Midlatitude Storm Tracks, Jet Streams and Oceanic Fronts

by

Hisashi NAKAMURA,1,2* Takeaki SAMPE,1 Youichi TANIMOTO,2,3

and Akihiko SHIMPO4

1Department of Earth and Planetary Science, University of Tokyo, Tokyo, Japan
2Frontier Research System for Global Change, Yokohama, Japan
3Graduate School of Earth Environmental Science, Hokkaido University, Sapporo, Japan.
4Climate Prediction Division, Climate and Marine Department, Japan Meteorological Agency, Tokyo, Japan.

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1Corresponding Author: Hisashi Nakamura, Department of Earth and Planetary Science, Graduate School of Science, University of Tokyo, Tokyo, 113-8654 JAPAN.
e-mail: hisashi@eps.s.u-tokyo.ac.jp
1. Introduction

Synoptic-scale baroclinic eddies migrating along midlatitude storm tracks not only influence daily weather but also play a critical role in the climate system by systematically transporting heat, moisture and angular momentum. Seasonal variations of synoptic-scale disturbances have been examined for the Northern Hemisphere (NH; Petterssen 1956; Klein 1958; Sanders and Gyakum 1980; Roebber 1984, 1989; Whittaker and Horn 1984) and the Southern Hemisphere (SH; Sinclair 1994, 1995, 1996; Simmonds and Keay 2000), through tracking the centers of individual moving cyclones (or anticyclones) at the surface. This “Lagrangean-type” approach based on cyclone tracks (“storm tracks” in this particular framework) or anticyclone tracks is a straightforward application of weather chart analysis by synopticians.

In addition to this “synoptic” viewpoint, another approach has been adopted, where an emphasis is placed on propagation behavior of wavy disturbances and their feedback as ensemble to the time-mean flow in which they are embedded. This “Eulerian-type” approach is based on high-pass filtering of daily time series at individual grid points, to extract subweekly fluctuations associated with migratory synoptic-scale disturbances (Blackmon 1976; Blackmon et al. 1977, 1984; Wallace et al. 1988). In this “wave-dynamic” approach, regions of particularly large variance in geopotential height, meridional wind velocity and air temperature are called “storm tracks”, and “storm track activity” signifies the magnitude of their variance or poleward temperature flux. Wallace et al. (1988) discussed the relationship between cyclone or anticyclone tracks in the synoptic framework and storm tracks in the wave-dynamic viewpoint. An

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Wallace et al. (1988) argued that the poleward (equatorward) deflection of the cyclone (anticyclone) tracks relative to the storm tracks (or “baroclinic waveguides” in their terminology) to the downstream of the latter
advantage of this “wave-dynamic” viewpoint is that the local correlation between high-pass-filtered time series of meridional wind velocity and air temperature or vertical motion can be used as a measure of baroclinic structure of migratory disturbances. The high positive correlation indicates that disturbances are in baroclinic structure that allows efficient energy conversion from time-mean flow for their baroclinic growth. Climatological seasonal variations in storm track activity in this framework were documented comprehensively by Trenberth (1991) for the SH and by Nakamura (1992) for the NH.

As an extension of this wave-dynamic viewpoint\textsuperscript{2}, studies during the last decade have substantiated the concept of downstream development, on the basis of the observational evidence that synoptic-scale baroclinic eddies do exhibit group-velocity propagation along

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 can be interpreted as a superposition of cyclonic (anticyclonic) eddies upon the mean surface pressure field with a deep, persistent low (e.g., the Icelandic or Aleutian Low in the NH) in higher latitudes and a subtropical high-pressure belt in lower latitudes. In addition to their argument, the deflection can be augmented by the westerly acceleration near the surface as a consequence of downward transport of mean-flow westerly momentum through poleward eddy heat transport along the storm tracks. Critical arguments on the wave-dynamic definition of storm track have been made by Held (1998) and Bosart (1998). With this definition, it is difficult to treat cyclonic and anticyclonic eddies separately, which is its obvious drawback. Yet, most of the results presented in this section are based on that definition, since our primary concern is ensemble behavior of synoptic-scale transient disturbances.
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\textsuperscript{2} Blackmon (1976), Blackmon et al. (1977, 1984) and Wallace et al. (1988) used a band-pass filter that extracts fluctuations with periods between 2.5 and 6 days. Since only a small amount of energy is contained in fluctuations with periods shorter than 2.5 days, this filter eventually works as a high-pass filter. Chang (1993) showed that the frequency window of that band-pass filter is too narrow to depict wave-packet behavior of baroclinic disturbances. In order to moderate the artificial filtering effect, statistics based on 8-day high-pass-filtered quantities have been used for the results presented in this article and its related works (Nakamura et al. 2002; Nakamura and Sampe 2002; Nakamura and Shimpo 2003). Chang (2003) and others used lightly high-pass-filtered data (e.g., daily departures from monthly averages) to emphasize the downstream development.
storm tracks in the form of wavepackets (Orlanski and Kazfey 1991; Lee and Held 1993; Chang 1993; Swanson and Pierrehumbert 1994; Orlanski and Chang 1995; Berbery and Vera 1996; Rao et al. 2002). Global aspects of wavepacket-like behavior of those eddies were summarized in Chang and Yu (1999) and Chang (1999), with a particular emphasis on their group-velocity propagation. A thorough review of recent progress on the storm track dynamics can be found in Chang et al. (2002).

A number of synoptic analyses and related forecast experiments have shown the importance of heat and moisture supply from the warm ocean surface of the Kuroshio and Gulf Stream in individual events of explosive cyclone development (e.g., Nuss and Kamikawa 1990; Kuo et al. 1991; Reed et al. 1993; Neiman and Shapiro 1993). Recently, Xie et al. (2002) showed through their regional-model experiment that cyclone development is sensitive to a fine frontal structure in a sea-surface temperature (SST) field just to the north of the Kuroshio axis caused by the bathymetric effect of the shallow East China Sea in its wintertime cooling. An important climatological aspect revealed from the synoptic viewpoint is a close association between region of rapid cyclone development and major oceanic currents. For the SH, for example, Sinclair (1995) pointed out that surface cyclogenesis over the open ocean is most likely near the strongest meridional SST gradient associated with an intense oceanic frontal zone (OFZ) over the Indian Ocean. For the NH, rapid cyclone intensification is most likely along the Gulf Stream and Kuroshio (e.g., Sanders and Gyakum 1980; Roebber 1984).

The presence of huge landmasses in the extratropical NH confines surface baroclinic zones to offshore of the midlatitude western boundaries of the ocean basins, where in winter the warm western boundary currents encounter air flows from the cooled nearby continents.
Figure 1 shows the climatological midwinter situation over the Far East and North Pacific. The mean near-surface atmospheric circulation is characterized by the East Asian winter monsoon with the prevailing northwesterlies between the Siberian High and Aleutian Low (Fig. 1a). With this steady monsoonal flow, this region marks the strongest poleward heat transport over the NH. Over this region, where the cold, dry continental air encounters a warm air mass to the south, the low-level meridional temperature gradient is extremely high in midlatitudes (Fig. 1c), consistent with an extremely intensified westerly jet aloft (Fig. 1a). The tight temperature gradient and an abundant supply of heat and moisture from the warm ocean surface to the monsoonal air (e.g., Esbensen and Kushnir 1981; see also Fig. 1e) sustain intense lower-tropospheric baroclinicity (Hoskins and Valdes 1990; Nakamura 1992). As they migrate along this distinct baroclinic zone, baroclinic eddies develop to form a well-defined storm track downstream (Blackmon et al. 1977; Wallace et al. 1988; Fig. 1b), marked with a belt of local maxima in precipitation across the Pacific basin (Xie and Arkin 1997; Trenberth and Guillemot 1998; Fig. 1f). These migratory eddies thus supply fresh water to the ocean along the whole extent of the storm track, influencing the stratification in the midlatitude upper ocean (Lukas 2001). At the same time, they act to relax the temperature gradient by systematically transporting sensible heat to higher latitudes (Fig. 1b), while acting to accelerate the upper-level westerlies along the storm track (Fig. 1d), which is located to the north of the axis of an intense subtropical jet stream (STJ). This is an indication of eddy poleward transport of mean-flow westerly (angular) momentum from the STJ. The momentum thus transported into the upper-level storm track is transferred downward by means of poleward eddy heat fluxes (Fig. 1b), acting to sustain the surface westerlies (Lau and Holopainen 1984). In addition to this
feedback through their heat and momentum fluxes, baroclinic eddies as ensemble yield
diabatic heating associated with precipitation, acting to force the planetary wave pattern in the
mean flow (Hoskins and Valdes 1990).

