On the Significance of the Sensible Heat Supply from the Ocean in the Maintenance of the Mean Baroclinicity along Storm Tracks

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ABSTRACT

The relative importance between the sensible heat supply from the ocean and latent heating is assessed for the maintenance of near-surface mean baroclinicity in the major storm-track regions, by analyzing steady linear responses of a planetary wave model to individual components of zonally asymmetric thermal forcing taken from a global reanalysis dataset. The model experiments carried out separately for the North Atlantic, North Pacific, and south Indian Oceans indicate that distinct local maxima of near-surface baroclinicity observed along the storm tracks can be reinforced most efficiently as a response to the near-surface sensible heating. The result suggests the particular importance of the differential sensible heat supply from the ocean across an oceanic frontal zone for the efficient restoration of surface baroclinicity, which acts against the relaxing effect by poleward eddy heat transport, setting up conditions favorable for the recurrent development of transient eddies to anchor a storm track. Unlike what has been suggested, the corresponding reinforcement of the near-surface baroclinicity along a storm track as the response to the latent heating due either to cumulus convection or large-scale condensation is found less efficient. As is well known, poleward eddy heat flux convergence acts as the primary contributor to the reinforcement of the surface westerlies, especially in the core of a storm track. In its exit region, a substantial contribution to the reinforcement arises also from a planetary wave response to the sensible heat supply from the ocean. In contrast, the surface wind acceleration as a planetary wave response to the latent heating is found to contribute negatively to the maintenance of the surface westerlies along any of the major storm tracks.

1. Introduction

Day-to-day weather variations in the extratropics are mainly caused by synoptic-scale high and low pressure systems migrating eastward. These transient eddies are important in our climate system because they systematically transport sensible heat, westerly angular momentum, and moisture from the subtropics. Fluctuations associated with those transient eddies that are typically on a time scale of several days or a week can be extracted with appropriate high-pass temporal filtering. The activity of a transient eddy can be measured as the variance in geopotential height or meridional wind velocity, or as a poleward heat flux based on their filtered time series. In both the Northern and Southern Hemispheres (NH and SH, respectively), distinct maxima of measured eddy activity are observed in longitudinally confined regions above the midlatitude oceans (Figs. 1a,c): over the North Atlantic and over the North Pacific (Blackmon et al. 1977; Hoskins and Valdes 1990, hereafter HV90; Nakamura 1992), and from the central South Atlantic to the south Indian Ocean in the SH (Trenberth 1991; Nakamura and Shimpo 2004). These regions of concentrated transient eddy activity or “storm activity” are called “storm tracks.”

Transient eddies are regarded as a manifestation of baroclinically growing wave disturbances in a vertically sheared westerly flow. Charney’s (1947) and Eady’s (1949) linear theory of baroclinic instability beautifully explains the structure, wavelength, frequency, and growth rate of the observed transient eddies. In the theory, the
The maximum growth rate of the most unstable disturbances is given by mean-flow parameters, including the meridional temperature gradient (equivalently, vertical shear of the westerlies via thermal wind balance) and the inverse of static stability. The product of these two quantities has been used as a measure of the mean-flow baroclinicity [see Eq. (3)]. Indeed, near-surface baroclinicity evaluated locally for the time-mean flow is particularly high in the storm-track regions, especially in their upstream portions, both over the NH and SH (Nakamura et al. 2004; Nakamura and Shimpo 2004), where the underlying sea surface temperature (SST) exhibits tight gradients along oceanic frontal zones (Figs. 1b,d). Because individual transient eddies grow in a baroclinic zone (i.e., a region of high baroclinicity) by converting mean available potential energy (APE) to eddy kinetic energy (KE), they systematically transport heat poleward, acting to reduce the background baroclinicity. Then, their subsequent development would be unlikely to occur in that region where the baroclinicity has been reduced. In this regard, storm tracks can potentially be self-destructive in nature. Therefore, for the storm tracks to be maintained there must be certain mechanisms operative that can effectively restore the baroclinicity against the relaxing effect by eddy heat transport.

Traditionally, differential radiative heating between high and low latitudes has been considered to be the primary restoration mechanism of atmospheric baroclinicity (e.g., Stone 1978); however, it cannot explain the existence of such meridionally narrow baroclinic zones, as observed near the surface along the storm tracks. HV90 was the first to address this issue. Using a planetary wave model linearized about the observed wintertime zonal-mean flow with each of the transient eddy momentum and heat fluxes and diabatic heating imposed in the NH storm-track regions—namely, the North Atlantic and the North Pacific—as forcing, they investigated whether a steady response to each of the forcings can act to reinforce the mean baroclinicity in each of the storm-track regions. In the North Atlantic and the North Pacific, the effect of poleward eddy heat flux that acts to reduce the mean-flow baroclinicity tends to be offset by a restoring effect by diabatic heating (see Figs. 4d, 6d, 7d in HV90). In HV90’s experiments, however, it is not possible to evaluate the relative importance among latent, sensible, and radiative heating because at that time they had to diagnose the diabatic heating as the residual of the thermodynamic equation. Nevertheless, they emphasized the particular importance of latent heat release in the free troposphere associated with precipitation. They argued that, in a weather system, a latent heat release that occurs in the warm sector acts to enhance cyclone development by generating APE, while a heat exchange with the underlying ocean via surface sensible heat flux acts to dampen near-surface temperature fluctuations.

Some of the previous studies, including HV90, focused on mean-flow baroclinicity at the steering level of synoptic-scale disturbances; however, it is the temperature gradient at the surface that is more essential for the linear theory of baroclinic instability and for cyclone development in more realistic circumstances from a viewpoint of potential vorticity (PV) thinking (Hoskins et al. 1985). Indeed, several studies have suggested the importance of surface baroclinicity for energizing transient eddies and thereby anchoring storm tracks. For example, Nakamura and Sampe (2002) studied the mechanisms of midwinter suppression of baroclinic eddy activity along the North Pacific storm track (Nakamura 1992), a puzzling phenomenon that cannot be explained by the linear theory of baroclinic instability. Their statistical analysis of observational data has revealed that
a surface baroclinic zone anchored by a sharp SST gradient across an oceanic frontal zone along the Oyashio Extension [the subarctic frontal zone (SAFZ)] acts to anchor the storm track along it, while the strong subtropical jet acts to trap upper-level synoptic eddies into its core and thereby retard their efficient coupling with the surface baroclinic zone. Nakamura and Shimpo (2004) found that a surface baroclinic zone anchored by the distinct SST frontal zone between the Antarctic Circumpolar Current (ACC) and the Agulhas Current in the south Indian Ocean [Antarctic polar frontal zone (APFZ)] coincides with the core region of the SH storm track throughout the year. A numerical study by Inatsu and Hoskins (2004) suggests that a midlatitude SST distribution is most responsible for the zonal asymmetries observed in the lower-tropospheric wintertime SH storm track, including its distinct core region over the south Indian Ocean. Through a set of idealized numerical model experiments, where zonally uniform SST distributions are assigned as the model lower-boundary condition (i.e., “aqua-planet experiments”), Brayshaw et al. (2008) have confirmed the high sensitivity of storm-track activity to a midlatitude SST gradient. With experiments using a regional climate model with a relatively large spatial domain, Woollings et al. (2009) found that imposing higher-resolution SSTs, which can represent its sharp frontal structure, can improve the model representation of the North Atlantic storm track by locating it closer to the SST front.

