6.1 WHY IS THERE MIDWINTER MINIMUM OF STORM TRACK?

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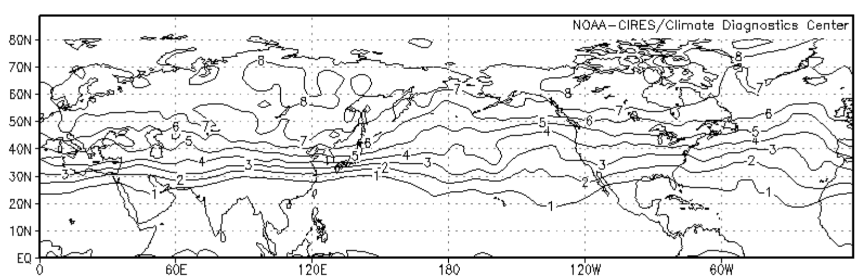
ABSTRACT

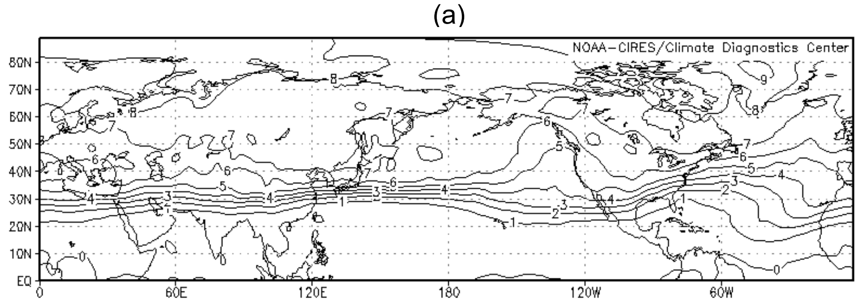
The counter-intuitive observation that the storm track over North Pacific has distinctly weaker intensity in mid-winter than in early- or late- winter remains largely unexplained. It is hypothesized that the phenomenon is a dynamical consequence of an increase in the isentropic potential vorticity (IPV) gradient across the tropopause over North Pacific during mid-winter. Such change would give rise to an increase in the barotropic-governor effect on local baroclinic instability. The observational basis of this hypothesis will be presented. The hypothesis is substantiated by a series of instability analyses with a quasi-geostrophic model that is constructed under the guidance of the dynamical considerations alluded to above. Sufficiently large downstream variation in the baroclincity and/or localized barotropic shear would give rise to baroclinic and/or barotropic localization of the unstable disturbances in the downstream region of the basic jet. The overall dynamical mechanism may be succinctly conceptualized as a mechanism of general wave-packet-resonance from the PV perspective. Additional insight of this process is deduced by diagnosing the local energetics of the unstable disturbances.

Model & Results

It is observed that the barotropic shear as well as the baroclinic shear of the North Pacific jet are strongest during mid-winter compared to early- or late- winter. This observed change is associated with a marked increase in the upper-tropospheric IPV gradient over North Pacific as exemplified by Fig.1a for November 1990 and Fig.1b for January 1991. This change suggests that the corresponding increase in the barotropic-governor effect could outweigh the increase in the baroclinic instability of the jet. It is hypothesized that the "mid-winter minimum" of the Pacific storm track could result from such change in the background flow. This hypothesis is tested in the context of a two-level QG channel model. The model basic PV distribution (Fig. 2) is an idealized version of the observed PV in January. The corresponding basic flow has a localized baroclinic jet and a downstream region of pronounced diffluence/confluence bounded by two wave-guides.



(a)



(b)

Fig. 1 Observed monthly mean 345K isentropic potential vorticity, in PV unit.
(a) November 1990 (b) January 1991 (midwinter).

