Storm Track Dynamics

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ABSTRACT

This paper reviews the current state of observational, theoretical, and modeling knowledge of the midlatitude storm tracks of the Northern Hemisphere cool season.

Observed storm track structures and variations form the first part of the review. The climatological storm track structure is described, and the seasonal, interannual, and interdecadal storm track variations are discussed. In particular, the observation that the Pacific storm track exhibits a marked minimum during midwinter when the background baroclinicity is strongest, and a new finding that storm tracks exhibit notable variations in their intensity on decadal timescales, are highlighted as challenges that any comprehensive storm track theory or model has to be able to address.

Physical processes important to storm track dynamics make up the second part of the review. The roles played by baroclinic processes, linear instability, downstream development, barotropic modulation, and diabatic heating are discussed. Understanding of these processes forms the core of our current theoretical knowledge of storm track dynamics, and provides a context within which both observational and modeling results can be interpreted. The eddy energy budget is presented to show that all of these processes are important in the maintenance of the storm tracks.

The final part of the review deals with the ability to model storm tracks. The success as well as remaining problems in idealized storm track modeling, which is based on a linearized dynamical system, are discussed. Perhaps on a more pragmatic side, it is pointed out that while the current generation of atmospheric general circulation models faithfully reproduce the climatological storm track structure, and to a certain extent, the seasonal and ENSO-related interannual variations of storm tracks, in-depth comparisons between observed and modeled storm track variations are still lacking.

1. Introduction

It has long been appreciated that mobile, O(1000 km) scale high and low pressure systems generate much of the day-to-day variability in sensible weather in the midlatitudes. Given this fact, it is natural that the geographical organization of these transients, whether in terms of their preferred paths of travel, relative frequency of occurrence, or the ''average'' magnitude of variability, has been and remains a topic of extreme relevance to the science and practice of weather forecasting. The synoptic classification of such preferred regions of storm (cyclone) activity, or *storm tracks,* dates at least to the mid-nineteenth century. Figure 1 shows the global distribution of storm activity as it was perceived in the mid-nineteenth century (Hinman 1888). Many features familiar to the modern synoptic picture of the geographical distribution of cyclone occurrence frequency and cyclogenesis, as discussed by Pettersen (1956), Klein (1957, 1958), and Whitaker and Horn (1982, 1984), among others, can be inferred even from this primitive figure (aside from the apparent confusion between warm core hurricanes and midlatitude baroclinic systems): a maxima in cyclone occurrence extending from the East China Sea across the Pacific, shading into weaker activity over the Rockies; a second maxima in occurrence extending from the eastern slopes of the Rockies across the Atlantic toward northern Europe, shading into weak-

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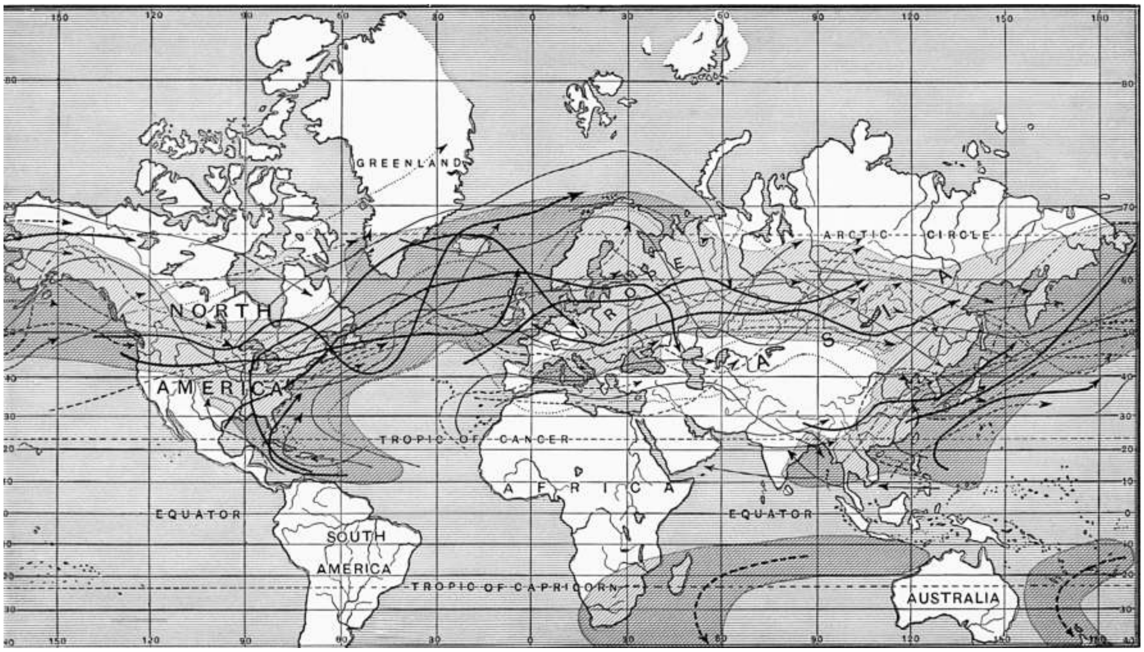


FIG. 1. A figure from an 1888 geography text showing storm frequency distribution as viewed in the mid-nineteenth century. The stipling denotes high storm frequency, while the arrows indicate individual storms. Reproduced from Hinman (1888).

er activity over central Asia; and a third weak maxima in cyclone activity located over the Mediterranean, extending into central Asia. Consistent with the eastward propagation of disturbances, cyclogenesis preferentially occurs on the westward fringe of the areas of maximum cyclone occurrence.

The advent of gridded atmospheric analyses at regular time intervals in the late 1970s heralded a new and dynamically more complete picture of storm track structure. Blackmon (1976) and Blackmon et al. (1977), following a methodology that can be traced to Klein (1951), showed that the atmosphere is described by a dispersion relation of sorts, as *time filtering* a series of gridded analyses maps to isolate disturbances with periods of 2–7 days (see also Hartmann 1974; Randel and Stanford 1985) isolates the $O(1000$ km) spatial-scale mobile transients familiar from the above synoptic classification of storm tracks. Further, this ''bandpass'' filtering has the distinct advantage vis-á-vis synoptic classification that it can be carried out at all levels in the atmosphere, allowing the development of a true three-dimensional picture of storm tracks. The original diagnoses of Blackmon and collaborators, along with numerous others since, provide an alternative definition of storm tracks as *geographically localized maxima in bandpass transient variance.* Examples of storm track structure that emerge from such an analysis are shown in Figs. 2a–c, where the storm tracks are marked in the various bandpass standard deviation fields by enhanced variability off the east coasts of Asia and North America, more or less coinciding with the regions of maximum cyclone occurrence described above.

With their strong connection to sensible weather, storm tracks play a prominent part in midlatitude climate dynamics. Regardless of how one chooses to define storm tracks, a systematic shift in either their geographical location or the level of storm activity will lead to substantial precipitation anomalies with consequent impacts on regional climates. A particularly pointed example of precipitation anomalies resulting from a change in storm track structure occurs during strong El Niño events, when the Pacific storm track extends much farther downstream than it does during ''normal'' winters. This downstream extension brings more active landfalling cyclones to California, resulting in flooding, landslides, and beach erosion.