Swanson and Pierrehumbert (1997) pointed out that, unlike in the upper troposphere, air parcels in the lower troposphere that move meridionally in association with transient eddies developing along the Pacific storm track must be subject to strong thermal damping through heat exchanges with the underlying ocean. They also pointed out that the damping is so efficient that it prohibits eddy temperature mixing in the lower troposphere and thus the relaxation of the temperature gradient. Nakamura et al. (2004) argued the air–sea heat exchange in the vicinity of an oceanic front may be of particular importance in maintaining baroclinicity along a storm track to maintain eddy activity. Aquaplanet numerical experiments by Nakamura et al. (2008) and Sampe et al. (2010) have indicated that the differential sensible heat supply across an oceanic frontal zone can effectively restore a surface air temperature (SAT) gradient. With this “oceanic baroclinic adjustment,” a surface baroclinic zone is maintained for the recurrent development of disturbances, acting to anchor a storm track and an eddy-driven polar-front jet (PFJ) around the SST front. In fact, high-resolution model experiments by Nonaka et al. (2009) and Taguchi et al. (2009) have elucidated how effectively the sharp SAT gradients across the ocean fronts can be restored by the near-surface sensible heating and cooling on their warmer and cooler sides, respectively, induced by cross-frontal cold- and warm-air advection associated with the passage of weather systems.

Nevertheless, one may still wonder whether the near-surface sensible heating is indeed more efficient in maintaining the surface baroclinic zones than other kinds of diabatic heating, including latent heating. In this study, we address this issue by assessing the relative importance of sensible and latent heating for the maintenance of the surface baroclinic zones observed near the storm-track cores in the wintertime NH and SH. We conduct numerical experiments with almost the same configuration as those in HV90 but impose the individual components of diabatic heating separately in a planetary wave model. Adopting virtually the same configuration as in HV90 allows us a straightforward comparison with their results. Unlike HV90, who evaluated mean-flow baroclinicity mainly for a single layer near the steering level of synoptic-scale disturbances, our evaluation is made for the entire troposphere with special attention to the near-surface layer, in recognition of the importance of surface baroclinicity in the storm-track formation, as suggested in the aforementioned recent studies. In addition, the response of the planetary wave model to each of the diabatic heating components is also used for assessing their relative importance in maintaining the mean surface westerlies. They are known to be important in midlatitudes for driving the ocean circulation, forcing heat and moisture releases from the ocean, and influencing the surface climatic conditions. Details of these experiments and their results are described in sections 2 and 3–5, respectively. A discussion and concluding remarks are presented in section 6, in which a new schematic view is proposed on the role of sensible and latent heat supply from the ocean in the maintenance of storm tracks.

2. Data and experimental settings

a. Data

The global circulation data and various kinds of diabatic heating rates used in this study are based on the National Centers for Environmental Prediction (NCEP)/Department of Energy (DOE) Global Reanalysis 2 (R-2) for the period from 1979 to 1998 (Kanamitsu et al. 2002), which is an improved version of the widely used NCEP–National Center for Atmospheric Research (NCEP–NCAR) Global Reanalysis (R-1) dataset (Kalnay et al. 1996). The improvements are especially in the representation of physical processes and thus in the
quality of the diabatic heating fields. Compared to the earlier periods, the inclusion of satellite-derived information has led to a significant improvement in the data quality, especially over the NH oceans and the entire SH, where not many station or sonde observations are available. A digital high-pass filter with a half-power cutoff period of 8 days was applied to the data time series on each grid point to extract fluctuations associated with transient eddies. The high-pass-filtered data were then used to evaluate the eddy thermal and vorticity forcing on the background flow.

Diabatic heating used as the forcing for the planetary wave model as described below includes turbulent sensible heat flux at the surface, which is represented in the reanalysis as subgrid-scale vertical turbulent temperature diffusion within the planetary boundary layer (SENS), radiative heating/cooling (RAD),1 and latent heat release associated with convective condensation (CONV) and large-scale condensation (LSC). The total diabatic heating (ALL) is defined as the sum of the four heating components mentioned above. Note that it does not include contributions from eddy thermal or vorticity forcing. The heating was taken from the R-2 data available as the rate of a local temperature change at 28 sigma levels and then interpolated onto the model grid points and vertical levels. Strictly speaking, the heating components CONV and LSC may not purely represent the direct effect of latent heat release. CONV includes the effects of vertical redistribution of sensible heat associated with subgrid convection and cooling associated with partial evaporation of raindrops or with the melting of snow. LSC includes the effect of cooling associated with cloud particle phase change. We, nevertheless, consider that it makes sense to include all those complexities into the effect of latent heat release, since they are, at least in the model for the reanalysis and perhaps also in the real atmosphere, always accompanied by latent heat release associated with precipitation. We have to keep it in mind that, though improved from its earlier version, the data quality of the R-2 dataset, especially parameterized physical processes and associated diabatic heating fields, is rather sensitive to the model representation of physical processes involved in the heating and therefore still subject to some uncertainties. Bauer and Del Genio (2006), for example, pointed out some deficiencies in midlatitude cloud representation in most of the climate and forecast models. Nevertheless, the decomposed heating components in the R-2 data are indispensable for our assessment. We have confirmed that the corresponding diabatic heating fields based on the R-1 dataset yield qualitatively the same results as shown below based on the R-2 data.

b. Model description

The steady linear response to a given tropospheric diabatic heating or to a thermal or vorticity forcing associated with transient eddies is investigated using a global time-dependent nonlinear primitive equation model with simplified physical processes, originally developed at the University of Reading (Hoskins and Simmons 1975). In our experiments the model resolution is set to be a triangular spectral truncation at wavenumber 42 (T42), with 15 equally spaced vertical sigma levels. As in HV90, topography is not included. Damping terms are incorporated in the form of Rayleigh drag and Newtonian cooling. A biharmonic hyperdiffusion is also included to stabilize the model integrations and to suppress the growth of transients. At the lowest and adjacent levels, time scales of the drag are set to be 1.0 and 5.0 days, and the time scales of the cooling are 5.0 and 10.0 days, respectively. In the free atmosphere, no drag is imposed and the cooling time scale is set to be 25.0 days. At the three highest levels placed in the model stratosphere, the drag time scales are set to be less than 1 day to avoid spurious reflections of waves at the top boundary.

c. Model integrations

Instead of constructing a linear version of the model, we obtained a steady “linear” response of the nonlinear model, following Held et al. (2002). The specific procedures are as follows:

1) From the initial state with complete zonal symmetry, the model was integrated with a prescribed forcing that had been taken from the R-2 data but scaled with a sufficiently small number $\varepsilon$ ($\ll 1$). In each of the integrations, the zonal-mean state in the model was fixed to the corresponding observed climatology, and a quasi-steady state was realized within ~20 model days.

