that the region of greatest surface stress was most important in restricting cyclone growth. This region, to the west of the surface low centre, is characterised by a well-mixed boundary layer, and so is not associated with a significant contribution to the PV budget. Instead, its effects are thought to occur via the Ekman pumping mechanism – the additional stress enhances convergence of near surface winds, which must create upward vertical motion due to continuity. This reduces cyclone growth by vortex squashing in the interior.

This note aims to clarify the differences between the two studies and demonstrate that both mechanisms can be seen in the same simulation.

2 Methodology

To mimic the large-scale atmospheric structure, both previous studies specified a mid-latitude jet with a potential temperature profile calculated from thermal wind balance. However, Adamson et al. (2006) took their sea-surface temperature field to be identical to that of the first model level at the initial time, whilst Beare (2007) chose a constant sea-surface temperature across the entire domain.

The cyclone initialisation also differed between the studies. Adamson et al. (2006) followed the same initialisation as for the lifecycle denoted LC1 by Thorncroft et al. (1993), in which a normal mode perturbation is added to the basic state, evolving slowly into a cyclone over 5–10 days. The study of Beare (2007) included a finite-amplitude upper-level vortex in the initial conditions, triggering rapid cyclogenesis over 2–3 days, and showing much similarity to the conceptual model of Shapiro and Keyser (1990). Therefore, the experiments represent different realisations of the spectrum of real-world cyclones, and neither can claim to be closer to reality than the other. Here we consider an intermediate experiment, repeating the simulations of Beare (2007), but choosing our sea-surface temperature in the same manner as Adamson et al. (2006).

3 Results

The choice of sea-surface temperature (SST) in Beare (2007) has considerable effect on the evolution of the boundary layer. The low SST relative to the overlying air in the warm sector gives rise to a shallow (as little as 50 m in places, but on average about 400 m) and highly stable boundary layer. The large negative surface heat fluxes in this region lead to the PV budget being dominated by the surface heat-flux term (term 2 in Equation 11 of Beare (2007)); i.e.,

\[
\frac{DP}{Dt} \approx -\frac{\xi_h H_s}{\rho_0 h^2} \tag{1}
\]
where $P$ is the Ertel–Rossby potential vorticity, $\xi_h$ is the vertical component of absolute vorticity on the boundary layer top, $H_s$ is the surface heat flux, $\rho_0$ is the density, $h$ is the boundary layer depth and the square brackets indicate a depth-average over the boundary layer, after Cooper et al. (1992). It is this PV which is seen in his Fig. 4(b) (our Fig. 1(b)), rather than the PV generated by the baroclinic mechanism which is the focus of Adamson et al. (2006).

The importance of the surface heat-flux pattern under conditions of a horizontally uniform SST field is demonstrated by Fig. 1(c), where we have repeated the control experiment of Beare (2007) without any turbulent heat fluxes. This makes the simulation closer to that of Adamson et al. (2006), since their study focussed on the drag, and so their boundary layer scheme only parameterised momentum transfer and had no turbulent heat fluxes present. Comparing Fig. 1(c) to Fig. 1(b) shows that the PV is now located to the north and east of the cyclone (between $5$–$15^\circ E$, $50$–$60^\circ N$), confirming that the
boundary layer PV in the Beare (2007) control run is predominantly generated by turbulent heat fluxes. The PV generated by heat fluxes remains close to the surface and ahead of the cyclone centre (shown between $10 - 20^\circ E, 40 - 50^\circ N$ in Fig. 1(b)) during the course of the Beare (2007) simulation; i.e., it is not vented from the boundary layer. It therefore never reaches a position above the low centre and cannot prevent communication between upper and lower-level anomalies.

Beare's (2007) conclusion that the heat-flux generated PV is not dominant in the spin-down process is therefore justified. But the results are not contradictory to those of Adamson et al. (2006). Indeed, when the turbulent heat fluxes are switched off, the cyclone shows a slight filling of 2 hPa after 48 hours, which is consistent with the results of the PV inversion in Beare (2007) that the PV generated by the heat fluxes acts to deepen the cyclone.

We consider now the low-level jet seen in the simulations of Beare (2007). Formed by a reversal of the north-south temperature gradient generating an easterly wind shear, this cold air wraps around the cyclone centre, producing a cold conveyor belt. This provides ideal conditions for generation of large surface stress. The location of maximum surface stress, or equivalently of the friction velocity, is then found to have a large impact on cyclone development, consistent with the Ekman pumping mechanism. Such a low-level jet is not apparent in the LC1 lifecycle simulations of Adamson et al. (2006). Figure 2 shows the low-level winds in both simulations at an early stage of development. If these low-level winds are assumed to lie within the surface layer, the surface stress is given by the bulk aerodynamic formula:

$$\tau_s = \rho_0 C_D |v_1| v_1$$

(2)

where $\tau_s$ is the horizontal surface stress vector, $C_D$ is the drag coefficient and $v_1$ is the horizontal wind vector on the lower model level. The surface stress exerted on the cyclone is therefore proportional to the square of the low-level wind. It is noticeable that the strongest wind-speeds in Fig. (2b) are to the southwest of the low centre, between $-15$ and $-5^\circ E, 50 - 55^\circ N$, whereas in Fig. (2a) they are to the southeast, within the warm sector ($65 - 60^W, 45 - 50^\circ N$). Therefore, in the Adamson et al. (2006) experiment, the strongest winds, and therefore surface stress, are in a region of horizontal temperature gradients and hence significant PV generation. However, in Beare (2007) the low-level jet wrapping around the cyclone centre enhances wind-speeds to the southwest of the low, making the largest surface stress in a region of small horizontal temperature gradients and hence little PV generation.

By imposing a meridional SST gradient on the Beare (2007) experiment, we see both the Adamson et al. (2006) and Beare (2007) mechanisms at work in the same cyclone. The boundary layer depth-averaged PV, shown in Fig. (1d), is now located to the north and east of the cyclone centre, similarly to Fig. (1a) and consistent with Figs. 4, 5 and 11(c) of Adamson et al. (2006).
from the boundary layer. Plant and Belcher (2007) discuss how this ventilation occurs by the cold conveyor belt at early stages of the lifecycle, transitioning to ventilation by the warm conveyor belt at later stages, as the cyclone wraps up. Once advected out of the boundary layer, the PV appears as a static stability anomaly above the cyclone centre.

4 Conclusions

This note aimed to clarify why two recent papers have provided different emphases for the boundary layer’s interaction with an extratropical cyclone. We have shown why the experiments of Beard (2007) did not find evidence for the baroclinic mechanism, due to the boundary layer PV distribution being dominated by dynamically unimportant PV generation from surface heat fluxes. We have also shown that the much weaker low-level jet in the cold air stream southwest of the low centre in the experiments of Adamson et al. (2006) limited the Ekman pumping.
Through the addition of a meridional SST gradient to the simulations of Beare (2007), we have produced results which show both mechanisms at work. It is beyond the scope of this note to establish the relative importance of each mechanism. That may, for instance, require the development of techniques for PV inversion within the boundary layer.

In reality, there is of course a spectrum of mid-latitude cyclones, and it is plausible that each mechanism could be dominant in different types of cyclone. Ekman pumping is a direct mechanism, reducing the angular momentum of a pre-existing barotropic circulation, whilst the baroclinic PV mechanism is somewhat indirect, weakening the growth of a baroclinic wave. Certainly more work is thus required to form a complete understanding of the boundary layer processes in baroclinic cyclones.

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