However, it is not only the ''obvious'' changes in precipitation patterns associated with shifts in storm track structure that explains why storm tracks are a topic of such vital importance to climate dynamics. Rather, over the past decade there has been a growing realization that storm tracks are symbiotically linked (following the terminology of Cai and Mak 1990) to the planetary-scale flow. To be concrete, consider a common problem in climate dynamics; namely, diagnosing an anomaly in the planetary-scale flow associated with some imposed external forcing, that is, anomalous tropical heating associated with El Niño SST anomalies. In general, a corresponding shift in the storm track structure will accompany the anomaly in the planetary-scale flow (Branstator 1995). However, diagnoses have shown that the storm track shift, through anomalous fluxes of heat and momentum, *often forces a larger component of the observed planetary-scale flow anomaly than the imposed*

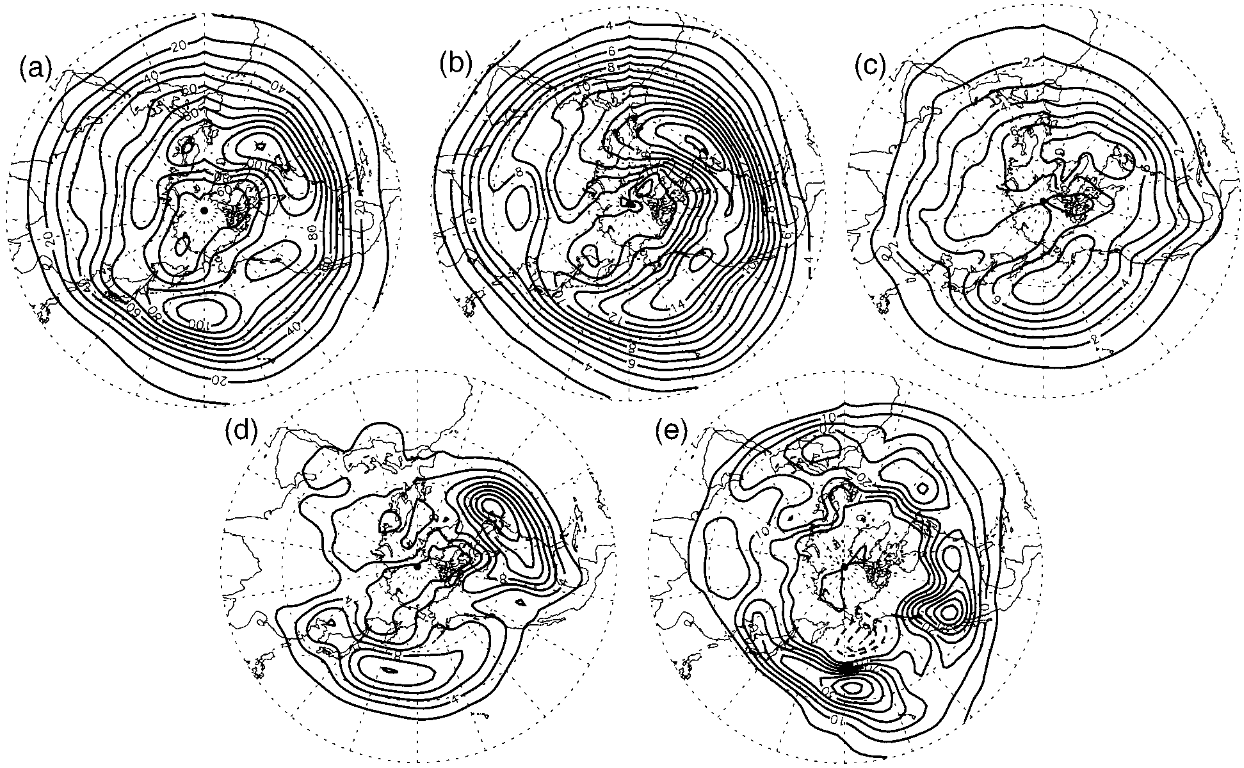


FIG. 2. Bandpass statistics from the NCEP–NCAR reanalysis. Std devs of (a) 250-hPa Z (contour 10 m), (b) 300-hPa v (contour 2 m s^{-1}), and (c) SLP (contour 1 hPa). Poleward fluxes of (d) 850-hPa heat (contour 2 K m s^{-1}) and (e) 250-hPa westerly momentum (contour 5 m^2 s^{-2}).

external forcing itself (Held et al. 1989; Hoerling and Ting 1994). Given this fact, climate simulation skill, whether in the context of seasonal-to-interannual forecasting or climate change scenarios, appears tantamount to proper representation of storm track dynamics in such models.

The apparent importance of storm tracks to midlatitude climate dynamics suggests that advances in the observational, theoretical, and modeling aspects of storm track dynamics will pay large dividends in the development of a ''theory'' of climate. To the authors' knowledge, at this point in time no single work exists that touches upon this triumvirate of storm track dynamics. The intent of this work is to provide an overview of the current state of the storm track problem in all of its aspects. As this topic has been one of the central foci of the Geophysical Fluid Dynamics Laboratory (GFDL)/University consortium over the past decade, in which all three authors were active, the viewpoint presented herein mirrors aspects emphasized during that project. Moreover, its focus is primarily on the Northern Hemisphere cool season, as it is during that season that synoptic-scale storm track activity is largest. As is inevitable in such a review, certain topics of undoubted importance will only be cursorily touched upon; the most glaring omission concerns a discussion of model simulations of storm track changes anticipated due to increasing CO_2. In part, this is because such simulations are relatively new and their place in the overall picture of storm track dynamics has yet to be firmly determined. Nonetheless, it is hoped that this review provides a useful framework within which to interpret such simulations.

Observed storm track structures compose the first member of the triumvirate, and are treated at length in section 2. The review not only touches upon the observed climatological structure of storm tracks, which given the availability of the various reanalysis projects now exists on quite solid footing in the extant literature, but also on the variability of storm track structures across a broad range of timescales. Examples abound that test the theoretician's intuition and the modeler's skill: the annual cycle of storm track activity in the Pacific exhibits a marked minimum during the midwinter, first noted by Nakamura (1992), which at first glance is inconsistent with the perceived annual cycle of baroclinic available potential energy generally thought to fuel storm tracks, which is largest during midwinter; the ENSO cycle on interannual timescales where large shifts in storm track structure occur in response to changes in the subtropical jet associated with anomalous tropical heating, as well as due to the two-way interaction between storm tracks and the midlatitude planetary-scale flow; and new research showing that storm tracks exhibit marked variations in their intensity on decadal timescales (Chang and Fu 2002). These observed variabilities in storm track structure provide a rigorous standard against which the ability of both numerical models and theoretical paradigms can be tested, and serve to focus the remainder of the review.

Physical processes vital to storm track dynamics form the second member of the triumvirate, and are treated in section 3. Taking for granted the longitudinal variations in the planetary-scale flow due to land–sea contrasts and stationary Rossby waves forced by topography and heating (see the review by Held et al. 2002, this issue), there are many fundamental questions regarding the nature of storm tracks that now have reasonable theoretical answers. These questions include the following: (i) What fuels transient eddy development within storm tracks? (ii) Are storm tracks self-sustaining in the sense that they can develop eddies in the absence of ‘‘seeding’’ by recirculation of disturbances around the globe? (iii) Do transient eddy life cycles change as a function of longitude within storm tracks? (iv) What terminates storm tracks? (v) To what extent is latent heating important to overall storm track structures? Answers to these questions form the core of our current knowledge of storm track dynamics, and provide a context within which both observational and modeling results can be interpreted.