2) In each of the integrations, the steady response to the weakened forcing was defined as the 10-day mean state, from the 16th to the 25th day of the integration.

3) Finally, the approximate linear steady response was obtained by rescaling the zonally asymmetric component of the steady response with the factor $1/\varepsilon$.

In each of the integrations, $\varepsilon$ was set to 0.01. This value is small enough to suppress nonlinearity in the response, since we have confirmed that replacing the value of $\varepsilon$ with 0.02, 0.01, or 0.005 yields almost the identical.

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1 Since the near-surface response to radiative heating was found to be rather weak, we do not present any figures of the response in this paper.
response. Although our model integrations do not reach a completely stationary state, the feedbacks from non-stationary components onto the background flow are negligible, because their amplitude is smaller than that of the stationary response by two orders of magnitude.

For each of the winter time North Atlantic, North Pacific, and south Indian Oceans, a linear steady response to each of the diabatic heating components and eddy forcing through their temperature or vorticity fluxes has been obtained through the method described above. The forcing prescribed for each of the model integrations was confined to the individual domain. Outside the domain, the forcing intensity was linearly reduced horizontally in such a way that it vanishes $5^\circ$ in latitude or longitude away from the domain boundaries to avoid any Gibbs phenomenon. For our experiments for the NH winter, the climatologies for the December–February period were prepared for both the zonal-mean basic state and the zonally asymmetric forcing fields, whereas their June–August climatologies were used for the SH winter.

d. Analysis procedure

As in HV90, the response field obtained was used to evaluate tropospheric baroclinicity, measured as the growth rate of the most unstable normal mode for the Eady (1949) problem in the log-pressure coordinate:

$$\sigma_{BI} = 0.31 \frac{f}{N} \left| \frac{\partial U}{\partial z} \right| \text{ with } (1)$$

$$N^2 = g \alpha \partial \theta / \partial z, \quad \alpha = T/T_{z=0} \cdot \theta, \quad (2)$$

where $f$ denotes the Coriolis parameter, $N$ the buoyancy (Brunt–Väisälä) frequency, $U$ the horizontal wind velocity vector, and $\theta$ potential temperature. In evaluating $N$ on a given level, $\theta$ was interpolated vertically with 1-km intervals in the log-pressure coordinate, and then $\partial \theta / \partial z$ was evaluated using centered finite differencing, where $z = H \ln(p_{00}/p)$ with the scale height $H (=10 \text{ km})$ and the reference pressure $p_{00} = 1000 \text{ (hPa)}$. For the lowest level ($z = 0, p = 1000 \text{ hPa}$), $\partial \theta / \partial z$ was evaluated using the first-order forward differencing of $\theta$ between that level and the level immediately above (i.e., 1 km above the surface). This rather nonstandard operation was applied because, as in HV90, in our model, which includes no representation of convective processes, unstable stratification may form in the linear responses, especially near the surface. As stated in the introduction, we evaluated the baroclinicity parameter $\sigma_{BI}$ for every tropospheric layer, rather than evaluating it for a single layer near the steering level as in HV90, with our special attention paid to the near-surface layers.

In addition, the Eady growth rate was also evaluated for the zonal-mean background state:

$$\langle \sigma_{BI} \rangle = 0.31 \frac{f}{(N)} \left| \frac{\partial U}{\partial z} \right| \text{ with } (3)$$

$$\langle N \rangle^2 = g \alpha \partial \theta / \partial z, \quad \alpha = [T]/([T]_{z=0} \theta), \quad (4)$$

where square brackets denote zonal averaging and any quantity with angle brackets is evaluated as a combination with the zonal means of the model’s basic variables. Finally, we evaluated the zonally asymmetric component of the growth rate, defined as the local departure of the model response $\sigma_{BI}$ from the zonally uniform basic state $\langle \sigma_{BI} \rangle$:

$$\sigma_{BI}^a = \sigma_{BI} - \langle \sigma_{BI} \rangle, \quad (5)$$

which can be considered as the response in baroclinicity and hereafter referred to as “baroclinicity response.” Note that both $\sigma_{BI}$ and $\sigma_{BI}^a$ are nonlinear functions of the model’s basic variables, including temperature and wind velocities. Hence, the zonal mean of $\sigma_{BI}$ is not necessarily equal to $\langle \sigma_{BI} \rangle$. More importantly, the baroclinicity responses ($\sigma_{BI}^a$) to the individual forcing components are not additive. In our actual evaluation, the model $\theta$ response was combined with $[\theta]$ to compute $\sigma_{BI}$ locally and then $\sigma_{BI}^a$ was calculated as the local deviation of $\sigma_{BI}$ from $\langle \sigma_{BI} \rangle$, which had been evaluated from $[\theta]$. In the following, we compare the baroclinicity responses based on the model experiments for the individual forcing components with that evaluated for the observed climatological state for the major storm tracks.