The substantiation of downstream development of baroclinic wavepackets from the
wave-dynamic perspective requires us to interpret eddy statistics along storm tracks in relation
to cyclogenesis from the viewpoint of initial value problem. This type of cyclogenesis has been
known as the "B-type cyclogenesis" (Petterssen and Smeye 1971), to which further elucidation
was added by Hoskins et al. (1985) from a potential-vorticity (PV) perspective. In the
framework of "PV thinking", baroclinic growth of synoptic-scale disturbances can be interpreted as mutual reinforcement between PV anomalies at the tropopause level and those at
the surface in the form of temperature anomalies. In cyclogenesis associated with the down-
stream development, thermal anomalies are triggered by wind fluctuations across a surface
baroclinic zone induced by an incoming upper-level vortex. Thus, near-surface baroclinicity is
of particular significance in baroclinic instability (Bretherton 1966; Hoskins et al. 1985). Never-
theless, the significance was somehow overlooked in most of the studies from the wave-
dynamic perspective. Rather, they focused on baroclinicity defined for a lower-tropospheric
layer between the 700 and 850-mb levels, which roughly corresponds to the steering level of
baroclinic waves (Blackmon et al. 1984).

The main purpose of this article is to substantiate the role of near-surface baroclinicity in
anchoring storm tracks and influencing the seasonal cycle of their activity from the wave-

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3 Takayabu's (1991) "coupling development", in which a strong upper-level cold vortex triggers a surface
cyclone development, is akin to Pettersen's "B-type cyclogenesis".
dynamic viewpoint and the PV thinking, in analyzing observed data for the two Hemispheres. After discussing the importance of storm tracks in midlatitude coupled ocean-atmosphere variability, we will show a close association among a storm track, a polar-front jet (PFJ; or subpolar jet stream) and an OFZ. We will also show how this association is disturbed by the intensification of a STJ in a winter hemisphere.

2. Importance of storm tracks in extratropical coupled ocean-atmosphere variability

The interaction between the midlatitude ocean and storm tracks is by no means a new concept. The importance of storm tracks has been emphasized in the concept of the "atmospheric bridge", through which the effects of SST anomalies (SSTAs) in the Tropical Pacific, associated mainly with the El Niño/Southern Oscillation (ENSO), is transferred into midlatitudes to drive SSTAs underneath with the opposite polarity (Alexander 1992; Lau and Nath 1994, 1996, 2001; Lau 1997). This mechanism must be operative also in interdecadal SST variability over the North Pacific driven by the corresponding tropical variability (Nitta and Yamada 1989; Trenberth 1990; Graham 1992; Gu and Philander 1997; Zhang et al. 1997). A comprehensive review of this concept is given by Alexander et al. (2002), and a related review is given by Hoerling and Kumar (2002). Once a stationary atmospheric teleconnection pattern with equivalent barotropic anomalies forms in midlatitudes as a remote response to tropical SSTAs, local storm track activity and associated poleward heat fluxes are altered (Hoerling and Ting

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4 For further discussion on this subject, refer to a section by Alexander and Scott (2003) in this volume.

5 For further discussion on the Pacific (inter-) decadal variability, refer to sections by Schneider and Deser (2003), Latif (2003), and Seger et al. (2003) in this volume.
1994). It is this anomalous heat flux through which the anomalous upper-level westerly momentum associated with the teleconnection pattern is transferred down to the surface (Lau and Holopainen 1984; Lau and Nath 1991). The surface wind anomalies thus generated in midlatitudes drive SSTAs locally by changing the latent and sensible heat release from the surface, entrainment at the bottom of the oceanic mixed layer, and a cross-frontal Ekman current (Frankignoul and Reynolds 1983; Frankignoul 1985).

The ocean-atmosphere interaction involved in the “atmospheric bridge” mechanism is thus primarily one-way forcing by atmospheric anomalies on the upper ocean, consistent with the general one-way nature of the extratropical air-sea interaction (e.g., Frankignoul 1985; Frankignoul et al. 1998). Generally, midlatitude SSTAs generated through this mechanism are extending over a large horizontal extent, reflecting the spatial scale of the atmospheric anomalies that have forced them (Namias and Cayan 1981; Wallace and Jiang 1987). The one-way nature has also been suggested by the fact that most of the atmospheric general circulation models (AGCMs) fail to generate any systematic response to prescribed extratropical SSTAs, as reviewed by Kushnir et al. (2002). Yet, several experiments in each of which an AGCM is coupled thermally with a simple ocean mixed layer model (e.g., Lau and Nath 1996, 2001; Watanabe and Kimoto 2000) indicated that extratropical SSTAs act to reinforce overlying atmospheric anomalies that have driven the former anomalies, by modulating the spatial structure of the former anomalies and prolonging their intraseasonal persistence. This weak local feedback from SSTAs to atmospheric anomalies is called “reduced thermal damping” (Kushnir et al. 2002), as elucidated in a linearized one-dimensional coupled model by Barsugli and Battisti (1998).
When atmospheric anomalies drive midlatitude SSTAs, as in the “atmospheric bridge”
mechanism, local correlation between a SSTA and an anomalous upward turbulent flux must
be negative (Cayan 1992ab; Hanawa et al. 1995; Tanimoto et al. 1997; Seager et al. 2000),
and the SSTAs generally exhibit strong negative correlation with the overlying anomalies in the
surface wind speed. Recently, however, Nonaka and Xie (2003) demonstrated through their
analysis of satellite-measured SST and surface wind fields that the SST-wind correlation is
positive along the Kuroshio south of Japan and its eastward extension. Analyzing wintertime
shipboard measurements compiled on a high-resolution grid over the North Pacific, Tanimoto
et al. (2003; hereafter referred to as TNKY03) found that latent heat flux anomalies are
positively correlated with local SSTAs over the subarctic frontal zone (SAFZ; Yasuda et al.
1996; Yuan and Talley 1996) located in the Kuroshio-Oyashio Extension (KOE). They also
found that this positive correlation is stronger when the SSTAs lead the flux anomalies. Unlike
in the vast central domain of the basin, the ocean-atmosphere interaction cannot be inter-
preted solely with local exchanges of heat and momentum through the ocean surface in the
KOE region where the thermal advection by the strong western boundary currents contributes
substantially to the upper-ocean heat budget (Qiu and Kelly 1993; Qiu 2000, 2002). These
recent results suggest that the role of SSTAs formed along the western boundary currents or
in SAFZs in the air-sea interaction can be more than the “reduced thermal damping”. Confined
to a meridionally narrow region as the KOE, the signal of the oceanic forcing is hardly captured
in SST data compiled on a coarse resolution grid (with latitudinal intervals of, say, 5°) or
through a statistical technique that preferentially extracts basin-scale anomaly patterns as a
singular-value decomposition as used in Deser and Timlin (1997).
A close association has been found in the extratropics between decadal SST variability and oceanic frontal zones (Nakamura et al. 1997a; Nakamura and Yamagata 1999; Nakamura and Kazmin 2003). Schneider et al. (2002) examined in detail the mechanisms of the North-Pacific decadal variability simulated in a coupled atmosphere-ocean general circulation model in relation to its observational counterpart. They pointed out that the KOE is the key region for the oceanic feedback on the atmospheric anomalies. TNKY03 argued how the quasi-decadal SSTAs observed along the SAFZ in association with the decadal climate variability inherent to the North Pacific (Deser and Blackmon 1995; Nakamura et al. 1997a; Nakamura and Yamagata 1999; Xie et al. 2000; Tomita et al. 2001) can reinforce the associated stationary atmospheric anomalies (Figs. 2-5). In the presence of warm (cool) SSTAs from early to mid-winter (Figs. 2a-b), enhancement (reduction) of latent heat release from the surface occurs along the SAFZ (Fig. 3a). Linearization of anomalous local heat fluxes, as in Halliwell and Mayer (1996), has revealed that the enhanced (reduced) heat release over the SAFZ is attributed to the local SSTAs (Fig. 3b), part of which is offset by a contribution from surface air-temperature anomalies (Fig. 3c). Concurrently, the poleward (equatorward) displacement of the storm track activity occurs (Fig. 4b), probably in response to changes in the near-surface baroclinicity. In fact, the axis of the SAFZ tends to shift poleward (equatorward) during warm (cool) periods associated with the decadal variability (Nakamura and Kazmin 2003). In the upper troposphere, the anomalous storm track activity exerts anticyclonic (cyclonic) forcing over the midlatitude North Pacific through anomalous vorticity transport\(^6\) (Fig. 4c), acting to

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\(^6\) The feedback forcing shown in Fig. 4c was evaluated in the same manner as in Nakamura et al. (1997b).
reinforce the pre-existing stationary anticyclonic (cyclonic) anomalies. The anomalies resemble the Pacific/North American teleconnection pattern (Wallace and Gutzler 1981), regarded as a preferred mode of the atmospheric variability in the exit region of the North Pacific jet (Simmons et al. 1983; see also Peng and Robinson 2001). The anomalous storm track activity shown in Fig. 4b is typical for this particular phase of the PNA pattern (Lau 1988; Lau and Nath 1991). Consistent with this local feedback forcing from the storm track, a wave-activity flux of stationary Rossby waves associated with those anomalies (Takaya and Nakamura 2001) is strongly divergent out of the forcing region (Fig. 4a). As in the “atmospheric bridge”, the anomalous westerly momentum associated with the PNA-like pattern is transferred downward through anomalous storm track activity, leading to the reinforcement of the existing anomalous Aleutian Low (Fig. 5). The surface wind anomalies thus reinforced exert thermal forcing upon the upper ocean over the central and eastern North Pacific (Fig. 3d), acting to extend warm (cool) SSTAs toward the east of the SAFZ while forcing cool (warm) SSTAs off the west coast of Canada, consistent with the observed tendency in SSTAs toward late winter (Figs. 2b-c).