The final member of the triumvirate concerns the ability to model storm tracks, and is treated in section 4. From a certain perspective, modeling provides the ultimate measure of our understanding of the physical processes relevant in the observed storm track structures, under the provisio ‘‘simulation is understanding.’’ While it is well known that early GCM simulations had difficulty reproducing the overall structure of storm tracks, this was primarily a problem of model resolution, and the current generation of climate models does a reasonable job at reproducing the climatological storm track structures. However, as indicated above, it is the ability of models to simulate the annual cycle of storm tracks as well as longer timescale variability in storm track structures that provides the true test of model fidelity; the fidelity and ability of models to simulate these phenomena are important and ongoing issues in the field, and will certainly remain the focus of research for some time to come. The review concludes with a discussion of the current state of the field, and a look toward the future regarding what the next decade will bring in the study of storm track dynamics.

2. Observations

a. Climatological structure

As indicated in the introduction, it is convenient to define storm tracks based upon bandpass transient variances (or covariances). Generally speaking, no matter which transient quantity is chosen to represent the storm track, eddy amplitudes during the Northern Hemisphere cool season are maximal over a band running across the midlatitudes, extending roughly from the western North Pacific, across North America and the North Atlantic, into northern Europe. Figure 2 shows that the bandpass standard deviations and covariances tend to have a more pronounced minima in some variables, most notably upper-tropospheric geopotential height (Fig. 2a) and lower-tropospheric heat flux (Fig. 2d), than in others, for example, meridional velocity (Fig. 2b). Usually, two peaks in the variance are found—over the eastern Pacific and North Atlantic—respectively marking the Pacific and Atlantic storm tracks. For other variances and covariances, refer to Blackmon et al. (1977) and Lau (1978, 1979). While not directly related to the current review, Trenberth (1991) showed the relationship between the different variables one can use to define storm tracks for the Southern Hemisphere, which generally applies to the Northern Hemisphere as well.

The structure of the transient eddy fluxes of heat and momentum strongly implicates baroclinic instability as the ultimate mechanism generating the transients that compose the storm tracks. Strong baroclinic conversion of the available potential energy from the time mean flow to the transients, marked by the downgradient heat fluxes off the east coasts of Asia and North America (Fig. 2d; see also Blackmon et al. 1977), coincides with the rapid growth in transient variance in the downstream (eastward) direction. This baroclinic growth peaks where the baroclinicity of the flow is largest, as measured, for example, by the Eady parameter (Lindzen and Farrell 1980):

$$\Gamma = \frac{f}{N}\frac{\partial \overline{u}}{\partial z}, \tag{1}$$

where $\overline{u}$ is the time mean zonal wind, f the Coriolis parameter, and N the static stability. The Eady parameter Γ has its maximal value at the core of the tropospheric jets, which as shown in Fig. 3b also lie just off the east coasts of Asia and North America.

It is well known from life cycles of nonlinear baroclinic waves on zonally homogeneous flows (see the review by Pierrehumbert and Swanson 1995) that downgradient heat fluxes occur during the growth stage of such waves in the lower troposphere, and are followed by large meridional fluxes of zonal momentum in the upper troposphere. Within the observed storm tracks, momentum fluxes tend to peak downstream of the peak heat fluxes, as seen by comparing Figs. 2d and 2e. This observation was originally interpreted to be a consequence of eastward phase propagation of baroclinic waves that are generally growing over the storm track entrance region and decaying over the storm track exit. However, it is now appreciated that storm tracks comprise an ensemble of wave packets with wave growth and decay occurring over all portions of the storm track. As such, the zonal variations in eddy heat and momen-

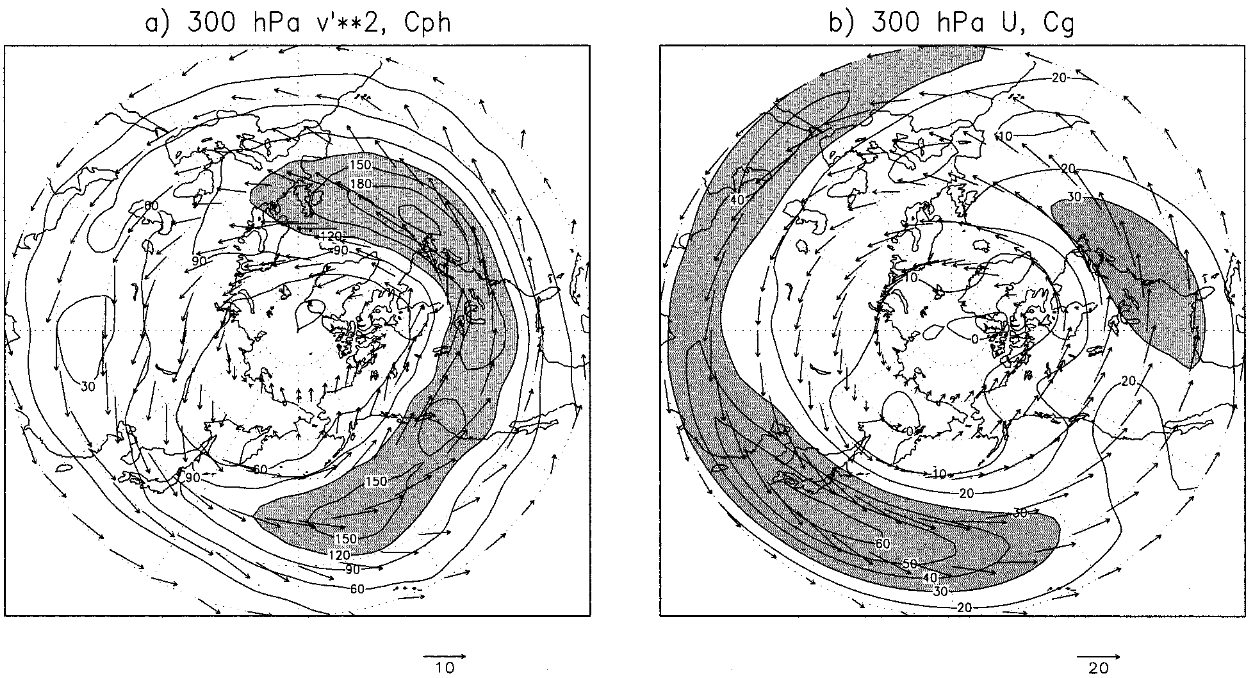


FIG. 3. (a) Bandpass 300-hPa v variance (contour 30 $m^2\ s^{-2}$) with phase velocity shown by arrows. (b) 300-hPa zonal wind (contour 10 $m\ s^{-1}$) with group velocity shown by arrows. The phase and group velocities are computed using time-lagged correlation statistics as described in Chang and Yu (1999). The velocity scales in $m\ s^{-1}$ appear below each plot.

tum fluxes reflect the modulation of the wave packets by the zonally varying background flow, more than the different stages in the life cycle of baroclinic waves (see Chang and Orlanski 1993). Nevertheless, *in the zonal mean*, the traditional paradigm of baroclinic growth and barotropic decay does hold. The implication is that this paradigm is appropriate only to the extent that a *collective* effect of the eddies on the zonal mean flow is concerned. We would like to note that even though the nonlinear life cycle paradigm may not be directly applicable to describing the majority of cases of the evolution of individual cyclones and troughs, aspects of it do strongly illuminate what we see when some individual weather systems grow mature and decay (e.g., Thorncroft et al. 1993).