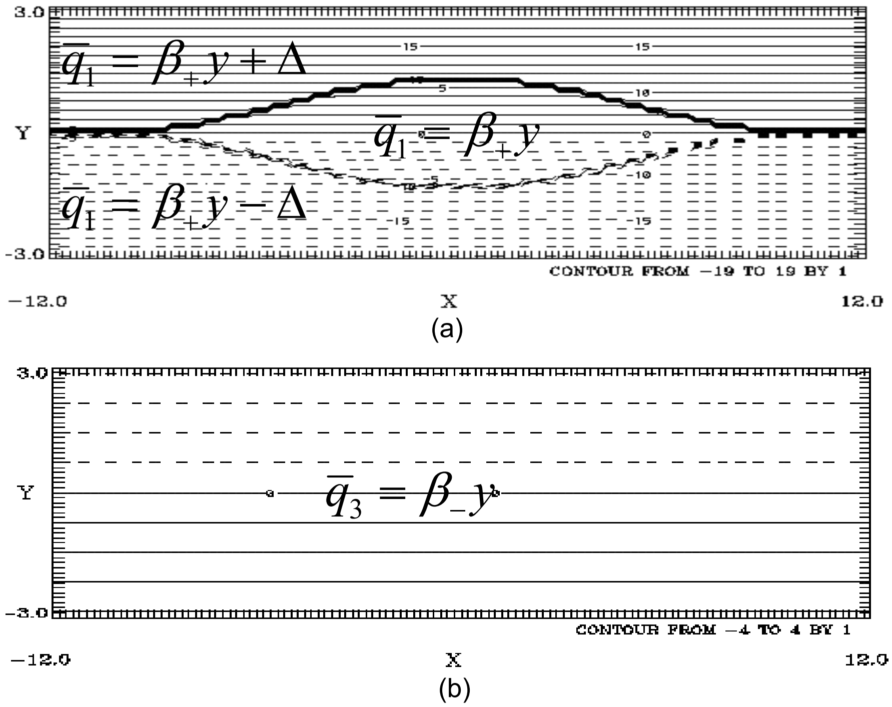


Fig. 2 Distribution of background non-dimensional PV in the model.
(a) Upper level (250mb) (b) Lower level (750mb). $\beta_{\pm} = \beta \pm \lambda^2 U_T$.

In Fig.2, U_T is a measure of the zonally uniform component of the basic baroclinicity; Δ is a measure of the zonally non-uniform component of the basic PV. An increase in Δ would increase both the local baroclinic and barotropic shears. A reasonable estimate for the increase of Δ from early-winter to mid-winter would be from 2.5 to 5.5. It is found that the growth rate of the most unstable normal mode decreases by about 50% from ~0.16 to ~0.09 as a result of the strengthening barotropic-governor effect (Fig. 3).