A similar mechanism has been postulated by Kushnir et al. (2002) as a paradigm for the coupling between a meridional dipole of atmospheric stationary anomalies and dipolar SSTAs as typically observed in the North Atlantic7. As in the mechanism of TNKY03, the critical factor in reinforcing the atmospheric anomalies is anomalous storm track activity in response to the anomalous surface baroclinicity associated with the dipolar SSTAs. Kushnir et al. (2002) considered a particular situation where those SSTAs have been generated locally by the

7 An overview of the coupled ocean-atmosphere variability in the Atlantic is given in this volume by Xie and Carton (2003).
atmospheric anomalies, as in the atmospheric bridge mechanism. Above the SSTAs along the Pacific SAFZ, in contrast, the observed surface wind anomalies are weak, particularly in the first half of winter (Fig. 5a; see also Fig. 3d), suggestive of a more active role of those SSTAs in the ocean-atmosphere interaction. In fact, it is recently suggested that interannual to decadal SST variations in the SAFZ or KOE are strongly influenced by changes in the oceanic conditions, including the gyre adjustment to the atmospheric forcing exerted far to the east (Xie et al. 2000; Seager et al. 2001; Schneider et al. 2002; Tomita et al. 2002). Once such zonally elongated SSTAs form in an OFZ through oceanic processes, they would immediately modify the near-surface baroclinicity in the overlying atmosphere.

The mechanism argued by TNKY03 is essentially the same as what Peng and Whitaker (1999) suggested through their careful diagnosis of an AGCM response to warm SSTAs prescribed along the Pacific SAFZ. Their diagnosis has revealed the critical importance of the involvement of a local storm track in generating the stationary atmospheric response with equivalent barotropic structure in the model. A similar suggestion was made by Watanabe and Kimoto (2000) for the generation of the stationary atmospheric anomalies over the North Atlantic in response to the local SSTAs. Peng et al. (1997) found that the AGCM response to the SSTAs prescribed in the North-Pacific SAFZ is sensitive to subtle differences in the model time-mean flow between January and February. The sensitivity stems from how effectively the remote barotropic response (the PNA-like pattern) is excited through the storm track feedback from near-surface anomalies as the robust direct response to the SSTAs (Peng and Whitaker 1999). Since the SSTA pattern prescribed as the model boundary condition was taken from the observation, a mismatch in their meridional positions can happen between the model
storm track and the direct thermal response to the local SSTAs in the SAFZ. The mismatch would be likely if the model climatology deviates substantially from its observational counterpart. The model sensitivity suggests the potential importance of the close association between the SAFZ and Pacific storm track in reinforcing the PNA-like anomalies, as a component of the feedback loop that is likely to be operative in the decadal-scale coupled ocean-atmosphere variability inherent to the North Pacific. We conjecture that horizontal and vertical resolutions of the AGCMs used in the previous attempts to simulate the response to SSTAs in frontal zones may not be high enough for fully resolving the meridionally confined anomalies in near-surface baroclinicity across OFZs and its association with nearby storm tracks.

3. Importance of storm tracks and ocean-atmosphere interaction in the extratropical Southern Hemisphere atmospheric circulation system

As discussed in the Introduction, a number of synoptic-oriented studies have suggested that the thermal influence from the warm ocean surface is an important factor for rapid development of individual cyclones. Nevertheless, the importance of the ocean-atmosphere interaction has been somehow overlooked until very recently in studies on storm track dynamics from the wave-dynamic perspective, where this subject has been regarded as a pure atmospheric dynamic issue. From the PV perspective (Hoskins et al. 1985), near-surface baroclinicity is of critical significance for baroclinic growth of atmospheric synoptic-scale disturbances. As the gradient of surface air temperature over the open ocean is closely linked to SST gradient underneath, surface baroclinic zones over the ocean are anchored along OFZs (Nakamura and Sampe 2002; Nakamura and Shimpo 2003, hereafter cited as NS03).
Over the SH, a close association between the major storm tracks and OFZs is evident both in winter and in summer (Figs. 6c and 6f). The stronger low-level storm track activity over the South Atlantic and Indian Oceans than over the South Pacific is in good correspondence with more intense SST gradient\(^8\) across the Antarctic Polar Frontal Zone (APFZ; Colling 2001) along the Antarctic Circumpolar Circulation (ACC) over the former oceans. Over any of the ocean basins, no well-defined storm track forms in the lower troposphere along a subtropical frontal zone (STFZ) located at \(\sim 30^\circ\)S. Both in the upper and lower troposphere (Fig. 6), the core region of the storm track is situated in the southwestern Indian Ocean (NS03), and that core almost coincides with the core of the APFZ. In fact, Sinclair (1995) demonstrated that the most frequent cyclogenesis in the SH occurs around this APFZ core. There, in the course of their seasonal march, the lower-tropospheric storm track activity exhibit high positive correlation with baroclinicity\(^9\) for a layer just above the surface (NS03), whose importance has previously been overlooked. The correlation is even higher than that with the baroclinicity defined for lower-tropospheric layers near the steering level of subweekly disturbances (NS03). The same is the case for the South Atlantic, where the storm track activity is also high (Figs. 6c

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\(^8\) In the following, we use satellite-measured monthly SST data compiled on a grid with 1\(^\circ\) resolution in both latitude and longitude (Reynolds and Smith 1994). The data were obtained from the web site of the Climate Diagnostics Center of the National Oceanic and Atmospheric Administration.

\(^9\) As in NS03, the baroclinicity has been locally evaluated as a non-dimensional number \(G = \left| \frac{g}{2 f_0} \right| \Delta \theta \partial / \partial N\), where \(N\) denotes the Brunt-Väisälä frequency, \(\theta\) potential temperature, \(g\) the acceleration of gravity, \(f\) the Coriolis parameter and \(f_0 = f(45^\circ\)S). In a situation where the thermal wind balance holds, \(G\) is equal to a quantity \(G^* = (f f_0) R_i^{-1/2}\), where \(R_i = N^2 A^2 + \partial^2 \partial z^2\) with horizontal wind velocity \((V)\). The maximum growth rate of the most unstable mode has been shown to be proportional to \(G\) (or \(G^*\)) in linear theories of baroclinic instability for the zonally uniform westerlies (Charney 1947; Eady 1949).
and 6f). The deep structure of the storm track over the Atlantic and Indian Oceans (Figs. 7d-e) reflects both the pronounced baroclinic growth of disturbances at the lower levels above the intense surface baroclinic zone and the well-defined upper-level PFJ (Figs. 7a-b). The jet acts as a good waveguide for baroclinic wave packets (NS03), recognized as strong eastward E-P flux along the upper-level storm track (Figs. 6a and 6d). Both the local baroclinic growth of disturbances in the lower troposphere and incoming wave activity as downstream development of the packets influence the upper-tropospheric eddy activity (NS03), which is maximized along the velocity core of the PFJ throughout the year.

As in the “atmospheric bridge” mechanism, poleward eddy heat fluxes across storm tracks transport westerly momentum downward, acting to locate belts of the strong surface westerlies in the proxies of the oceanic storm tracks (NS03). The storm track core is collocated with the core of the near-surface westerly jet (Figs. 6b-c and 6e-f), situated below the core region of the PFJ (Figs. 6a-b and 6d-e; NS03). The strong surface westerlies are a manifestation of the deep structure of the PFJ\(^\text{10}\) (Figs. 7a-b), as a consequence of the downward transport of the mean-flow westerly momentum due to eddy heat fluxes (Lau and Holopainen 1984; Lau and Nath 1991). The close association among the cores of the storm track, surface westerly jet and APFZ can be observed throughout the year (Figs. 6b-c and 6e-f; NS03). Consistently with the annual persistence of their association, the strongest annual-mean wind stress within the world ocean is observed in the SH storm core region (Trenberth et al. 1988),

\(^{10}\)In austral winter, the tropospheric PFJ is connected into the stratospheric polar-night jet (Randel and Newman 1998). This deep westerly jet acts as a waveguide for upward propagating Rossby wave trains. Subseasonal fluctuations in the lower-stratospheric wintertime circulation are associated largely with those wave trains originated from tropospheric anomalies (Nishii and Nakamura 2003).
which is suggestive of the importance of the storm track activity in driving the ACC and associated APFZ. A close inspection of Figs. 6b-c and 6e-f reveals that the APFZ tends to be displaced slightly equatorward of the surface westerly axis. This displacement is consistent with the tendency that the surface turbulent heat fluxes, wind stirring effect on the oceanic mixed layer, and cold Ekman advection are all maximized along the wind velocity axis.