Waves within the storm tracks generally propagate eastward. Figure 3a shows the vector phase velocities of the transients that compose the tracks, computed using one-point lagged correlation statistics (e.g., Wallace et al. 1988; Lim and Wallace 1991; Chang and Yu 1999). The phase propagation is primarily zonal, and the preferred zones of propagation with the largest time-lagged phase coherence, or ''baroclinic waveguides,'' more or less coincide with the storm tracks themselves (Wallace et al. 1988). More significantly, storm track transients also behave as wave packets, characterized by a well-developed group velocity property (Lee and Held 1993; Chang 1993). Chang and Yu (1999) diagnose this group property by performing one-point lagged correlations on the envelopes of transient activity. The resultant vector group velocities are shown in Fig. 3b, along with the distribution of the 300-hPa zonal wind. The group velocity coincides with the general direction of the storm track axes, while the group speed is on the order of (but somewhat less than) the speed of the jet itself. This contrasts with the phase speed of the transients, which has a value more characteristic of a lower-tropospheric steering level, consistent with the transients being forced from a lower-tropospheric critical layer, that is, arising from baroclinic instability. This wave propagation characteristic suggests that the high bandpass variability not only marks regions of high cyclone frequency, but also is oriented along the paths of phase and group propagation of synoptic transients. As such, the description ''storm tracks'' is truly apt.

b. Seasonal variability

The association of storm tracks with midlatitude baroclinic zones suggests that storm tracks, like the zonal mean pole-to-equator temperature gradient, will experience a pronounced annual cycle. However, that cycle holds some surprises. Nakamura (1992) provides the most complete picture of the seasonal variations of the Northern Hemisphere storm tracks, and also raises a number of possible mechanisms for the ''midwinter minimum'' of the Pacific storm track. As shown in Figs. 4a,b, both the Atlantic and Pacific storm tracks, here indicated by an average in 300-hPa bandpass meridional velocity variance over the indicated longitude band,

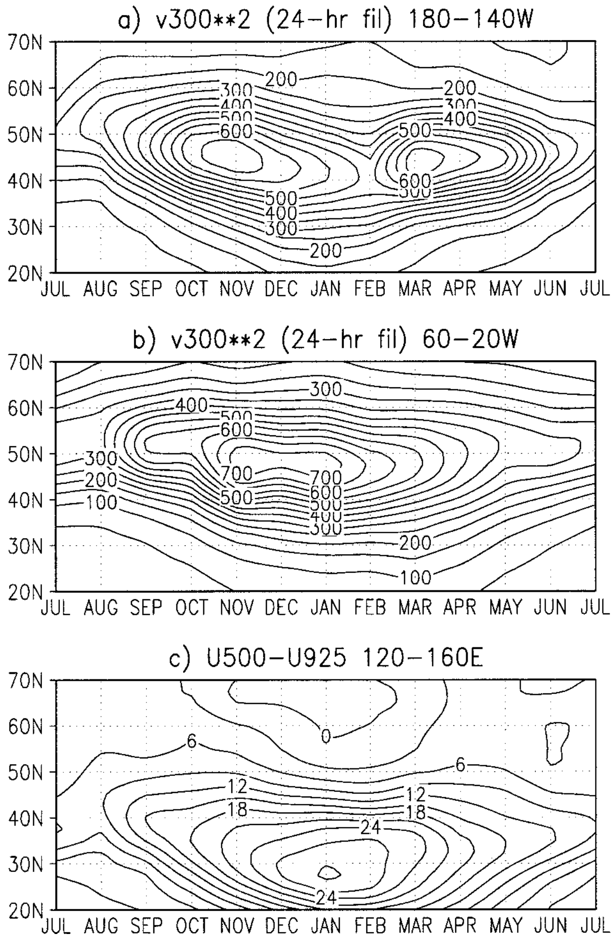


FIG. 4. (a) Lat–time sections showing the seasonal march of baroclinic wave amplitude in 300-hPa v^2, filtered using the 24-h difference filter described in Wallace et al. (1988). The variance has been averaged over the lon band 180°–140°W (contour 50 $m^2 s^{-2}$). (b) As in (a) except for the lon band 60°–20°W. (c) As in (a), except for vertical shear of the zonal wind between 500- and 925-hPa levels, averaged over the lon band 120°–160°E (contour 3 m s^{-1}). The plots are patterned after Figs. 2 and 6 of Nakamura (1992), but a different parameter is shown here for comparison.

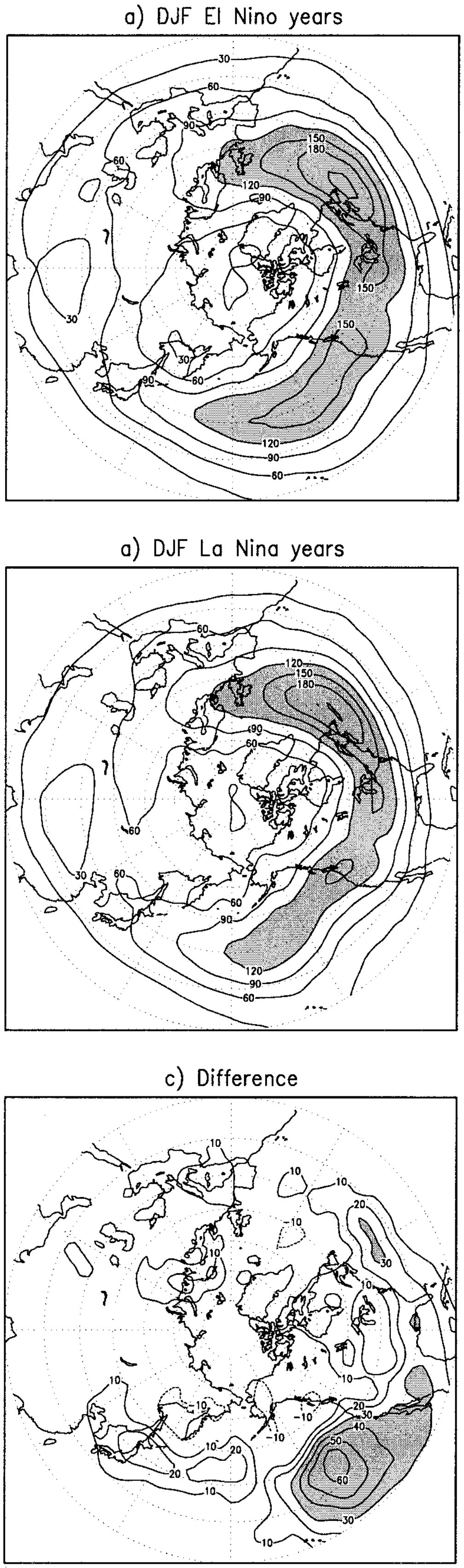
shift equatorward in step with the jet stream from fall to midwinter, and then migrate poleward after January. However, while the Atlantic storm track attains its maximum amplitude around midwinter, the Pacific storm track is strongest during fall and spring, and shows a minimum in eddy amplitude during midwinter. This minimum is not only found in upper-tropospheric velocity and geopotential height variance fields, but also in sea level pressure variations, transient eddy heat fluxes, as well as eddy energy. The fact that the atmospheric condition is deemed to be more unstable during midwinter, indicated in Fig. 4c by greater zonal wind shear over the Pacific during midwinter than in fall or spring, makes this finding quite surprising. Christoph et al. (1997) confirm this result using a longer period of analyzed observational data (1946–89), and also found similar variations in simulations using the Hamburg version of the European Centre atmospheric GCM (ECHAM3) T42 model.