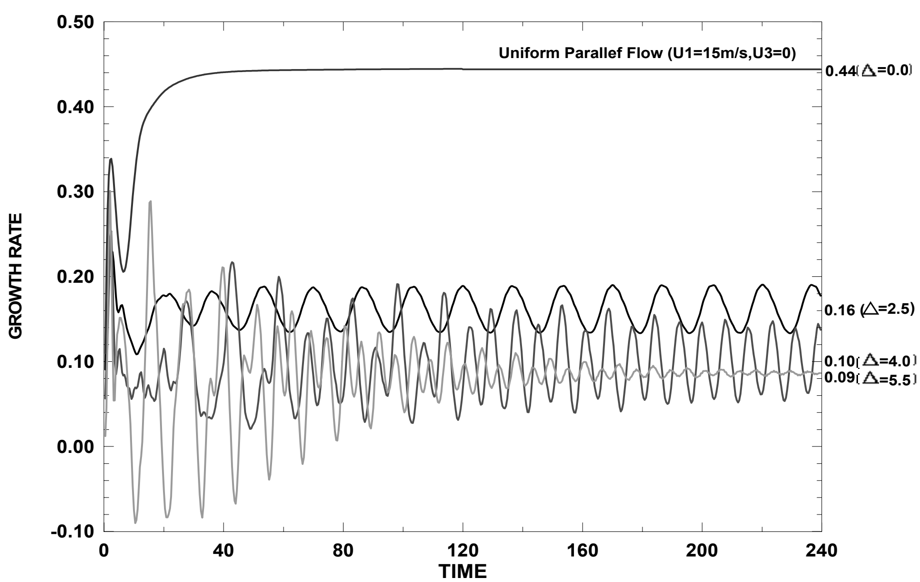
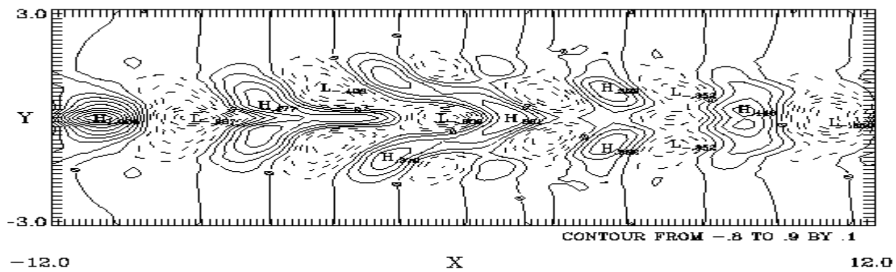
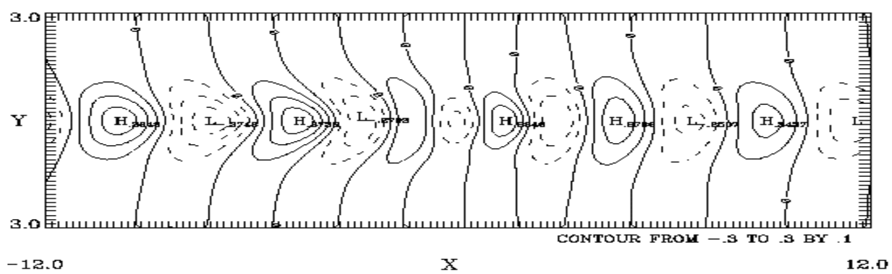


Fig. 3 Instantaneous growth rate as a function of non-dimensional time
for different values of Δ.

The most unstable mode for $U_T = 1.5, \Delta = 5.5$ has downstream localization of eddy activity on both levels (Fig. 4) with a clear westward vertical tilt associated with baroclinic growth. The horizontal tilt of the wave-packet is such that some kinetic energy is transferred from the disturbance to the basic flow. The downstream localization of the disturbance is also manifested in a downstream shift of the maximum local baroclinic conversion rate $C(P', K')$ relative to the jet (Fig. 5). We call this process "baroclinic localization".



(a)



(b)

Fig. 4 Normalized non-dimensional perturbation streamfunction for $U_T = 1.5, \Delta = 5.5$.

(a) Upper level ψ_1 (b) Lower level ψ_3.

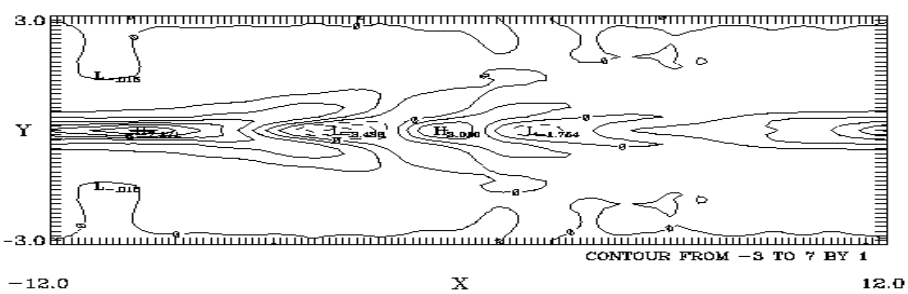


Fig. 5 Baroclinic energy conversion rate, $C(P', K')$, for $U_T = 1.5, \Delta = 5.5$.