In sharp contrast with the situations over the Atlantic and Indian Oceans, the storm track activity over the South Pacific is strongly influenced by the formation of the intense STJ over the Indian and Pacific Oceans (NS03). In the absence of the intense STJ during austral summer and autumn, the Pacific storm track is part of a well-defined circumpolar storm track along the PFJ along the ~50°S circle (Fig. 6d). During winter and spring (Fig. 6a), in contrast, upper-level storm track activity bifurcates from the core region into the main branch along the STJ and the secondary branch along the PFJ, in the presence of the distinct double-jet structure in the basic state (Figs. 6a-b; Karoly et al. 1998; Bals-Elsholz et al. 2001). In the lower troposphere, a storm track forms only along an enhanced low-level baroclinic zone below the PFJ (Fig. 6c). Unlike the PFJ, the STJ does not accompany strong westerlies near the surface. The STJ axis almost coincides with the axis of the surface subtropical high-pressure belt (Fig. 6b), particularly below the entrance and core regions of the jet. The intense velocity core of the STJ, which is confined around the tropopause (Fig. 7c), acts as an excellent waveguide for synoptic-scale eddies. In fact, the extended Eliassen-Palm (E-P) flux

\[ \dot{E} = \left[ \left( \nu^2 - \bar{u}^2 \right) / 2, -\bar{u} \bar{v}', f
\bar{v}' / (\partial \Theta / \partial p) \right]^T \cos \phi, \]

where an overbar signifies time averaging, a superscript T the vector transpose, \( \phi \) latitude, an \( \Theta \) the average of \( \theta \) over the entire extratropical SH. Note that a negative value of the vertical component in the \( p \)-coordinate means the
associated with subweekly disturbances is dominantly eastward along the STJ (Fig. 6a), indicative of the strong trapping of wave activity into the jet core (Fig. 7f). The trapping contributes to the dispersion of wave activity accumulated in the core of the upper-level storm track over the Indian Ocean mainly toward the STJ (Fig. 6a). Located over the subtropical high, the STJ does not favor the baroclinic eddy growth (NS03), despite the presence of modestly strong near-surface baroclinicity associated with the South Pacific STFZ.

The dominant influence of the STJ upon the upper-tropospheric storm track activity over the South Pacific can be illuminated further by contrasting the two winters during the unprecedented ENSO cycle in the 1997–98 period\(^{12}\) (Fig. 8). In good agreement with previous studies (Chen et al. 1996; Kiladis and Mo 1998), the double-jet structure became even more evident in the El Niño winter of 1997, with the profound intensification of the STJ especially over the Pacific (Figs. 8a and 8c). Contrastingly, the double-jet structure was diminished in the La Niña winter of 1998\(^{13}\), especially over the central and eastern domains of the South Pacific, in association with a dramatic weakening of the STJ (Figs. 8d and 8f). Under these drastic

\[\text{upward flux. This form of the flux differs slightly from the original definition by Hoskins et al. (1983), but it is more suited for representing the three-dimensional group velocity propagation of migratory eddies relative to the locally-defined basic flow in which they are embedded. Another dynamical significance of the flux is that it represents the propagation (relative to the basic flow) of wave activity pseudomomentum, the second-order easterly momentum of the mean flow carried conservatively with waves. In other words, baroclinic eddies act to transport mean-flow westerly momentum in the direction opposite to the E-P flux.}\]

\(^{12}\) With respect to the upper-tropospheric jet structure over the South Pacific, these two winters are such distinct outliers that they are worth examining in a case study.

\(^{13}\) Precisely speaking, the July-August period of 1998 was at the initiation stage of the La Niña event right after the termination of the 1997 El Niño event.
changes in the jet stream structure\textsuperscript{14}, the storm track activity was also changed dramatically. In the 1997 winter (Fig. 8a), the bifurcation of the upper-level storm track activity was evident with the main storm track branch extended consistently along the STJ, to which wave activity is dispersed from the storm track core not only over the eastern Indian Ocean but also over Australia (Fig. 8a). In the lower troposphere, a well-defined storm track formed only along the PFJ (Fig. 8b), slightly poleward of the Pacific branch of the APFZ and closer to the seasonally expanded ice edge (Fig. 8c). In sharp contrast, the bifurcation of the upper-level storm track activity was much less apparent in the 1998 winter (Fig. 8d). The wave-activity dispersion from the storm-track core region to the STJ was limited to the eastern Indian Ocean. The subtropical branch of the upper-level storm track was confined to the weakened STJ core region over Australia, and no well-defined storm track formed over the subtropical South Pacific. Rather, a substantial fraction of the upper-level wave activity propagated downstream out of the core region over the Indian ocean, and the eddy activity over the South Pacific was organized into a single storm track along the PFJ both in the upper and lower troposphere (Figs. 8d-e). This situation deviated distinctly from the climatology, and rather it resembled the summertime situation (Figs. 6d-f). Despite the somewhat weaker lower-tropospheric eddy activity (measured as a heat flux) than in the previous winter, the surface westerlies along the Pacific PFJ was stronger in the 1998 winter (Fig. 8e), which seems attributable to more coherent vertical structure of the storm track along the PFJ (Figs. 8d-e). In the 1998 winter, 

\textsuperscript{14} Aoki and Hirota (1998) found a tendency that late-winter propagation of planetary waves into the stratosphere is stronger in the presence of the enhanced double-jet structure. Hio and Hirota (2002) showed that variations in the double-jet structure also influence the phase of the stratospheric planetary waves.
anomalously enhanced equatorward dispersion of wave activity occurred from the subpolar storm track over the eastern Pacific, where the weak STJ was confluent with the PFJ (Fig. 8d). The mean-flow westerly (angular) momentum transported poleward due to this dispersion was then transferred from the upper troposphere down to the surface through eddy heat transport, to sustain the strong surface westerlies.

From a macroscopic viewpoint, the OFZs were surprisingly robust for those two winters (Figs. 8c and 8f), despite the overlying atmospheric circulation underwent such dramatic changes in response to the 1997~98 ENSO cycle. The OFZs, particularly those along the ACC, kept acting to anchor the low-level storm tracks along them. Only the noticeable change between the two winters is the poleward displacement of the Pacific STFZ in 1998 relative to its position in the previous year. Equatorward broadening of the low-level storm track occurred slightly to the south of the poleward shifted STFZ over the eastern Pacific. The associated baroclinicity acted to maintain the enhanced eddy activity through which the enhanced surface westerlies in this region could be maintained. The anticyclonic wind shear associated with the westerlies could, in turn, act to sustain the displaced STFZ underneath via the cold Ekman advection, suggestive of a positive feedback among the STFZ, westerly jet, and storm track.

4. Importance of storm tracks and ocean-atmosphere interaction in the extratropical Northern Hemisphere atmospheric circulation system

Over the two ocean basins in the NH, the upper-tropospheric westerlies and associated PV gradient both weaken substantially in summer (Nakamura 1992). Thus, storm track activity also weakens substantially (Klein 1958; Nakamura 1992), although the SST gradient in SAFZs
is as strong as in winter\textsuperscript{15} (Nakamura and Kazmin 2003). Over each of the basins in winter, a major storm track extends eastward from an intense surface baroclinic zone anchored in a SAFZ off the western boundary of the basin\textsuperscript{16} (Figs. 9a and 9c). From a macroscopic view, the storm track is along the boundary of the subtropical and subpolar gyres. Over the North Atlantic, a belt of the surface westerlies between the Icelandic Low and Azores High is situated slightly to the south of the storm track axis. Over the wintertime North Pacific, the equatorward displacement of the surface westerly axis relative to the low-level storm track is more apparent. The former is closer to the STJ axis aloft particularly over the western half of the basin.

Despite the modest intensity of the upper-level westerly jet (Fig. 9b), the midwinter storm track activity over the North Atlantic is stronger than over the North Pacific (Fig. 9a), especially in the lower troposphere (Fig. 9c). The lower-tropospheric storm track axis is closer to a SAFZ in the North Atlantic than in the North Pacific, and the SST gradient in the Atlantic SAFZ is substantially stronger than in the Pacific SAFZ (Fig. 9c). Although the main surface frontal axis extends along the Oyashio Extension at $\sim$42°N, the North Pacific SAFZ is meridionally broad, associated with the Kuroshio-Oyashio Interfrontal Zone between the Extensions of the Oyashio and the Kuroshio (Yasuda et al. 1996; Lin and Talley 1996; Qiu 2002). The North

\textsuperscript{15} Nakamura and Kazmin (2003) showed that SST contrast across the North Pacific SAFZ is even somewhat stronger in summer than in winter.

\textsuperscript{16} In addition to OFZs, surface baroclinicity associated with the thermal contrast between a warm western boundary current (the Gulf Stream or the Kuroshio) and its adjacent cooler landmass is important for the amplification of individual cyclones in winter and thus influential in the storm track formation (e.g., Dickson and Namias 1976; Gulev et al. 2003). Gulev et al. (2003) also pointed out that intensity of sub-synoptic scale disturbances in the lower troposphere is highly correlated with the surface baroclinicity along the SAFZ and the east coast of North America.
Atlantic SAFZ is sharper and thus more intense, which acts to sustain strong surface baroclinicity, contributing to the pronounced baroclinic growth of synoptic-scale eddies.