Recently, Nakamura and Izumi (1999) showed that the midwinter suppression is modulated by interannual and decadal variations of the midwinter Pacific storm track intensity, such that during the late 1980s and early 1990s, when the Pacific storm track is stronger than its climatological average during midwinter, the suppression is not as apparent as during the 1970s and early 1980s. While the exact mechanisms responsible for this midwinter suppression in the Pacific are still being actively debated, several factors that in principle may lead to such variations have been proposed, and these will be discussed in later sections.

c. *Interannual variability*

Lau (1988) examined the month-to-month variations of the wintertime storm tracks. One of the first two leading modes corresponds to fluctuation of the storm track intensity, while the other leading mode corresponds to a north–south shift of the storm tracks. Lau (1988) also showed that these leading patterns of storm track variability are linked to larger-scale, lower-frequency variability in the monthly averaged flow. Metz (1989) used canonical correlation analysis to investigate the relationship between changes in atmospheric low-frequency variabilities and eddy flux convergences. Metz found two robust canonical modes: one apparently related to Pacific blocking, and the other to a regional jet anomaly over the Atlantic. Both of these studies established that storm track variances and covariances are closely related to mean flow changes.

On interannual timescales, storm tracks change in response to the El Niño–Southern Oscillation (ENSO) cycle. Figures 5a,b show that the Pacific storm track shifts equatorward and downstream during El Niño years (see also Trenberth and Hurrell 1994; Straus and Shukla 1997; Zhang and Held 1999), apparently in response to local enhancement of the Hadley circulation over the eastern Pacific (Bjerknes 1966, 1969), while La Niña events mark opposite shifts. Although the tropical SST-induced heating may be the ultimate driver behind these structural changes, attributing all of these storm track structural changes to the direct tropical forcing is improper. As stated earlier, Held et al. (1989) suggested that the direct midlatitude stationary wave response to tropical SST-induced heating is weak, and eddy forcing associated with changes in the storm tracks plays an important role in setting up the extratropical response to ENSO. Because the storm track eddies are in turn organized by the stationary wave (Branstator 1995), nonlinear interaction among the tropical heating, storm track eddies, and the midlatitude stationary wave must be accounted for to make correct attributions of the storm track structural changes.



d. *Decadal variability*

On decadal timescales, recent observational analyses suggest that the amplitude of the storm tracks vary significantly on interdecadal timescales. Nakamura and Izumi (1999) point out that the midwinter Pacific storm track was much stronger during the late 1980s and early 1990s than during the 1970s and early 1980s. Meanwhile, Ebisuzaki and Chelliah (1998) showed evidence that the Atlantic storm track was much weaker during the 1960s than in recent decades. Recently, Geng and Sugi (2001) and Graham and Diaz (2001) have shown that the frequency and intensity of extreme cyclones over both the Atlantic and Pacific basins have increased over the second half of the twentieth century. Chang and Fu (2002) conducted an empirical orthogonal function (EOF) analysis on the interannual storm track variations, and found that the leading mode represents a simultaneous change in the intensity of both storm tracks, and the principal component time series displays marked interdecadal variability, with the storm tracks nearly 40% stronger during the 1990s than during the 1960s. An example is shown in Fig. 6, where the decadal mean December–January–February (DJF) 300-hPa bandpass eddy meridional velocity variances for the decade 1989/90–1998/99 when the storm tracks were strong, and the 1961/62–1970/71 when the storm tracks were weak, together with their differences, are plotted. Similar, but smaller, differences in variability are found in radiosonde observations along the storm tracks over areas (apart from Japan) where such observations are available. Chang and Fu (2002) also showed that even if storm track variations linearly dependent on the Arctic Oscillation (Thompson and Wallace 1998, 2000; Thompson et al. 2000), and ENSO-like interdecadal variabilities (Zhang et al. 1997) are removed, the residual storm track data still show significant interdecadal variability. The implications of such large interdecadal variations, as well as what causes such variabilities, are questions that remain unanswered.

3. Processes

a. *Baroclinic source*

From an energetics point of view, the ultimate source of energy for the zonally asymmetric part of the flow is baroclinic generation (e.g., Oort and Peixoto 1983; Peixoto and Oort 1992). The main physical mechanism involves release of mean flow available potential energy associated with poleward and upward motion of warm air, along with equatorward and downward motion of

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FIG. 5. (a) Bandpass 300-hPa v variance averaged over nine El Niño winters (contour 30 m^2 s^{-2}). (b) As in (a) but for nine La Niña winters. (c) Difference between El Niño and La Niña winters (contour 10 m^2 s^{-2}).

cold air within baroclinic waves and frontal cyclones (e.g., Browning and Roberts 1994). This interpretation is consistent with the theoretical expectation that midlatitude cyclogenesis occurs as a result of baroclinic instability (Charney 1947; Eady 1949). While there are doubts that the majority of cyclones grow from infinitesimal perturbations (e.g., Farrell 1984, 1985), the energy source for nonmodal cyclone growth is still baroclinic, and such growth can be incorporated into a more generalized paradigm of baroclinic instability (e.g., Hoskins et al. 1985; Pierrehumbert and Swanson 1995).

The association of the baroclinic source (and hence, the storm tracks themselves) with strong meridional temperature gradients (baroclinicity) raises the question of what determines the zonal variation in baroclinicity. This question is in reality very deep, as at first glance it is not apparent that storm tracks should be self-maintaining. The same eddies that compose the storm tracks act to mix temperature in the lower troposphere, which in principle should act to destroy the very meridional temperature gradients upon which the existence of the storm tracks depends. Hoskins and Valdes (1990) have investigated this issue using a linear stationary wave model, and suggest that a threefold process allows for the enhanced baroclinicity over the storm track entrance regions in the Northern Hemisphere. First, storm track eddies in general are most vigorous *downstream* of the regions of peak baroclinicity, and the actual mixing of temperature by eddies is relatively benign where the baroclinicity is largest. Further, the enhanced baroclinicity itself is actively maintained by condensational heating over the storm track entrance region. Since the diabatic heating maxima are caused by the eddies themselves, Hoskins and Valdes argue that in a certain sense, the storm tracks are self-maintaining. Finally, the wind stress of the low-level flow induced by the eddies acts to drive the warm western boundary currents in the ocean, which establishes zones of high baroclinicity due to land–sea temperature contrasts.

An alternative perspective on the maintenance of baroclinicity over the storm track entrance regions is given by Broccoli and Manabe (1992), who showed that in general circulation model experiments without mountains, the stationary waves are considerably weaker, and the storm tracks more zonally symmetric, even in the presence of land–ocean contrast. Supporting this result, Lee and Mak (1996) showed that, in a dry nonlinear model driven by relaxation to the observed winter zonal mean temperature distribution, enhanced baroclinicity over the storm track entrance region could be maintained just by stationary waves induced by mountains