As mentioned in the introduction, we evaluate $\sigma_{BI}^a$ and $\sigma_{BI}$ over the entire troposphere with special attention to the near-surface layer. There are some critical arguments on the validity of near-surface baroclinicity as a measure of a background condition favorable for storm development, especially in the classical quasigeostrophic (QG) framework, where the thermal structure within the planetary boundary layer (PBL) cannot be represented explicitly. Many previous studies in the QG framework, therefore, regarded baroclinicity just above the PBL (e.g., 700–850-hPa levels) as an appropriate measure of baroclinic eddy growth. As described in the introduction, however, recent observational or GCM studies have revealed that storm-track activity tends to exhibit better correspondence with near-surface baroclinicity than at the bottom of the free troposphere (e.g., Nakamura and Sampe 2002; Nakamura and Shimpo 2004; Inatsu and Hoskins 2004; Brayshaw et al. 2008). In the present paper, the baroclinicity at the 925-hPa level, which is closely inspected in later sections, has been evaluated for a single layer between the 850- and 1000-hPa
FIG. 2. (a),(b) Distribution of ALL over the North Atlantic based on the NCEP–DOE reanalysis. (a) Horizontal distribution shown as column-averaged heating with contour intervals of 25 W m$^{-2}$. (b) Vertical structure shown as the rate of temperature change averaged between 80°W and 20°E with contour intervals of 0.25 K day$^{-1}$. Negative contours (cooling) are dashed. Zero contours are omitted. (c),(d) As in (a),(b), respectively, but for heating due to eddy heat flux convergence. (e),(f) Distribution of (subgrid scale) SENS over the North Atlantic based on the NCEP–DOE reanalysis. (e) Horizontal distribution shown as the rate of temperature change at the lowest sigma level ($\sigma = 0.9667$) with contour intervals of 1 K day$^{-1}$ [in place of column-averaged heating in (a),(c),(g),(i)]. Local maxima and minima are labeled with “+” and “−”, respectively. (f) As in (b), but for SENS with contour intervals of 0.5 K day$^{-1}$ (doubled from those in (b),(d),(h),(i)). (g),(h) As in (a),(b), respectively, but for heating solely due to LSC. (i),(j) As in (a),(b), respectively, but for heating solely due to CONV.
levels. Thus, the evaluated baroclinicity reflects the gross thermal structure nearly over the entire depth of the PBL. It can therefore be in reasonable correspondence with a theoretical treatment of the surface thermal condition in the QG framework proposed by Bretherton (1966), where the surface acts as an infinite source (δ function) of potential vorticity (cf. Hoskins et al. 1985).

In the subsequent sections, variables averaged longitudinally across each of the major storm tracks are illustrated in meridional sections. As stated in the introduction, a parameter as a measure of baroclinicity (i.e., $\sigma_{BI}$ or $\sigma_{BI}^*$) is more important for eddy development near the entrance of a storm track than at its exit, and is therefore averaged over the upstream portion of each of the NH storm tracks for being illustrated in a meridional section. Other parameters, including diabatic heating and eddy flux convergence, have been averaged across the entire zonal extent of a storm track. For the SH storm-track core over the south Indian Ocean, all the parameters, including baroclinicity, have been averaged over the entire zonal extent of the storm track. Our choice of the zonal extent for the averaging is rather subjective, but we
have confirmed that the results presented below are not sensitive to the particular choice.

e. Validity of our methodology

As stated in the introduction, our main interests are in distinguishing the individual roles of sensible and latent heating in the light of the recent finding that near-surface baroclinicity is important for anchoring the storm tracks. Before proceeding to our main assessments, we briefly show the results for the eddy thermal or vorticity flux forcing imposed in the North Atlantic sector (80°W–20°E). The purpose here is to verify the validity of our model simulation and methodology by assessing the consistency with previous studies, especially HV90.

Figures 2c,d show the horizontal and vertical structures, respectively, of the eddy heat flux forcing onto the background flow. A dipole of heating and cooling is found to the north and south of the storm-track axis, respectively, in the sense of reducing the mean baroclinicity. The node of the dipole is tilted slightly in the southwest–northeast direction, almost parallel to the storm-track axis. The dipolar heating–cooling exists in a deep layer with their peaks located around the 850-hPa (σ = 0.85) level. The steady linear response to this thermal forcing is presented in Fig. 3. In the upper troposphere (Fig. 3a), the response is anticyclonic and cyclonic on the poleward and equatorward sides, respectively, of the storm track around 45°N. The wavy pattern extends southeastward with a weaker anticyclonic response elongated zonally over North Africa and the Middle East. The response acts to weaken the upper-level westerlies along the storm-track axis. The lower-tropospheric response shown in Fig. 3b is opposite of the upper-level response, acting to enhance the low-level westerlies along the storm-track axis by inducing cyclonic circulation to the north and anticyclonic circulation to the south. The above-mentioned features are totally consistent with Lau and Holopainen (1984), who elucidated through their QG tendency method that transient eddy heat transport acts to reduce the mean-flow baroclinicity and to transport westerly momentum downward. Indeed, the baroclinicity parameter σBI evaluated with the inclusion of the response to the eddy heat flux forcing exhibits its minimum in the storm-track region (Figs. 3d–f). A region of enhanced baroclinicity is found in the subpolar North Atlantic around 55°N at the 925-hPa level (Fig. 3f), but it is well north of the storm track and much weaker than the responses to other types of forcing shown later.

Despite some differences in observational data used for the model basic state, model resolution, and procedures to obtain the linear response, the response obtained here is overall consistent with that obtained by HV90. A minor discrepancy is found in the wave-train-like response that tends to propagate further downstream than in their evaluation. Presumably, this is because of the different climatology used in the zonal-mean zonal wind between our evaluation and theirs. As pointed out by Ostermeier and Wallace (2003), the Northern Hemisphere annular mode (NAM) underwent a significant positive trend in the last quarter of the twentieth century. Thus, the zonal-mean westerlies in our climatology for 1979–98 must be stronger than the corresponding climatology for 1979–84 used in HV90.

The corresponding steady linear response to the eddy vorticity flux forcing, which is fundamentally barotropic, was also computed (not shown). In the entire depth of the troposphere, the response is in the sense of enhancing the westerlies along the storm track, which is again consistent with a finding by Lau and Holopainen (1984). The response in the near-surface temperature field is negligible, with no noticeable effect on the mean baroclinicity. This
result is in agreement with HV90, confirming the validity of our method for obtaining the steady linear response.

3. Results for the North Atlantic

In this section we discuss the results of our experiments, in which each of the forcing fields was restricted to the North Atlantic sector (20°–60°N, 80°W–20°E). Note that the longitudinal averaging performed exerts a slight smoothing effect, because of the southwest-northeast-tilted structure of the North Atlantic storm track, whose axis is observed at ∼45°N (Fig. 4a) along a surface baroclinic zone (Figs. 4b,c). It should be noted that the Eady growth rate in Figs. 4b,c, based on the reanalysis data, can be regarded as a response to all the possible forcing, including diabatic heating, topography, feedback from transient eddies, and quasi-stationary anomalies.

Figures 2a,b show the horizontal and vertical structures, respectively, of the total diabatic heating over the North Atlantic. As shown later, the midlatitude near-surface maximum of the heating is a manifestation of the sensible heat supply from the ocean. A pair of shallow maxima extending up to the 800-hPa level at 40° and 60°N is related to convective heating probably associated with the storm track and the partially resolved active polar lows, respectively. The upper-level cooling is radiative. Figure 5 illustrates the response to this total diabatic heating.