As well known, the atmospheric circulation over the North Atlantic changes on intraseasonal and interannual time scales under the profound influence of the North Atlantic Oscillation (NAO; van Loon and Rogers 1978; Hurrell 1995a), the most preferred mode of the atmospheric variability in this sector. The NAO not only changes the monthly or seasonal mean circulation but also systematically alters the preferred paths of individual cyclones (Rogers 1990) and thus the storm track activity (Hurrell 1995b). As in the SH case, the North Atlantic SAFZ is robust regardless of the phase of the NAO (Fig. 10), acting to anchor the surface baroclinic zone in the same location. Both in the upper and lower troposphere, the storm track over the western Atlantic extends nearly along the SAFZ in either phase of the NAO, although the axis is slightly displaced poleward in the positive phase (Fig. 10a). In the positive phase, the upper-level westerlies are organized into a well-defined single jet that extends northeastward across the basin over the entire span of the SAFZ. Over the eastern half of the basin, the jet (at ~50°N) is so well separated from the STJ (at ~20°N) that it must behave as a PFJ (Fig. 10a). The storm track almost coincides with that upper-level jet, so that the disturbances are allowed to migrate along the entire span of the SAFZ. This collocation among the jet stream, the SAFZ and the storm track favors the baroclinic growth of the disturbances, consistent with the observed tendency of the enhanced storm track activity in the positive phase of the NAO (Hurrell 1995b). As well known, the surface westerlies tend to be intensified underneath of that well-defined jet, again consistent with the enhanced downward transport of the westerly momentum due to the enhanced storm track activity. In the negative phase, in contrast, the
midlatitude jet is strongly diffluent over the eastern Atlantic, and thus it is not well separated from the STJ (Fig. 10b). This jet diffluence acts to enhance the equatorward dispersion of wave activity. In fact, the storm track axis is deflected to the south and thus detached from the SAFZ, which is consistent with the reduced storm track activity in the negative phase of the NAO, especially in the central and eastern portions of the Atlantic. The midlatitude surface westerlies also tend to be weaker under the reduced eddy activity.

An intriguing aspect of the mean seasonal cycle in the NH storm tracks is the “midwinter minimum (suppression)” in the North Pacific storm track activity (Nakamura 1992), which occurs in spite of the local tendency that both the low-level baroclinicity and westerly intensity aloft are maximized in midwinter\(^{17}\) (Fig. 11b). Bosart (1998) questioned the reality of the phenomenon, speculating that it might merely be a statistical artifact due to Nakamura’s sampling of eddy amplitude on the 250-hPa surface that, along the western Pacific storm track, tends to be situated above the tropopause only in midwinter. However, his speculation is inconsistent with the fact that the eddy-activity suppression can be seen also near the surface (Nakamura 1992; Nakamura et al. 2002). The phenomenon has been reproduced in AGCMs (Christoph et al. 1997; Zhang and Held 1999; Chang 2001). Moreover, analyzing unassimilated aircraft and rawinsonde data obtained for 15 recent years, Chang (2003) has confirmed the midwinter suppression of eddy activity over the Pacific as a climatological-mean signature. He has also confirmed its decadal-scale modulation found by Nakamura et al. (2002) in reanalysis data. They found that the “midwinter minimum” has disappeared since the late 1980s due to the

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\(^{17}\) Nakamura’s (1992) statistics were obtained for the 1965-1984 period, during which the East Asian winter monsoon and STJ were as strong as in the early and mid-1980s for which Fig. 11b is plotted.
enhanced eddy activity in midwinter (Fig. 11a), under decadal weakening of the East Asian monsoon and associated relaxing of the upper-level STJ.

The mechanism that causes the “midwinter suppression” has not been fully understood yet. Diagnosing an AGCM output, Chang (2001) concluded that fall-winter differences in the precipitation pattern within individual cyclone systems are particularly important in the climatological suppression of eddy activity. In midwinter, more precipitation tends to occur in a severe cold-air outbreak behind the cold front of a cyclone, and associated latent heat release in a colder air does not favor the generation of eddy available potential energy. Nakamura (1992), Chang (2001) and Nakamura et al. (2002) pointed out that, though maybe of secondary importance, the excessively strong westerlies act against eddy amplification, driving eddies too fast away from the baroclinic zone in the KOE region. They also showed that, in the presence of the strong westerlies, coherency tends to be lowered between subweekly fluctuations in temperature and the meridional or vertical wind component, leading to the less efficient energy conversion from the mean flow to the eddies.

In search of the reason for the structural changes observed in subweekly eddies, Nakamura and Sampe (2002, hereafter cited as NS02) noticed that, in the course of the seasonal cycle, the Pacific storm track axis underwent greater equatorward displacement in 5 midwinter periods with the most distinct eddy-activity minimum (Fig. 11b) than in 5 other midwinter periods without the activity minimum (Fig. 11a). Amplitude of upper-tropospheric subweekly

18 In NS02, while all of the 5 winters with the least apparent suppression were from the period since 1987, 4 out of the 5 winters with the most distinct midwinter suppression of eddy activity were sampled from the period before 1987, and the 1985/86 winter was the 6-th most distinct. Thus, this modulation in midwinter storm track activity is a decadal-scale phenomenon.
disturbances in the north-western Pacific storm track is significantly correlated only with the eddy amplitude over northern China and Mongolia (NS02), indicating that subweekly eddies propagating into the North Pacific storm track have been mainly through the westerlies along the PFJ rather than along the STJ. Over the wintertime Far East, the strong monsoonal flow near the surface and the enhanced STJ aloft, as observed in the early through mid-1980s, are associated with the pronounced deepening of a planetary-wave trough. With this deepening, the PFJ tends to merge itself into the STJ over the Far East (Mohri 1953; Palmén and Newton 1969). Behind the deepened trough, the upper-tropospheric PFJ has a stronger equatorward component, driving the subweekly eddies strongly toward the intensified STJ located at ~30°N (NS02). Accompanying tight PV gradient, the intensified STJ core acts as an excellent wave-guide that can efficiently trap the eddy activity (Chang and Yu 1999). In fact, for nearly 40% of the time during the midwinter periods with the most pronounced activity minimum, the eddy amplitude maximum\(^{19}\) in the 160°E meridian was located in the vicinity of the STJ core around 32°N at the 200-hPa level (NS02, Fig. 12b). The core is ~12 km in altitude, ~3 km higher than the midlatitude tropopause through which eddies propagate from the Asian Continent into the storm track region over the KOE. The STJ axis is displaced equatorward by 500~800 km relative to the near-surface baroclinic zone anchored by the SAFZ (Figs. 13b and 14b). In sharp contrast, for most of the time during the midwinter periods with no activity minimum (Figs.

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\(^{19}\) As in Nakamura and Wallace (1990) and Nakamura (1992), local, instantaneous eddy amplitude has measured by what is called "envelope function" (\(Z_e\)). The quantity \(Z_e\) was defined locally as twice of the squared time series of the 8-day high-pass filtered geopotential height, which was then smoothed by an 8-day low-pass filter, and finally, the square-root was taken. \(Z_e\) has been multiplied by a factor \([\sin 45°\text{N}/\sin(\text{lat.})]\), so as to mimic eddy amplitude in the geostrophic streamfunction.
13a and 14a), the eddy amplitude maximum stays at the midlatitude tropopause (300-hPa) almost above the SAFZ (at ~40°N). This is an indication of the diminished trapping effect by the relaxed STJ intensity under the weakened winter monsoon. Interestingly, both in late autumn (Fig. 12c) and spring (Fig. 12d) when the Pacific storm track activity is maximized, the eddy amplitude maximum tends to be away from the relaxed STJ core (NS02). Rather, the maximum stays at the midlatitude tropopause over the SAFZ or slightly to its north.

The eddy-trapping by the STJ core exerts some obvious impact upon the eddy structure (NS02). Under no significant trapping effect by the STJ, upper-level eddies coming from the continent can propagate right over the SAFZ (Fig. 14a), which allows them efficient growth through their interaction with the surface baroclinicity. The secondary amplitude maximum just above the surface is an indication of that strong interaction (Fig. 13a). The eddy structure is deep with a strong heat flux in the lower troposphere just above the baroclinic zone. When trapped into the STJ core (Fig. 12b), eddies are lifted up by 3 km from the midlatitude tropopause and displaced by 500~800 km from the SAFZ (Figs. 13b and 14b). By distorting the eddy structure, the trapping impairs the eddy interaction with the surface baroclinic zone (Fig. 13b). The eddy amplitude decays rapidly downward with no secondary maximum at the surface, and the associated heat flux is much weaker than under the diminished trapping. NS02 argued that before the late 1980s, the excessively intense STJ in midwinter trapped the eddy activity to its core and thereby impaired the coupling of upper-level eddies with the surface baroclinic zone along the SAFZ, thus contributing to the midwinter eddy-activity minimum as observed (Fig.11b). Since then, under the reduced trapping effect by the relaxed STJ intensity, upper-level eddies were allowed to propagate right above the baroclinic zone,
while strongly interacting it, even in midwinter as in fall and spring. NS02 argued that this is why the North Pacific storm track activity enhanced since the 1986-87 winter, leading to the disappearance of its midwinter minimum (Nakamura et al. 2002). This tendency is consistent with Nakamura (1992), Chang (2001) and Nakamura et al. (2002), who found through correlation analyses that the structure of subweekly disturbances tends to be more suited for their baroclinic growth under the relaxed STJ.