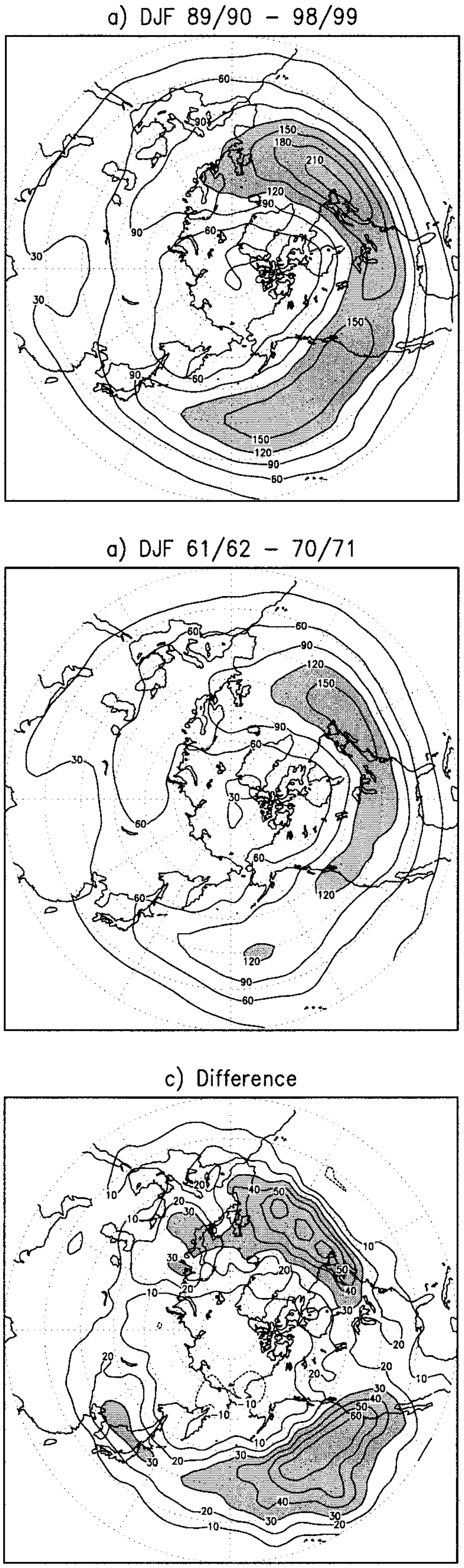

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FIG. 6. (a) Bandpass DJF 300-hPa v variance averaged over the decade 1989/90–1998/99 (contour 30 $m^2 s^{-2}$). (b) As in (a) but for the decade 1961/62–1970/71. (c) Difference between (a) and (b) (contour 10 $m^2 s^{-2}$).

alone, without the need for diabatic heat sources near the storm track entrance regions. The above results lead us to conclude that planetary stationary waves are crucial for organizing the storm tracks, and that the storm tracks are not completely self-maintained. Whether stationary waves are primarily forced by diabatic heating or by orography is still an actively debated question, and is reviewed in this issue by Held et al. (2002).

While geographical maxima in the baroclinic conversion are associated with the geographical maxima in the baroclinicity, it does not follow that baroclinic conversion and the conversion from the eddy potential energy (PE) to eddy kinetic energy (KE) are necessarily stronger for larger baroclinicity. As discussed in section 2, Nakamura (1992) shows that the Pacific storm track is weaker during the midwinter than during the fall or spring, even though the baroclinicity is strongest during the midwinter. Zhang and Held (1999) showed that the same is true in GCM simulations for interannual variations in midwinter. Chang (2001a) provides evidence that a number of factors, including changes in eddy structure, eddy residence times over the baroclinic region, and differences in diabatic heating contributions to eddy energy generation, influence the ability of eddies to tap into the baroclinicity of the large-scale flow. The links between baroclinicity, baroclinic conversion, and eddy amplitudes are complex, and will require further study to clarify.

b. Linear modal instabilities

Following the success of the Charney (1947) and Eady (1949) models at predicting the observed scale and structure of linear disturbances that can efficiently tap the baroclinic source of energy and grow on zonally symmetric flows, it is only natural to inquire whether linear disturbances to more realistic zonally varying basic-state flows might replicate the structure of storm tracks. Indeed, early results by Fredriksen (1983) along these lines do reveal exponentially growing modes that at least qualitatively resemble storm tracks. However, there are reasons to doubt such an approach. Point correlations within the storm tracks are in general localized to an individual storm track, and specifically, do not span the globe. Thus, if a linear instability is indeed responsible for storm track structure, it must be *local*, with all eddy properties, that is, growth rates, phase speeds, scale selection, determined as a function of the local properties of the background flow. This distinction was first noted by Pierrehumbert (1984), who emphasized the contrast between local modes and *global modes* that require a reentrant domain for their existence and in general are sensitive to flow properties far away from the storm track region.

As noted by Pierrehumbert (1984), the distinction between local and global modes is intimately related to the concept of absolute and convective instabilities to zonally symmetric flows in an unbounded zonal domain [see Huerre and Monkewitz (1990) for a review, and Pierrehumbert and Swanson (1995) for applications to baroclinic systems]. In an unbounded domain, a system is said to be absolutely unstable if an initially localized disturbance ultimately leads to disturbance growth throughout the domain. In this situation, the system can be considered closed in the sense that it will generate its own disturbances in the absence of any external noise. In the zonally varying problem, this situation can lead to a geographically fixed, temporally amplifying local mode that is independent of flow conditions far removed from the seat of instability. The self-contained nature of considering a storm track as a single local mode is theoretically appealing, as it is a natural extension of exponentially amplifying normal modes familiar from the Charney and Eady models.

In contrast, in a convectively unstable flow an initially localized disturbance will not fill the domain, but rather will be advected downstream faster than it locally grows. In such a situation, provided one waits long enough, any initially localized disturbance will eventually pass by a given point, leaving undisturbed flow in its wake. Such a system is considered open, as disturbance properties at any given point are tied to the nature of an externally imposed excitation.

In the Wentzel–Kramers–Brillouin (WKB) limit, the criteria for the existence of a local instability on a zonally varying flow are identical to the criteria for absolute instability applied to individual zonal slices of that flow. However, the necessary criteria for absolute instability, and hence for the existence of local modes, do not appear to be met in the midlatitude atmosphere. Pierrehumbert (1986) showed that the Charney model only possesses absolute instabilities if easterlies are found at the surface, and this situation is only exacerbated for more realistic flows with surface damping (Lin and Pierrehumbert 1993).

In the absence of local modes, linear theory also allows for the possibility that storm tracks result from the constructive/destructive interference between several global modes, which overcomes the phase coherence issue noted above, while still potentially yielding a localized storm track. As Pierrehumbert (1986) points out, this possibility recognizes a great deal of connectivity between storm tracks, with each track actively seeding its downstream neighbors, presumably in the form of upper-level troughs circumnavigating the globe. In fact, Whitaker and Barcilon (1992) demonstrate with an idealized model that a linear superposition of several unstable global modes can yield a localized structure that resembles the observed storm tracks. Lee (1995a) shows that when large-scale mountains and associated baroclinic zones are sufficiently close to each other, global unstable modes that resemble observed storm track structures are spontaneously generated. Further, realizing that nonmodal interactions can lead to disturbance growth rates over short time periods that greatly exceed the growth rates of normal modes, Farrell (1982, 1985)