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2 A polar low is a mesoscale phenomenon with a horizontal scale of ∼1000 km or less and is not fully resolved in the R-2 data with a horizontal resolution of ∼100 km in the polar regions. Nevertheless, polar lows with relatively larger horizontal extent, which typically accompanies baroclinic instability, can be represented in the reanalysis.
over the North Atlantic. The low-level warm response downstream of the heating maxima (Fig. 5c) can be regarded as the direct response to the shallow near-surface heating (Kushnir et al. 2002). In the response, the baroclinicity is considerably enhanced locally in the storm-track region and in higher latitudes between 65° and 70°N (Figs. 5e,f). The distinct maximum in $\alpha_{BI}$ observed in the real atmosphere (Fig. 4c) is reproduced well in the model response, except near the surface (Fig. 5d). Again, the above-mentioned result is essentially the same as in HV90.

As shown in Figs. 2e,f (note the different convention in these panels from others), the sensible heating in the North Atlantic is very shallow and maximized along the Gulf Stream (GS), particularly off Cape Hatteras. Perhaps this strong heating is contributed to by outflows of very cold continental air to the warm GS associated with individual storms passing along the coastline. Through a set of GCM experiments with idealized SST distribution and landmass configurations, Brayshaw et al. (2009) argued that the strong sensible heating that contributes to the strong baroclinicity along the GS is a consequence of a pronounced land–sea thermal contrast, and that the heating is enhanced by the reversed triangular shape of the North American continent and by planetary waves.
forced by the Rockies. It should be noted that the pronounced meridional gradient in the climatological sensible heating across the GS off Cape Hatteras (Fig. 2e) is also contributed to by the oceanic cooling of the warm southerlies ahead of individual cyclones, which is particularly effective only to the north of the GS. This cooling tends to offset the contribution from the sensible heat release associated with cold-air outbreaks from the continent (Nonaka et al. 2009; Taguchi et al. 2009; Sampe et al. 2010).

Both in the upper and lower troposphere, the streamfunction response to the sensible heating shown in Figs. 6a,b is similar to that of the total diabatic heating (Figs. 5a,b), indicating the dominance of sensible heating. The near-surface temperature response in Fig. 6c also bears marked similarities with its counterpart to the total heating (Fig. 5c). Its dominance is also evident in the mean baroclinicity (Figs. 6d–f). The distinct near-surface maximum in $\sigma_{bb}$ forms near 45°N along the North Atlantic storm track (Fig. 6d), which is in good correspondence to the observations (Fig. 4c). The response to the sensible heating is confined to the layer below the 850-hPa level. These results suggest the primary importance of the sensible heat supply from the ocean in the maintenance of a surface baroclinic zone, in agreement with several of the latest studies (Nakamura et al. 2008; Taguchi et al. 2009; Nonaka et al. 2009; Sampe et al. 2010).

It should be pointed out that, although the strongest sensible heating occurs along the GS just off the U.S.
coast (Fig. 2e), the maximum positive responses at the 925-hPa level, both in temperature (Fig. 6c) and barocliniticy (Fig. 6f), are displaced downstream by ~30° in longitude of the heating maximum. This displacement may be attributable to the following factors. First, the steady linear response can be viewed as a stationary Rossby wave, whose eastward group velocity may contribute to the downward displacement. Second, as we mentioned in section 2d, the linear response, by nature, does not include vertical adjustment processes that would be activated in the real atmosphere if the stratification becomes unstable by the influence of the sensible heating, which may locally lead to an artificial enhancement of the barocliniticy response in the linear model. As evident in Fig. 5f, the latter problem can be alleviated when the sensible heating is combined with the latent heating that is maximized above the PBL and can therefore counteract against the destabilization effect of the former. Similar arguments can also be applied to our experiments for the North Pacific and south Indian Oceans discussed below.

Though weaker than the maximum response near the surface, there is another maximum of the baroclinicity response in the lower free troposphere (600–800 hPa) between 30° and 45°N (Fig. 6d). As indicated in Fig. 2f, diffusive cooling acting on the temperature inversion at the PBL top is confined to the shallow layer (below the 0.8-σ level) in the subtropics in the presence of subsidence associated with the Hadley cell, while the cooling is deeper in the midlatitudes. This latitudinal

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**Fig. 8.** As in Fig. 3, but for latent heating associated with CONV in the North Atlantic.
contrast in the diffusive cooling acts to maximize baroclinicity just above the PBL just the north of 30\°N. In reality, however, this effect is completely offset by strong longwave cooling in the cloud-free subsidence area in the subtropical lower troposphere, as evident even in the total diabatic heating in Fig. 2b. In fact, the baroclinicity response to the combined radiative and sensible heating (not shown) exhibits no maximum in the subtropical lower troposphere. The same interpretation can be applied to the corresponding subtropical maxima found in the baroclinicity response to the sensible heating in the storm tracks over the North Pacific and south Indian Oceans (Figs. 11c, 13c).

Figures 2g,h show that large-scale condensation heating over the North Atlantic is strongest around the storm track. Compared with the sensible heating, the particular latent heating is much deeper and meridionally broader, probably representing the contribution from precipitation ahead of warm fronts of synoptic-scale cyclones. The streamfunction response to large-scale condensation heating can enhance the mean-flow baroclinicity around the storm track, with enhanced and weakened westerlies at the upper and lower levels (Figs. 7a,b), respectively. However, the near-surface temperature response to large-scale condensation heating (Fig. 7c) is one order weaker than that to the sensible heating (Fig. 6c), and its effect on the near-surface baroclinicity is therefore quite limited, especially along the storm track (Figs. 7d–f).

Convective latent heating exhibits a deep maximum south of the storm track (Figs. 2i,j), in association with convection, probably in the warm sectors of individual cyclones traveling along the GS (Kuwano-Yoshida et al. 2010). In fact, the band of convective heating maxima (Fig. 2j) is located slightly south of the corresponding band of large-scale condensation heating (Fig. 2h). Convective heating also exhibits a shallower secondary maximum south of Iceland, perhaps because of the relatively shallow cumulus activity associated with polar lows. As shown in Fig. 8, the response to the convective heating is in the same sense as that to large-scale condensation heating. It reinforces the mean baroclinicity in the storm-track region (Fig. 8e) with enhanced westerlies at the upper levels, while reducing them at the lower levels (Figs. 8a,b). Though slightly stronger than that of large-scale condensation, however, the effect of convective heating on the maintenance of the near-surface baroclinicity along the storm track is also notably weaker (Fig. 8f) than that of the sensible heating, which is consistent with the weaker thermal response near the surface (Fig. 8c).