As in the South Pacific case shown in Fig. 8, the North Pacific SAFZ was robust and quite consistent between the two types of winters (Fig. 14). Only the noticeable difference is the somewhat enhanced SST gradient in the STFZ core during the winters with the eddy-activity minimum (Fig. 14b) relative to the winters without the minimum (Fig. 14a), indicating that the change in the surface baroclinicity was unlikely the reason for the observed changes in the storm track activity. The axis of the surface westerlies nearly followed the STJ axis, located between 30°N and 35°N, in the winters of the suppressed storm track activity (Fig. 14b). In the winters of the enhanced storm track activity (Fig. 14a), the surface westerly axis was displaced poleward by ~10°. The axis was situated along the northern fringe of the SAFZ over the western Pacific, whereas it was systematically below the upper-tropospheric storm track axis over the eastern Pacific. Nakamura (1992) showed that, in the course of the seasonal march, the axis of the barotropic component of the lower-tropospheric westerlies tends to follow the axis of the upper-tropospheric storm track axis over the eastern Pacific. He also showed that, over the western Pacific, the former axis almost coincides with the storm track axis only in spring and fall when the storm track activity is maximized seasonally.
5. Discussion and concluding remarks

It turns out from our argument that the close association among a storm track, a PFJ and a SAFZ (including the APFZ) is crucial for the two-way interaction between the midlatitude ocean and atmosphere, as exemplified in a diagnosing study by Tanimoto et al. (2003) and also in an AGCM experiment by Peng and Whitaker (1999) both on the decadal variability inherent to the North Pacific (Nakamura et al. 1997). It also turns out from our argument that the whole dynamical picture of the formation and climatological seasonal variations of the observed storm tracks and PFJs is unlikely to be obtained without taking their interaction with the underlying OFZs into account, as implied by NS02, Inatsu et al. (2003) and NS03. It has been widely accepted that the differential radiative heating acts to restore the mean baroclinicity in midlatitudes against relaxing effect by eddy heat transport, but it provides no clear explanation why such intense surface baroclinic zones as observed are maintained. The fact that strongest surface baroclinic zones over the ocean are anchored by SAFZs (including APFZs in the SH) suggests the effective restoring of the atmospheric baroclinicity by those midlatitude OFZs, owing to the large thermal inertia of the ocean mixed layer and also to the differential thermal advection between to the north and south of the OFZs by strong oceanic currents. Within each of the NH ocean basins, the core region of a surface baroclinic zone forms off the east coast of a continent, where the westerly boundary currents coming from the north and south join to form a core region of a SAFZ, and a cold, dry continental air mass frequently intrudes equatorward in winter. The anchoring of a storm track by the underlying SAFZ is persistent but not quite strong, so that their association can be disturbed by some external forcing including seasonal or interannual intensification of a STJ. The latter intensi-
fication may be due to a teleconnection from the Tropics or an upstream continent. An important scientific issue emerging is how the surface baroclinicity is determined within the oceanic boundary layer. NS03 locally evaluated it as a quantity equivalent to the Eady growth rate for the layer between the 925 and 1000-hPa levels, calculating the horizontal temperature gradient and static stability based on a reanalysis data set. They repeated the evaluation by using satellite-measured SST in place of the re-analyzed air temperature. Although the overall features are similar between the two evaluations, some differences are found with respect particularly to the sharpness of baroclinic zones.

Another aspect of the air-sea coupling associated with a storm track is that the mean westerly momentum carried downward with upward wave-activity transfer along the storm track (Lau and Holopainen 1984; Lau and Nath 1991) is organized into a surface westerly jet, which drives oceanic gyres (or the ACC), thus contributing to the maintenance of SAFZs (or the APFZ). Along the ACC, the core regions of the storm track, surface westerlies and APFZ almost coincide with each other, indicative of the local feedback forcing among them. Over each of the NH ocean basins, in contrast, the SAFZ core is located at the confluent region of the western boundary currents that are driven mainly through gyre adjustment by the surface westerlies that are strongest over the central part of the basin. The strong surface wind along the storm track also enhances the surface evaporation, whereas precipitation associated with individual migratory storms mostly determines the fresh water supply to the midlatitude ocean (Lukas 2001). Kinetic energy input into the ocean by the strong surface westerlies and vigorous storm activity acts to sustain the mixed layer structure. The input also becomes a source of oceanic turbulence available for deep-layer mixing (Nagasawa et al. 2000).
The findings in this article and related papers (NS02, NS03) require some modification to conceptual models for the zonally symmetric circulation in the wintertime troposphere\textsuperscript{20}, including the one by Palmén (1951) as shown in Fig. 15b. Palmén and Newton (1969) modified Palmén’s model by incorporating the concept of air masses (Fig. 15c). While resembling their original version (Fig. 15a) given by Rossby (1941), Palmén’s models emphasize the concentration of mean westerly momentum into a STJ and PFJ in their respective association with the Hadley cell and a polar frontal zone\textsuperscript{21}. That frontal zone is depicted in Rossby’s model as a convergence zone at \(~60^\circ\) in latitude that separates the midlatitude surface westerlies and the surface easterlies in the polar region, whereas in Palmén’s models the zone is situated within the Ferrel cell and thus in the low-level westerlies as observed. The distinct poleward tilt of the polar frontal zone with height in Palmén’s models probably reflects the fact that his models were based on a number of instantaneous sections each of which depicts a cold front associated with a cyclone. We know these days that this kind of conceptual models better applies to the SH, as in the analysis of NS03, where zonal asymmetries in the jet streams exerted by planetary waves are much weaker than in the NH. On the basis of our findings for the SH, the most fundamental modification that must be added to Palmén’s models is the close association among a PFJ, storm track, surface baroclinic zone and SAFZ (or APFZ), as schematically illustrated in Fig. 16. The mean atmospheric circulation system in midlatitude

\textsuperscript{20} Refer to an excellent review by Lorenz (1967) for the history of the development of conceptual models for the atmospheric general circulation.

\textsuperscript{21} The concepts of fronts and air masses were introduced by V. Bjerknes, J. Bjerknes and their collaborators in the “Norwegian school”, whereas the concept of jet streams was by C.-G. Rossby and his collaborators in the “Chicago school” (Palmén and Newton 1969).
must be considered in the context of ocean-atmosphere interaction. Furthermore, in winter a PFJ is connected to a stratospheric polar-night jet over the SH and North Atlantic. We also know that this kind of conceptual models should be representative of a mean state from which fluctuations associated with individual transient eddies have been eliminated. As shown in Fig. 7 (see also NS03), mean baroclinicity in the core region of the SH storm track increases downward below the PFJ with only a slight poleward tilt with height, indicating that baroclinic zone extends almost vertically above the APFZ. In adding the above modifications to Palmén’s models, we need to clarify the meaning of the “steadiness” of a jet stream. Palmén and Newton (1969) contrasted the steadiness of a STJ, as first pointed out by Namias and Clapp (1949), with the unsteadiness of a PFJ. The latter is a manifestation of the enhanced storm track activity along a PFJ, whereas the former means suppressed eddy activity along a STJ. On seasonal time scales, PFJs and STJs exhibit the opposing tendency in their steadiness, especially over the SH. While the PFJ is apparent throughout the year, the STJ becomes distinct only in winter and spring. Such distinctions are less apparent over the NH.

One of the crucial factors that control the observed seasonal evolution of storm tracks is a distinct characteristic between the STJ and PFJ, as our schematics in Fig. 16 are classified into two types depending upon the strength of a STJ. The PFJ has a deep structure throughout the depth of the troposphere accompanied by a distinct baroclinic zone near the surface (Fig. 7). From a PV viewpoint, this vertical structure favors the growth of baroclinic eddies. Over the SH, the cores of the storm track and PFJ are anchored above the APFZ core throughout the year (Fig. 6). Over the North Pacific and Atlantic, the good correspondence is also observed between the storm track core and SAFZ (Fig. 9). The strong near-surface
westerlies along the PFJ are sustained by the downward transport of the mean flow momentum by eddies. Thus, the deep structure of the PFJ is consistent with the eddy-driven nature of the jet (Lee and Kim 2003). The westerly momentum transported downward by eddies must be transferred into the ocean via turbulent fluxes, acting to drive the ocean current and maintain the associated SAFZ (or APFZ). In light of this atmospheric forcing on the ocean below the PFJ, their coexistence among the storm track, PFJ and surface baroclinic zones and OFZs is likely a manifestation of a feedback loop among them, which is another indication of the significance of storm tracks in midlatitude ocean-atmosphere interaction.

As speculated by Palmén (1951) and elucidated later in a simple theory of the zonally symmetric Hadley cell (Held and Hou 1980; Lindzen and Hou 1988), a STJ forms through the poleward transport of angular momentum and it is much stronger in the winter hemisphere22. Zonal asymmetries in the tropical SST distribution or the presence of large landmasses can lead to the formation of a STJ core region (Inatsu et al. 2002). In fact, the STJ formation over the SH is known to be related to the activity of the Asian summer monsoon (Nogues-Paegle and Zhen 1987; Berbery and Nogues-Paegle 1993). Thus, a STJ may not necessarily be accompanied by a distinct surface baroclinic zone. Indeed, the STJ axis lies in a latitudinal band between the SAFZ and STFZ over the North Pacific (Figs. 9 and 14). Over the SH, a STJ is located coincidentally over STFZs and thus accompanied by modestly strong baroclinicity near the surface over the Indian and Pacific Oceans. Below the jet, however, a belt of subtropical high extends zonally near the surface, which counteracts the baroclinic eddy growth.