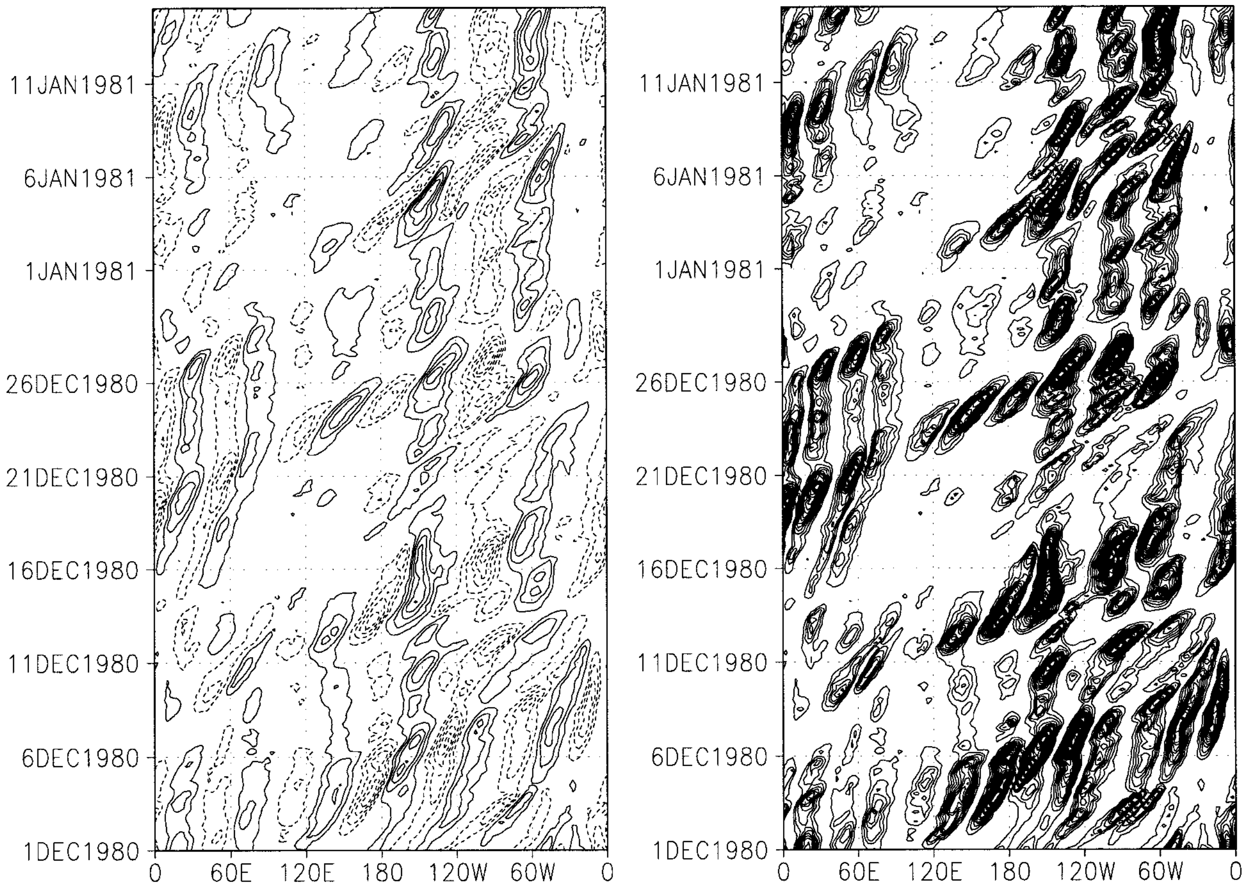

FIG. 7. Hovmoeller (lon–time) diagram of 300-hPa v and v^2, unfiltered except for removal of seasonal mean, for the period 1 Dec 1980 to 14 Jan 1981 (contours 10 m s^{-1} and 100 m^2 s^{-2}). This figure is similar to Fig. 2 in Chang (1993), except the the plots are for a different period of time, and v has been averaged over a 20° lat band centered around the upper-tropospheric waveguide as defined in Chang and Yu (1999) instead of being averaged over 30°–60°N.

argues that exponential growth is irrelevant. Instead, the background noise can be incorporated into a stochastic model, forcing a system, which in reality is stable, to exponential disturbance growth. Such an approach has recently shown success in reproducing storm track structure based entirely on dynamics linearized about the time mean flow (Whitaker and Sardeshmukh 1998; Zhang and Held 1999), and appear to provide the logical next step in linear investigations of storm track structure. Further details on such an approach are given in section 4.

c. Downstream development

The fact that a storm track is not simply a single mode leads us to investigate other processes that might yield the observed storm track structures. One important realization over the past decade has been that storm tracks, in contrast to being a single time-independent structure, in reality are more aptly characterized as a composite resulting from the modulation of individual nonlinear baroclinic wave packets. Coherent nonlinear baroclinic wave packets were first noted in the Southern Hemisphere and in simple models by Lee and Held (1993). The study of Chang (1993) more strongly supports this viewpoint by illustrating the prevalence of such wave packets over the NH storm track regions. Figure 7 shows a Hovmoeller diagram of meridional wind on the 300-hPa surface for 45 days during the winter season of 1980/81; the phase evolution of the individual synoptic eddies is apparent, characterized by a typical phase speed of 10 m s^{-1}. However, most remarkable is the presence of a number of coherent wave packets that survive one or more circuits around the globe. The synoptic eddies that compose these packets amplify as they propagate over the zones of high baroclinicity in the western Pacific and western Atlantic, but the continual group character of the packets throughout their transit is the most striking aspect.

The fact that the synoptic eddies that compose nonlinear packets deviate from the ''traditional'' synoptic eddy life cycle of Simmons and Hoskins (1978) is the key to understanding this group behavior. In the traditional synoptic eddy life cycle, baroclinic conversion at the surface is followed by disturbance growth aloft, with disturbance decay following soon after due to wave breaking on jet flank critical layers and consequent barotropic energy conversions from eddy kinetic energy to mean flow kinetic energy. In nonlinear wave packets, *individual* synoptic eddies decay primarily by transferring their energy to neighboring eddies downstream (Orlanski and Katzfey 1991; Chang 1993; Orlanski and Chang 1993), a process called *downstream development* (see Namias and Clapp 1944; Cressman 1948; Yeh 1949; Simmons and Hoskins 1979). This gives the packets their coherence, as individual eddies that compose the packet primarily transfer energy to their downstream neighbors, with the energy lost to barotropic conversion and dissipation mostly balanced by gains from baroclinic conversion, such that *upon zonal average*, the eddy activity resembles the traditional eddy life cycle (Chang 2000).

In the storm track entrance region, upper-tropospheric disturbances associated with nonlinear baroclinic wave packets initiate type-B cyclogenesis, tapping into the baroclinicity and amplifying. As such, in these regions, eddy growth in general is dominated by baroclinic conversion (Plumb 1986). However, as one moves farther downstream within the storm track, downstream development becomes increasingly more important to the development of individual eddies. In these regions, eddy growth primarily occurs from ''recycled'' energy from eddies over the upstream end of the track. In the storm track exit region, energy lost by eddies is enhanced by barotropic deformation (see also section 3d). Thus, to a certain extent, the collective behavior of the eddies within the storm track is similar to the traditional eddy life cycle paradigm in that the amplitude of eddies are enhanced baroclinically in the upstream and suppressed barotropically in the downstream. The differences between the two paradigms that we wish to stress here are the following: 1) it is not a single eddy that undergoes baroclinic growth in the upstream and barotropic decay in the downstream, but embedded between the upstream and downstream are eddies that recycle their energy toward downstream eddies; 2) eddies undergo growth and decay over both the storm track entrance and exit regions; and 3) this downstream development is able to extend storm track from source regions within zones of strong baroclinicity into regions unfavorable for baroclinic conversion (Chang and Orlanski 1993).

d. Barotropic effects

The realization that large amplitude baroclinic waves are not confined within the two oceanic storm track regions, but rather, in the guise of coherent nonlinear wave packets, span the entire midlatitudes, naturally suggests further inquiry into what mechanisms other than zonally varying baroclinicity serve to localize the storm tracks. Some of these processes are relatively straightforward, such as zonally varying surface roughness that results in stronger eddy decay in certain regions (Chang and Orlanski 1993). However, recent studies have shown that barotropic effects resulting from a zonally varying background flow also act to localize storm tracks.