A comparison among the model responses to the individual forcing components mentioned above reveals the particular importance of sensible heating in maintaining strong near-surface baroclinicity along the storm track. To summarize the aforementioned results, contributions to $\sigma_{BI^*}$ from sensible heating (Fig. 6d) and from other diabatic heating components combined (Fig. 9) are compared in their distributions averaged zonally over the western North Atlantic (80\°–20\°W). It should be noted that, because of the nonlinearity described in section 2d, the sum of the responses shown in Figs. 6d, 9 does not coincide with that shown in Fig. 5d. As discussed earlier, the very strong near-surface baroclinicity response in Fig. 6d to sensible heating is likely overestimated in our linear model, and the corresponding response in Fig. 5d is modified because of the compensating effect on static stability by latent heating. Nevertheless, it is evident that the surface baroclinic zone, as observed along the storm track at 45\°N, cannot be reproduced without the contribution from sensible heating, while the contribution from other diabatic effects is noticeable only in the free troposphere above the 850-hPa level (Fig. 9).

4. Results for the North Pacific and south Indian Oceans

This section discusses the results of our planetary wave model experiments in which the thermal forcing was imposed separately over the North Pacific (20\°–60\°N, 110\°E–100\°W) and south Indian Oceans (20\°–60\°S, 0\°–120\°E). Since the essential features of these
results are the same as in the case for the North Atlantic, brief summaries are presented below for the two maritime regions.

As discussed by Nakamura et al. (2004), the relationship between the storm track and the mean baroclinicity observed in winter over the North Pacific is more complicated than over the North Atlantic because of the excessively strong subtropical jet (STJ) at ~30°N due to its merger with the PFJ. Nevertheless, the maximum eddy activity is observed at ~40°N along the strong SST gradient across the SAFZ along the Oyashio Extension (Figs. 1a, 10a). In addition to this surface baroclinic zone, the mean baroclinicity is also strong in the midtroposphere below the STJ (Figs. 10b,c). As argued by Nakamura and Sampe (2002) and Nakamura et al. (2004), the axial location and activity of the storm track tend to be determined through competing effects between the surface baroclinic zone and the trapping of upper-level eddies by the STJ.

Another notable difference between the North Pacific and the North Atlantic is in how sensible heat is released from the Sea of Japan, and then it receives a greater amount of latent heat when traveling over the Tsushima warm current. After traversing the Japanese Islands, it experiences further latent heating over the Kuroshio or its extension (Taguchi et al. 2009). Hence, unlike in the North Atlantic, the very cold continental air does not directly encounter the zone of strong meridional SST gradient over the Kuroshio–Oyashio Extension. Nevertheless, the continental air that has been warmed in traveling across the Sea of Japan is still cold enough relative to the SST to receive a substantial amount of sensible heat even over the warm side of the oceanic frontal zone (Taguchi et al. 2009). Thus, despite the differences in the land–sea configuration, particularly in the presence of the Japanese islands, the pronounced meridional gradient is generated in the sensible heating, especially to the east of Japan, as we observe in the North Atlantic.

In these surface baroclinic zones, the baroclinicity associated with the response to sensible heating is extremely strong (Fig. 12c), especially below the 900-hPa level, where not only the SAT gradient is tight but also the static stability is low under cold air masses coming out of the Asian continent over the relatively warm sea.
Fig. 11. As in Fig. 2, but for the North Pacific. Longitudinal averages in the right panels are taken over 110°E–180°.
surface. As in the case of the North Atlantic, the maximum response in 925-hPa temperature or baroclinicity (not shown) is located slightly downstream of the maximum sensible heating. Compared with this dominant effect of sensible heating, the corresponding effect of latent heating on the maintenance of the surface baroclinic zone around the Pacific storm track is substantially weaker. As shown in Fig. 12d, large-scale condensation heating, which is deep and strongest over the storm-track region, induces a baroclinicity response at the 925-hPa level that is too weak to maintain the surface baroclinicity efficiently. Likewise, the response to deep convective heating yields no noticeable enhancement of the surface baroclinicity along the storm track (Fig. 12e).

As shown by Nakamura and Shimpo (2004), storm-track activity over the south Indian Ocean is positioned close to a distinct surface baroclinic zone along the APFZ with a strong meridional SST gradient (Figs. 1c,d, 10d,f). Again, the shallow near-surface sensible heating that exhibits sharp meridional gradient across the APFZ (Fig. 13) is most efficient in maintaining the surface baroclinicity in the vicinity of the core region of the SH storm track (Fig. 14). Another maximum in near-surface baroclinicity in Fig. 14c located at ~60°S, which can also be seen in the observations (Fig. 10f), is due to a ice–sea contrast (Nakamura and Shimpo 2004). Neither large-scale condensation heating nor convective heating induces any significant thermal response that can contribute substantially to the maintenance of the surface baroclinic zone.

5. Responses in the near-surface westerlies

The main purpose of the above-mentioned experiments was to assess individual contributions from the latent and sensible heating components to the maintenance of near-surface baroclinicity. Still, it is worth discussing the corresponding responses in the surface westerlies, which are the main driving force of oceanic gyres and also influence SST distribution by controlling entrainment at the bottom of the oceanic mixed layer, Ekman advection, and air–sea heat exchanges via turbulent heat fluxes (e.g., Alexander et al. 2002; Kushnir et al. 2002; Kwon et al. 2010). Figure 15 shows the climatological distribution of the observed 925-hPa zonal wind, from which the zonally symmetric components have been removed for a straightforward comparison with the model responses shown in Figs. 16–18. In each of the ocean basins, the surface westerlies are strongest in a storm-track region, associated with a deep, eddy-driven polar-front jet (Nakamura et al. 2004).

An inspection of Fig. 16, which shows responses in the near-surface westerlies to the various forcing components imposed in the North Atlantic, reveals a significant contribution from the eddy heat flux forcing to the maintenance of the surface westerlies, especially in the core region of the storm track (Fig. 16b).
Fig. 13. As in Fig. 2, but for the south Indian Ocean. Longitudinal averages in (right) are taken over 0°–120°E.
result is consistent with diagnostic studies by Lau and Holopainen (1984) and Hoskins et al. (1983). Our experiment shows that sensible heating can also contribute comparably to the maintenance of surface westerlies (Fig. 16c), but its maximum is shifted slightly downstream relative to the response to the eddy heat flux. As argued by Palmer and Sun (1985), low-level convergence induced as a direct response to the shallow sensible heating acts as a source of positive vorticity, which is balanced by the negative planetary vorticity advection by the anticyclonic response generated downstream. Unlike the argument by HV90, the surface westerlies induced by the two latent heating components are located too far equatorward (Figs. 16d,e), and the induced wind is easterly in the storm-track region.