22 Refer to Xie (2003) for a historical, comprehensive review of the Hadley cell.
Recently, Lee and Kim (2003) performed intriguing experiments, through which they examined how storm track activity depends on the intensity of a STJ. They attempted to compromise two conflicting-looking views of the tropospheric circulation: one is a STJ-PFJ double-jet system with a main storm track branch along the PFJ as in Palmén’s models, and the other is a STJ-dominant system where a storm track forms only along the poleward fringe of the jet as in Lindzen’s (1993) argument. In their idealized setting, a STJ, as it intensifies, becomes dominantly favorable for baroclinic growth of synoptic-scale eddies, because of the increasing atmospheric baroclinicity below the jet. Meanwhile, a main storm track branch forms along a PFJ only when the jet forms into a double-jet structure with a weak STJ. They concluded that in winter the North Atlantic is in the double-jet regime whereas the North Pacific is in the STJ-dominant regime. Some of their results sound consistent with our observational findings and those in NS02 and NS03. However, the greatest discrepancy is that, as opposed to what is suggested from their idealized experiments, the intensification of a STJ in the real atmosphere is unfavorable for the baroclinic eddy growth and hence the storm track formation. Over each of the North and South Pacific Oceans, the wintertime intensification of a STJ causes trapping of upper-level eddy activity into the jet core, keeping it away from a midlatitude baroclinic zone anchored by a SAFZ (including the APFZ). The trapping impairs the baroclinic growth of synoptic-scale eddies, resulting in a reduction of storm track activity, despite the enhanced baroclinicity below the STJ core. Over the South Pacific, where the STJ is well separated latitudinally from the PFJ, the trapping leads to the meridional separation of the main storm track branch between the upper and lower troposphere (NS03). This is a situation of the STJ-dominant regime (Fig. 16b), as observed most typically in the 1997 El Niño winter.
No such separation is apparent over the North Pacific, where a PFJ is merged into a STJ associated with a deep planetary-wave trough (Mohri 1953). Still, when extremely intensified, the STJ traps migratory eddies into its core away from the surface baroclinic zone along the SAFZ, resulting in the suppression of storm track activity in midwinter (Nakamura 1992). This is a situation between the two types shown in Fig. 16. The eddy activity enhances when synoptic-scale eddies can propagate right above the SAFZ away from the jet core under the weakened STJ as in autumn and spring, or even in winter since the late 1980’s under the weakened winter monsoon (NS02). In the real atmosphere, as usually observed over the South Indian Ocean in austral winter, or even over the South Pacific in the 1998 winter, the main storm track branch exhibits apparent preference for staying with a PFJ even when a STJ intensifies. This “weak-STJ regime” (Fig. 16a) appears in the purest manner in the summertime SH under the weakened STJ. Owing to the anchoring effect by the underlying SAFZ (including the APFZ) through its interaction with the PFJ and storm track, the preference in the real atmosphere must be more robust than in the idealized experiments by Lee and Kim (2003) that include no well-defined surface baroclinic zones anchored by SAFZs. In light of its eddy-driven nature, a PFJ owes its presence to its interaction with the underlying SAFZ, while the lack of an intense OFZ underneath must be one of the main reasons why baroclinic eddy growth is suppressed along a STJ despite the high atmospheric baroclinicity under its core.

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Fig. 1 (a) Climatological Jan.-Feb. distribution of sea-level pressure (hPa; contoured for every 4, heavy lines for 1000 and 1020) and 1000-hPa wind (m s$^{-1}$; arrows with scaling at the lower-right corner), and 250-hPa zonal wind speed (m s$^{-1}$; shaded for 40–50 and 60–70). (b) As in (a), but for 850-hPa meridional heat fluxes (K m s$^{-1}$) associated with the Pacific storm track. Contour intervals: 3. Shading for the values exceeding 15. (c) As in (a), but for 500-mb vertical pressure velocity (contoured) superimposed on westerly wind shear between the 700 and 1000-mb levels (shaded lightly and heavily where the values exceed 10 and 14 m s$^{-1}$, respectively). Contour intervals: 0.02 (Pa s$^{-1}$); dashed for negative values (upward) and zero lines omitted. (d) As in (b), but for the 250-mb feedback from the storm track measured as the westerly acceleration migratory transient eddies act to induce through their vorticity flux. Contour intervals: 1 (m s$^{-1}$/day); dashed for negative values (easterly) and zero lines omitted. Bold line: mean storm track axis as defined by local maxima of the envelope function of 250-mb height at each meridian. (e) As in (a), but for a latent heat flux from the surface (every 40 W m$^{-2}$). (f) As in (a), but for precipitation (contoured for every 1 mm/day; shaded lightly and heavily where the values exceed 4 and 6, respectively). Based on (a-e) reanalysis data by the National Centers for Environmental Prediction (NCEP) and National Center for Atmospheric Research (NCAR) and (f) Climate Prediction Center Merged Analysis of Precipitation (CMAP) data for 1979–1995. (b-f) After Nakamura et al. (2002).
Fig. 2  Difference maps of bi-monthly SSTA (°C) associated with the dominant mode of the North Pacific decadal climate variability between a pair of 4-year periods of 1968/69~1971/72 and 1982/83~1985/86 (i.e., “warm” minus “cold”) for the (a) Nov.-Dec., (b) Jan.-Feb. and (c) Mar.-Apr. periods. Based on the Comprehensive Ocean-Atmosphere Data Set (COADS) compiled on a 2°×2° latitude-longitude grid. Coloring convention is shown at the bottom. After Tanimoto et al. (2003).

Fig. 3  (a) As in Fig. 2, but for total upward latent heat flux anomalies (W m⁻²) at the surface for the December-January period. (b) As in (a), but for a contribution only from local SSTAs. Based on linearization of Halliwell and Meyer (1996) applied to the total anomalous flux in (a). (c) As in (b), but for a contribution only from local air temperature anomalies. (d) As in (b), but for a contribution only from local wind speed anomalies. The SAFZ-KOE region is indicated with a rectangle. (e-h): As in (a-d), respectively but for the Feb.-Mar. period. After Tanimoto et al. (2003).
Fig. 4  As in Fig. 3, but for (a) 250-hPa height (contoured with 20-m intervals), (b) storm track activity measured as the 850-hPa meridional heat flux associated with subweekly disturbances (colors; K m s\(^{-1}\)), and (c) feedback forcing from anomalous storm track activity measured as 250-hPa geopotential height tendency (colors; unit: m day\(^{-1}\)) calculated only from the vorticity flux convergence, all for the December-January period. Based on the NCEP/NCAR reanalyses. In (a), 250-hPa wave-activity flux (m\(^2\) s\(^{-2}\)) associated with this anomaly pattern is superimposed with arrows. The flux has been evaluated from 250-hPa anomalous geostrophic streamfunction using a formula developed by Takaya and Nakamura (2001). In (c), the 250-hPa height tendency is associated with subweekly fluctuations at the same level (Nakamura et al. 1997b). Scaling of arrows is given at the top. After Tanimoto et al. (2003).