For a fixed U_T, a sufficiently large Δ would make the jet and horizontal shear so localized that downstream barotropic energy conversion process dominates over baroclinic conversion process. We call this "barotropic localization". A series of computation using different combinations of U_T and Δ reveals the existence of three regimes distinguished in terms of the ratio of area-integrated barotropic conversion rate to baroclinic conversion rate, $\alpha = \langle \vec{E} \bullet \vec{D} \rangle / \langle C(P', K') \rangle$. Regime I is characterized by $\alpha < 0$ in which the instability is due to baroclinic process partly counteracted by barotropic-governor effect. Regime II is characterized by $\alpha > 0$ in which both baroclinic and barotropic processes contribute to the instability. Regime III is characterized by $\alpha < 0$, in which the instability is due to barotropic process partly counteracted by "baroclinic-governor" effect. The circle in Fig.6 indicates the range of parameter conditions that is relevant to the mid-winter background flow over North Pacific. In light of the results, we conclude that the change in the storm track intensity from early-winter to mid-winter may be interpreted as a consequence of a significant increase in the value of Δ for a given U_T over North Pacific.

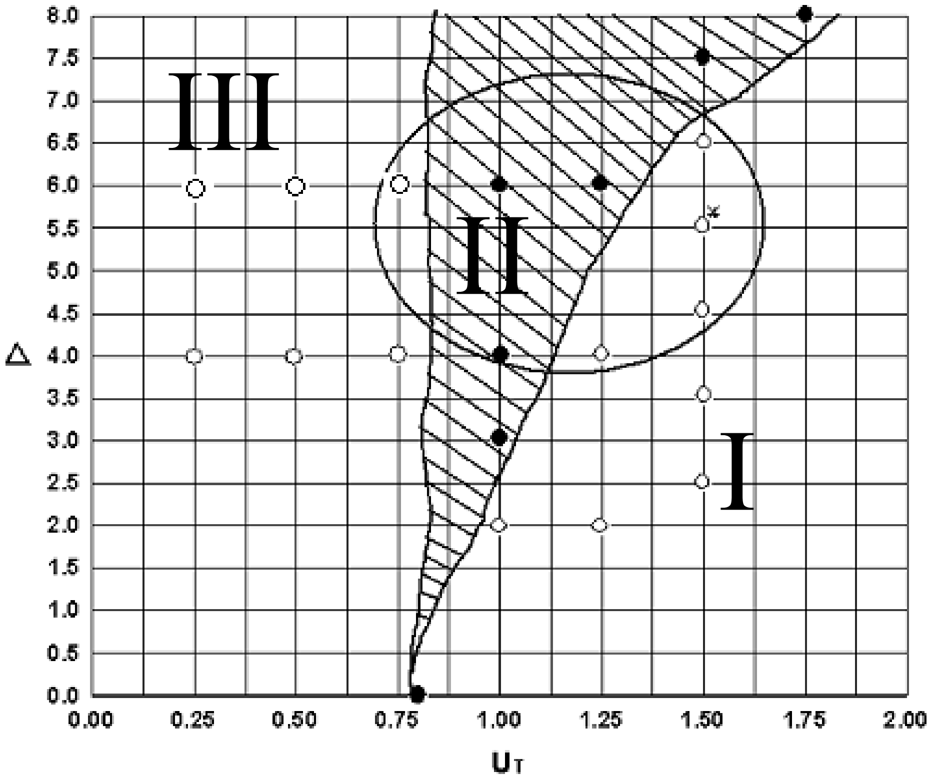


Fig. 6 Linear instability regime diagram on the basis of parameter α , where $\alpha = \langle \vec{E} \bullet \vec{D} \rangle / \langle C(P', K') \rangle$, Shaded area has positive values of α . Computations have been made at those parameter points indicated by ● and ○. The results shown in Figs. 4 and 5 are for the parameter condition indicated by a "star".

The dynamical nature of local baroclinic/barotropic instability is most clearly interpretable with the use of wave-guides in the basic state by invoking the concept of general wave-packet resonance. It emphasizes the mutual reinforcement among different constituent wave-packets through their advection of basic PV.