Essentially the same argument as for the North Atlantic can apply to the North Pacific and south Indian Oceans. In the North Pacific (Fig. 17), the eddy heat flux convergence can effectively induce the surface westerlies in the storm-track region, while the two latent heating components induce surface winds in the sense of decelerating the observed surface westerlies. The response to sensible heating can accelerate the surface westerlies in the storm-track region. Unlike the case of the North Atlantic, however, its maximum acceleration over the North Pacific is located somewhat downstream of the core region of the observed surface westerlies. The particular response is forced by radiative cooling localized off the Namib Desert (not shown), which seems to be due to marine stratiform clouds associated with

![Fig. 14. As in Fig. 12, but for the south Indian Ocean, based on longitudinal averaging for 0°–120°E.](image)

![Fig. 15. Climatological wintertime distribution of the zonally asymmetric component of 925-hPa westerlies (every 1 m s⁻¹; stippled for ≥4 m s⁻¹), based on the NCEP–DOE reanalysis, for (a) NH and (b) SH. Negative contours are omitted.](image)
a subtropical anticyclone (cf. Miyasaka and Nakamura 2010).4

The surface responses over western North Africa (Fig. 16a) and west of Mexico (Fig. 17a) to the total diabatic heating are also forced by low-level radiative cooling over the subtropical eastern North Atlantic and Pacific, respectively (not shown). Again, the cooling is perhaps due to marine stratiform clouds associated with the subtropical anticyclones. As indicated by Miyasaka and Nakamura (2005, 2010), radiative cooling due to marine stratus associated with each of the summertime subtropical anticyclones in the two hemispheres can reinforce them by yielding a Rossby wave response. Our experiments seem to suggest that, though weaker than in summer, low-level cloud radiating cooling associated with the subtropical anticyclones can also force the lower-tropospheric planetary waves; however, the zonal phase relationship between the thermal forcing and circulation response can differ from the summertime situation because of the stronger low-level westerlies in winter. This argument is, of course, beyond the scope of the present study.

Adiabatic arguments, including those by Hoskins et al. (1983) and Lau and Holopainen (1984), tell us that a storm track reinforces the mean surface westerlies primarily in its entrance and core regions via eddy heat transport. Our results suggest that if the effect of accompanied diabatic heating, especially sensible heating, is also taken into account, a storm track can effectively accelerate the mean surface westerlies across its entire length.

6. Discussion and concluding remarks

In this study, the importance of individual components of diabatic heating in maintaining the mean baroclinicity against the destructive effect of poleward eddy heat transport has been assessed in a “linear, steady” response of a planetary wave model to prescribed thermal

4 The surface responses over western North Africa (Fig. 16a) and west of Mexico (Fig. 17a) to the total diabatic heating are also forced by low-level radiative cooling over the subtropical eastern North Atlantic and Pacific, respectively (not shown). Again, the cooling is perhaps due to marine stratiform clouds associated with the subtropical anticyclones. As indicated by Miyasaka and Nakamura (2005, 2010), radiative cooling due to marine stratus associated with each of the summertime subtropical anticyclones in the two hemispheres can reinforce them by yielding a Rossby wave response. Our experiments seem to suggest that, though weaker than in summer, low-level cloud radiating cooling associated with the subtropical anticyclones can also force the lower-tropospheric planetary waves; however, the zonal phase relationship between the thermal forcing and circulation response can differ from the summertime situation because of the stronger low-level westerlies in winter. This argument is, of course, beyond the scope of the present study.
forcing based on a reanalysis dataset. Our primary interest is in assessing the relative importance of sensible and latent heating in the maintenance of near-surface baroclinicity along the major storm tracks, which is necessary to maintain them. The results of our experiments with the total diabatic heating are consistent with those of HV90 based on diabatic heating diagnosed as the residual of the thermodynamic equation.

Unlike in HV90, our model experiments with individual diabatic heating components taken from a reanalysis dataset allow us to assess the relative importance of a contribution from a particular component of heating among other contributions. The assessment suggests the primary importance of near-surface sensible heating in maintaining a surface baroclinic zone in the vicinity of a storm track. The sensible heating is shallow and arises from a surface turbulent sensible heat flux, with a sharp meridional contrast across the major oceanic frontal zones that is collocated with the major storm tracks over the North Atlantic, North Pacific, and south Indian Oceans. Our results are in contrast to the argument by HV90, who emphasized the primary importance of latent heat release in the free troposphere along the storm tracks in maintaining the baroclinicity along them. In fact, our experiments show that convective latent heating acts to enhance the baroclinicity in the free troposphere along the storm tracks. For the maintenance of the near-surface baroclinicity that is considered to be essential for cyclone development, however, the effect of latent heating is much weaker than the corresponding effect by sensible heating, primarily through the differential heat supply from the underlying ocean across the oceanic frontal zones. While sensible heating alone may not be able to reproduce the location of the surface baroclinic zone perfectly, our experiments have nevertheless revealed the primary importance of the sensible heating for the maintenance of the near-surface baroclinicity in each of the oceanic frontal zones, which cannot be replicated without that heating.

It has been considered that sensible heat exchanges with the ocean act as damping to transient eddies, especially their SAT fluctuations, as argued by HV90. Swanson and Pierrehumbert (1997) argue, however, that the damping must be necessary for maintaining the strong near-surface thermal gradient against the relaxing effect by the eddy heat transport. The present study further suggests that the sensible heat supply from the ocean, in turn, acts to maintain the observed storm tracks, in good agreement with Nakamura et al. (2004, 2008), Taguchi et al. (2009), Nonaka et al. (2009), and Sampe et al. (2010). The relaxing of the meridional gradient of SAT by poleward eddy heat transport enhances the air–sea temperature difference on the warmer side of the frontal zone and reduces (or even
reverses the sign of) the temperature difference on its cooler side. The subsequent augmentation of the cross-frontal differential sensible heat supply from the ocean surface acts to damp SAT fluctuations; however, at the same time, it efficiently restores the cross-frontal SAT gradient, setting up a condition favorable for the subsequent development of another weather system along the oceanic front. Nakamura et al. (2008) called this efficient restoration of surface baroclinicity “oceanic baroclinic adjustment.”