Fig. 5  As in Fig. 2, but for surface wind velocity (arrows) and scalar wind speed (colors; unit: m s\(^{-1}\)) for the (a) December-January and (b) February-March periods. Coloring convention and scaling of the arrows are given at the bottom. After Tanimoto et al. (2003).
Fig. 6. (a) Climatological Jul.-Aug. distribution of 250-hPa storm track activity (stippling) and horizontal component of 250-hPa extended E-P flux (arrows; scaling at the bottom with unit: m² s⁻²), with 250-hPa westerly wind speed (heavy solid lines for 30, 40, 50 and 60 m s⁻¹; heavy dashed line for 20 m s⁻¹) and 250-hPa). Light and heavy stippling is applied where amplitude of subweekly fluctuations in geopotential height is between 90 and 130 (m) and above 130 (m), respectively, with thin lines for every 10 (m). Polar stereographic projection poleward of 20°S. (b) As in (a), but for 925-hPa westerly wind speed (heavy lines for 3, 6, 9, 12 and 15 m s⁻¹; dashed for zero wind line), superimposed on 250-hPa westerly wind speed (light stippling for 20~30 m s⁻¹; heavy stippling for 40~50 m s⁻¹). (c): As in (a), but for 850-hPa storm track activity measured as poleward heat flux associated with subweekly eddies (heavy lines for 4, 8, 12 and 16 K m s⁻¹). Light and heavy stippling indicates oceanic frontal zones where meridional SST gradient (deg./110 km) exceeds 0.6 and 1.2, respectively (thin lines are drawn for every 0.6). Dark shading indicates data-void regions. (d-f): As in (a-c), respectively, but for the Jan.-Feb. period.
Fig. 7 (a-c): Climatological-mean meridional sections of zonal wind speed (contoured for every 5 m s$^{-1}$; dashed for the easterlies), superimposed on meridional (m$^2$ s$^{-2}$) and vertical (Pa m s$^{-2}$) components of the extended E-P flux (arrows; scaling at the lower-right corner), for (a) January and (b) July in the Indian Ocean sector (50$^\circ$~90$^\circ$E) and (c) for July in the western Pacific-Australian sector (120$^\circ$~160$^\circ$E) over the Southern Hemisphere. Hatching indicates data void regions due to topography. (d-f): As in (a-c), but for eddy amplitude ($Z_e$; unit: m; heavy lines for every 20 from 40) and mean baroclinicity (G; thin lines for every 0.05 from 0.15; light stippling for 0.2~0.35 and heavy stippling for above 0.35). In (a-c), stippling for $Z_e > 80$ (m). After Nakamura and Shimpo (2003).
Fig. 8  (a) Horizontal component of 250-hPa extended E-P flux (arrows; scaling at the bottom with unit: m$^2$ s$^{-2}$), superimposed on 250-hPa westerly wind speed (heavy solid lines for 30, 40, 50 and 60 m s$^{-1}$; heavy dashed line for 20 m s$^{-1}$) and 250-hPa storm tracks (stippling), for July-August, 1997. The light and heavy stippling is applied where frequency of the eddy amplitude maximum passing through a given data point (with 2.5$^\circ$ latitudinal intervals) on a given meridian, defined as the number of days over the 62-day period, exceed 3 and 6 (thin lines for every 3). Polar stereographic projection poleward of 20$^\circ$S only for the South Indian and Pacific Oceans. (b): As in (a), but for 925-hPa westerly wind speed (heavy solid lines for 5, 10 and 15 m s$^{-1}$), superimposed on lower-tropospheric storm track activity measured as 850-hPa poleward heat flux associated with subweekly fluctuations (unit: K m s$^{-1}$; light contours for every 4; light stippling for 8~16 and heavy stippling for 16~28). (c): As in (a), but for oceanic frontal zones where local meridional SST gradient (deg./110 km) is 0.6~1.2 (light stippling) and above 1.2 (stippling with light contours for every 0.6), superimposed on 250-hPa westerly wind speed (heavy lines). (d-f): As in (a-c), respectively, but for July-August, 1998.
Fig. 9  (a) Climatological January-February distribution of the horizontal component of 250-hPa extended E-P flux (arrows; scaling at the bottom with unit: m$^2$ s$^{-2}$), superimposed on 250-hPa storm track activity measured as the envelope function of subweekly height fluctuations (contoured for every 10 m). (b) As in (a), but for 925-hPa westerly wind speed (heavy lines for 3, 6, 9, 12 and 15 m s$^{-1}$), superimposed on the mean 250-hPa westerly wind speed with light stippling for 30–40 (m s$^{-1}$) and heavy stippling for 50–60 (m s$^{-1}$). (c): As in (a), but for lower-tropospheric storm track activity measured as 850-hPa poleward heat flux associated with subweekly fluctuations (heavy lines for every 4 K m s$^{-1}$). Light and heavy stippling indicates oceanic frontal zones where local meridional SST gradient (deg./110 km) is 0.6–1.2 (light stippling) and above 1.2 (heavy stippling with thin contours for every 0.6). Dark shading indicates data-void regions.
Fig. 10  Relationship over the North Atlantic among the oceanic frontal zones (stippling), storm track axes at the 1000-hPa (solid lines) and 300-hPa (series of vertical lines) levels, and westerly jet axes at the 1000-hPa (dashed lines) and 250-hPa (dotted lines) levels, (a) for 4 midwinter (Jan.-Feb.) periods with the strongest positive phase of the NAO and (b) 4 midwinter periods for its strongest negative phase during the 1980~1998 period. Position and intensity of oceanic frontal zones are represented as region of strong meridional SST gradient (deg./110 km) with stippling as indicated on the right hand side of (n). The storm track axis is defined as a local maximum of envelope function of subweekly geopotential height. The NAO index used for the classification was based on the first principal component time series of wintertime (Dec.-Mar.) sea-level pressure anomalies over the North Atlantic domain $[20^\circ N\sim70^\circ N, 90^\circ W\sim40^\circ E]$ (Hurrell 1995ab).
Fig. 11  Seasonal march in NW storm track activity, composited in latitude-time sections for (a) 5 winters for the weakest midwinter suppression of the activity and for (b) 5 winters with the most distinct suppression, during the 1979-1995 period. The storm track activity is measured as 850-hPa poleward eddy heat flux averaged 100°E-180° (every 2 K m s$^{-1}$; heavy lines for 10). Tick marks along the abscissa indicate the first days of months. From Nakamura et al. (2002) and Nakamura and Sampe (2002).

Fig. 12  (a) Frequency of the eddy amplitude maximum passing through a given data point on the meridional plane at 160°E, indicated as the number of days over a 59-day midwinter (Jan.-Feb.) period (heavy sold lines; every 2), for 5 of those periods with the weakest suppression of the Pacific storm track activity during the 1979-1995 period. Dotted lines for basic-state PV (every 1 PV unit). Westerly jet is indicated by shading (U: 20~30, 40~50 and 60~70 m s$^{-1}$). (b): As in (a), but for another 5 midwinter periods with the most distinct activity suppression. (c): As in (a), but for 5 late autumn (mid-Oct. through mid-Dec.) periods preceding the 5 midwinter periods in (b). (d): As in (a), but for 5 spring (Mar.-Apr.) periods following the 5 midwinter periods in (b). After Nakamura and Sampe (2002).
Fig. 13  Typical meridional structure of a baroclinic eddy in the core region (170°E~170°W) of the Pacific storm track, on the basis of subweekly fluctuations in geopotential height (Z') field regressed linearly upon 300-hPa Z' at [47°N, 105°E] with a 2-day lag, for (a) 5 midwinter (Jan.–Feb.) periods with the weakest midwinter suppression of the Pacific storm track activity and for (b) another 5 periods with the most distinct suppression during the 1979-1995 period. Eddy amplitude in Z' is normalized by its maximum (30, 50 70 and 90%). Associated poleward temperature flux based on the regression (K m s^{-1}; density adjusted) is plotted with dashed lines for 0.56, 0.84, 1.12 and 1.40 in (a) and 0.2, 0.3 and 0.4 in (b). Note that eddy amplitude is larger in (a) by 67%. The westerly jet is indicated with shading (U: 20~30, 40~50 and 60~70 m s^{-1}), and meridional gradient of basic-state surface air temperature is plotted at the bottom (K/ 100 km; sign reversed). After Nakamura and Sampe (2002).

Fig. 14  Relationship over the North Pacific among the oceanic frontal zones (stippling), storm track axes at the 1000-hPa (solid lines) and 300-hPa (series of vertical lines) levels, and westerly jet axes at the 1000-hPa (dashed lines) and 250-hPa (dotted lines) levels, (a) for 5 midwinter (Jan.–Feb.) periods with the weakest suppression of the Pacific storm track activity and (b) for another 5 midwinter periods of its most distinct suppression during the 1979-1995 period. Position and intensity of the frontal zones are represented with local magnitude of meridional SST gradient indicated with stippling as at the bottom of (a). The storm track axis is defined as a local maximum of envelope function of subweekly geopotential height. The classification of the periods is based on Nakamura et al. (2002).
Fig. 15  Schematics of zonal-mean tropospheric general circulation over the winter hemisphere proposed by (a) Rossby (1941), (b) Palmén (1951) and (c) Palmén and Newton (1969). In (c), STJ and PFJ denote a subtropical and polar-front jet streams. Three types of principal air masses are signified by TA (tropical), MLA (midlatitude) and PA (polar).

Fig. 16  Schematics of tropospheric general circulation over the ocean based on our argument. (a) In the case where, in the presence of a relatively weak subtropical jet (STJ), the main storm track (thick dashed line) forms over an intense surface baroclinic zone (stippled at ~45° latitude) anchored by a subarctic oceanic frontal zone (SAFZ), as in the North Atlantic, the South Atlantic and the South Indian Oceans. Wave-activity dispersion toward the STJ (wavy arrow) leads to the formation of a well-defined polar-front jet (PFJ) above the SAFZ. Downward westerly-momentum transport occurs in association with poleward heat fluxes due to synoptic-scale eddies (stippled arrow) migrating along the storm track, leading to the formation of a surface westerly jet (circled W) along the SAFZ. This eddy transport of the mean westerly angular momentum is indicated by an open arrow. (b) In the case as in the wintertime South Pacific, where an intense STJ traps most of the eddy activity. The main branch of the upper-level storm track forms along the STJ core with suppressed baroclinic eddy growth below. The main low-level storm track forms along an intense baroclinic zone anchored by the SAFZ, but the PFJ is weak in the Southern Hemisphere, the SAFZ corresponds to the Antarctic Polar Frontal Zone, and the subpolar and subtropical gyres to the cyclonic and anticyclonic shear sides of the Antarctic Circumpolar Current. The situation of the wintertime North Pacific is in between (a) and (b), with a SAFZ not well separated from an intense STJ.