To gain further insight into the process of oceanic baroclinic adjustment, we have conducted a lag regression analysis for the wintertime south Indian Ocean (Fig. 19), based on 6-hourly data of the NCEP–DOE reanalysis and SST data (Reynolds et al. 2002). In this analysis, near-surface meteorological variables were linearly regressed on high-pass-filtered 850-hPa meridional wind velocity at a particular location in the APFZ, regarded as the reference time series. The timing of lag zero corresponds to the phase at which the equatorward wind velocity across the APFZ is maximized on passages of cold fronts associated with synoptic-scale cyclones (Fig. 19b). Since their period is typically ~4 days, lags of −1 and +1 day correspond to the peaks of cold- and warm-air advections, respectively, while lags of −1 and +1 day correspond to the passages of cyclone and anticyclone centers, respectively. Figure 19a depicts the typical time evolution of air–sea heat exchanges associated with migratory weather systems. Ahead of a cyclone, warm subtropical air advected poleward is cooled primarily on the cooler side of the SST front with an enhanced downward sensible heat flux. Behind the cyclone, the meridional wind reverses its direction, advecting cold subpolar air equatorward, which is then warmed rapidly, especially on the warmer side of the SST front, with the greatly enhanced sensible heat supply from the ocean. Poleward heat transport associated with individual eddies acts to relax the SAT gradient (Fig. 19c), which is counteracted by the differential sensible heat supply from the ocean across the SST front (Nakamura et al. 2004, 2008; Nonaka et al. 2009; Taguchi et al. 2009; Sampe et al. 2010). As a result, the relaxed SAT gradient by cold-air advection is restored rapidly within 1.5 days (Fig. 19e). In the case of the North Atlantic and the North Pacific, the above-mentioned scenario may require some modifications. Since subpolar air advected equatorward behind a cold front is coming out of the nearby continent, it is extremely cold in winter to induce substantial sensible heat release from the ocean even on

![Fig. 18. As in Fig. 16, but for the south Indian Ocean.](image)
the cooler side of an oceanic front. Therefore, it may take longer to restore the SAT gradient across the SST front.

Finally, we point out some implications of our study for scenarios on self-maintaining mechanisms of storm tracks in which oceanic processes are involved. HV90 postulated a positive feedback loop among a storm track, mean baroclinicity, latent heat release, and moisture supply from a warm western boundary current (WBC), as schematically summarized in Fig. 20a: (a) through eddy poleward heat and vorticity fluxes, surface westerlies are maintained along mid-latitude storm tracks; (b) The induced westerlies reinforce the warm WBC; (c) from which moisture supply to the atmosphere is enhanced; (d) Then latent heat release associated with precipitating storms is also enhanced; (e) inducing a steady planetary wave response in which mean baroclinicity along the storm track and the surface westerlies over the ocean basin are reinforced; and (f) the enhanced mean baroclinicity, in turn, maintains the storm-track activity. Finally, the process (a) follows again to form a positive feedback loop.

Recognizing the importance of near-surface baroclinicity and close association among a storm track, jet stream and oceanic front, Nakamura et al. (2004, 2008) proposed another positive feedback loop, which may be viewed as a modified version of the one by HV90. Its essence can be schematically summarized in Fig. 20b: (a) downward momentum transport associated with the poleward eddy heat flux maintains a deep westerly PFJ along a storm track that forms in the vicinity of an oceanic frontal zone; (b) the associated surface westerlies reinforce the WBC or ACC (c) to maintain the oceanic frontal zone perhaps via thermal advection; (d) then, the differential sensible heat supply from the ocean surface across the oceanic front restores the cross-frontal SAT gradient against the relaxing effect by the eddy heat flux (e) to maintain the storm-track activity. It should be emphasized that (f) the moisture supply through surface evaporation from the warm WBC plays an importance
role in the feedback loop proposed by Nakamura et al. (2004), as the latent heat release energizes individual storms whose SAT perturbations are subject to effective damping via heat exchanges with the underlying ocean. Moreover, as indicated in Fig. 20b with a dotted arrow, surface evaporation may have another effect, that of influencing the sensible heat supply through controlling the PBL depth (e.g., Troen and Mahrt 1986).

Our results in sections 3 and 4 suggest that the process (d) of Nakamura et al. (2004) is seemingly operative, while the process (e) of HV90 may be less effective. Our results in section 5 also support the process (a) of Nakamura et al. (2004, 2008), while questioning the effectiveness of the process (e) of HV90, that is, the ability of convective diabatic heating to accelerate the surface westerlies directly around a storm track. Rather, our results are overall supportive of the working hypothesis postulated by Nakamura et al. (2004, 2008). Furthermore, the sensible heat supply from a warm WBC reduces the static stability within the PBL to enhance the turbulent transport of westerly momentum down to the surface (Nonaka and Xie 2003; Minobe et al. 2008; Small et al. 2008; Booth et al. 2010).

It should be noted that the global reanalysis data currently available, including the NCEP–DOE reanalysis, are likely to underestimate the convective heating concentrated on narrow regions along intense WBCs, including the GS (Minobe et al. 2008) and the Kuroshio Extension (Tokinaga et al. 2009). Therefore, their contribution to the thermal forcing of planetary waves must be reevaluated in a future study. Furthermore, cumulus convections in a vertically sheared westerly flow act to enhance surface westerlies by transporting the westerly momentum downward (e.g., Schneider and Lindzen 1976; Kershaw and Gregory 1997), which can be operative in cyclones developing along a PFJ. This subgrid process, however, is not included in the forcing for our planetary wave model experiments; however, it may contribute to the maintenance of the surface westerlies along oceanic fronts. Still, it is not certain how effective these subgrid-scale processes acting locally along a warm WBC are in reinforcing the gyre circulation.

To put the feedback loop depicted in Fig. 20 on a firmer basis, more elaborative studies are required, in which all the processes involved in the extratropical air–sea interaction are treated explicitly, including oceanic processes in oceanic frontal zones, atmospheric processes in the planetary boundary layer, those processes involved in the maintenance of storm tracks, and the influence of the STJ that tends to trap eddies into its upper-level core. Since the hypothesis proposed by Nakamura et al. (2004) may be operative also under the zonally symmetric lower-boundary condition, especially in the determination of

![Fig. 20. Schematic pictures of the feedback loop postulated by (a) HV90 and (b) Nakamura et al. (2004, 2008). Refer to text for details.](image-url)
the latitude of the storm tracks or of the polar front jets, further effort with explicit treatment of the zonal-mean circulation is required as in Brayshaw et al. (2008), Nakamura et al. (2008), and Sampe et al. (2010).

The present study has examined how a climatological-mean near-surface baroclinic zone can be maintained in a particular longitudinal sector as a response of a zonally symmetric mean flow to zonally confined thermal forcing. The next step will be how variability in an oceanic frontal zone can influence the nearby storm track by modifying the strength and/or position of a near-surface baroclinic zone. For this purpose, it is necessary to use a stationary wave model with a zonally varying basic state or a stochastic “storm track model,” as in Peng and Whitaker (1999) and Watanabe and Kimoto (2000). Furthermore, the present study is based on a stationary response to time-mean heating. A more concrete picture of the restoration processes of a surface baroclinic zone via the sensible heat supply from the ocean related to individual cyclone systems requires a detailed analysis of time integrations of high-resolution numerical models, as in Taguchi et al. (2009) and Nonaka et al. (2009).

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