Taking coherent nonlinear baroclinic wave packets in the upper troposphere as given, one may pose a question as to how the amplitude and length of the wave packet change due to the barotropic component of the zonally varying background flow. With a highly idealized barotropic model, Lee (1995b) showed that zonal variation of the zonal wind alone is capable of modulating high-frequency ''storm track eddies,'' resulting in zonally localized storm tracks. Specifically, provided that the eddies are meridionally elongated, that is, that their zonal wavenumber k is much larger than their meridional wavenumber l, and that the advection of eddy vorticity by the background zonal wind is much greater than the advection of background vorticity by the eddy winds, that is, $k^2 \gg |\partial Q/\partial y|/U$, where Q is the background vorticity and U the zonal flow, then we have

$$\langle \psi^2 \rangle^{1/2} \approx k(x)^{-2} \langle \zeta^2 \rangle^{1/2} \approx [U(x)/\omega]^2 \langle \zeta^2 \rangle^{1/2} \quad \text{and} \tag{2}$$

$$\langle v^2 \rangle^{1/2} \approx k(x)^{-1} \langle \zeta^2 \rangle^{1/2} \approx [U(x)/\omega] \langle \zeta^2 \rangle^{1/2}, \tag{3}$$

where ψ, ζ, and v are the disturbance streamfunction, vorticity, and meridional velocity, respectively; the angled brackets denote average over phase; and quantities that vary in the zonal direction are denoted as functions of x. Under the assumptions stated above, the local dispersion relation is simply $\omega = Uk$; since this is a linear problem, ω is fixed, and as such k is inversely proportional to U, that is, eddies become more anisotropic as they propagate into regions of weak zonal flow. Since the above assumptions also imply that the eddy enstrophy, $\langle \zeta^2 \rangle$, is independent of x, it follows from (2) and (3) that the rms disturbance streamfunction and meridional velocity must both decrease when disturbances propagate into regions of weak zonal flow. In other words, changes in eddy anisotropy caused by changes in U modulate eddy amplitude when measured using nonconservative quantities such as disturbance streamfunction variance or disturbance kinetic energy. Because storm tracks usually terminate near the jet minimum, the implication is that the localization of the storm tracks is, at least in part, due to the fact that they are described by nonconservative quantities.

The above results are refined by Swanson et al. (1997). Once again adopting a barotropic framework, they examined a Rossby wave packet propagating along a zonally varying background flow whose potential vorticity (PV) distribution is piecewise constant, rather than constant as in Lee (1995b). Rossby waves propagating along a PV discontinuity are meridionally trapped [see appendix A in Swanson et al. (1997)], and thus do not disperse in the meridional direction. This property allows the waves to propagate in the zonal direction without irreversibly losing their energy through meridional dispersion. It is also for this reason that a constant PV distribution was chosen by Lee (1995b). While neither uniform nor piecewise constant PV distributions accurately reflect the observed PV distribution in the storm track regions, the latter is closer to reality as the former does not support Rossby waves. In addition, the nonzero PV gradient associated with the PV discontinuity allows Swanson et al. (1997) to interpret the ''barotropic modulation'' of disturbances in the context of conservation of wave action (see Andrews and McIntyre 1978).

As stated above, eddy anisotropy is central to the barotropic amplitude modulation of nonconservative quantities. However, edge waves propagating along a single barotropic PV jump (or contour) are inherently isotropic. Thus, Swanson et al. (1997) also consider Rossby waves propagating along two PV contours; the anisotropy can be controlled by changing the distance between the two contours. With these contour models, they found that the amplitude of the eddy streamfunction scales as $\langle \psi^2 |_{\mathrm{contour}} \rangle^{1/2} \propto U$, a rather modest amplitude modulation compared to that given by (2) above.

Both Lee (1995b) and Swanson et al. (1997) found that if the local minimum value of the background zonal wind is sufficiently weak, nonlinearities can lead to an irreversible loss of wave energy. Frequent wave breaking events observed in regions of weak upper-tropospheric flow suggest that such an irreversible loss of energy is realistic, and as such plays an important role in storm track dynamics.

The dissipation of the upper-tropospheric storm track eddies, either due to wave breaking at that level or to surface friction, must ultimately be balanced by a baroclinic source. This statement can be summarized (see Swanson et al. 1997) in terms of a conceptually useful, heuristic wave action equation for the upper troposphere. Restating (5.1) of their paper,

$$\frac{\partial \mathcal{W}}{\partial T} = -\frac{\partial}{\partial X}(c_g \mathcal{W}) - \text{dissipation} + \text{baroclinic source}, \qquad (4)$$

where $\mathcal{W}$ is wave action, c_g is the local group velocity, and X and T, respectively, denote slowly varying space and time variables corresponding to the basic state. In order to close the above problem, the baroclinic source term must be represented in terms of upper-tropospheric wave action (or any other appropriate conserved quantity) and mean flow parameters. There is some observational basis for parameterizing the baroclinic wave source in terms of upper-tropospheric wave action, as baroclinic growth of upper-tropospheric waves is often triggered by preexisting disturbances in the upper troposphere (Pettersen and Smebye 1971; Whitaker et al. 1988; Uccellini 1986). However, it is not well understood precisely how the baroclinic source term is related to the upper-tropospheric wave action and mean flow parameters, nor how the former can be parameterized in terms of the latter. Finally, it appears difficult to quantitatively test barotropic modulation in the atmosphere. Although there has been such an attempt (Lee 2000), the ambiguity in formulating conserved quantities for finite amplitude storm track disturbances and the existence of large regions of nearly vanishing potential vorticity gradients present formidable obstacles to such investigations.

As alluded to earlier, we emphasize that the theory reviewed in this section does not address the growth of ''baroclinic eddies''; instead, it only considers barotropic effects and takes baroclinic eddies as given. However, a comprehensive barotropic effect in a baroclinic atmosphere is far more complex, as barotropic shear can play an important role in baroclinic growth itself. Studies of baroclinic instability in the presence of horizontal shear (more precisely, shearing deformation) show that baroclinic growth can be severely limited by this horizontal shear (James 1987). This effect is known as the ''barotropic governor'' mechanism. Whitaker and Dole (1995) find that the model storm track maximum occurs at the jet entrance, just downstream of the point of the minimum horizontal deformation, when the equilibrium state in their model is configured so that baroclinicity is zonally uniform while the horizontal deformation varies in the zonal direction. While zonal variation of wave breaking and associated irreversible wave energy loss could explain this result, as they speculate, the barotropic governor mechanism is certainly a viable explanation for their model behavior.

e. Effects of diabatic heating

As discussed above, there are indications that condensational heating associated with the midlatitude storm track regions helps to maintain the enhanced baroclinicity in the storm track entrance regions (Hoskins and Valdes 1990). However, the effect of diabatic heating is not limited to this. Latent heat release due to condensation in rising air over the warm sector of cyclones in general acts as an additional energy source to fuel the development of those eddies, and numerous theoretical (e.g., Mak 1982; Emmanual et al. 1987; Fantini 1995), numerical, and diagnostic (e.g., Gutowski et al. 1992; Davis et al. 1993; Reed et al. 1992; etc.) studies have shown that the growth rate and amplitude of baroclinic waves are enhanced by condensational heating. On the other hand, surface sensible heat fluxes, especially over the oceans, strongly damp temperature perturbations near the surface (Swanson and Pierrehumbert 1997), and as such generally act as an energy sink (Hall and Sardeshmukh 1998; Branscome et al. 1989). However, in certain situations surface sensible heat fluxes could induce strongly unstable, shallow short waves (Mak 1998).

In spite of the interest in the role of diabatic heating on the evolution of individual storms, few studies have explicitly dealt with the effects it has on aggregate storm tracks. One reason for this is the lack of reliable global observation of diabatic heating rates. Black (1998) used the Goddard Earth Observing System (GEOS-1) reanalysis assimilated data, together with the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis data, to estimate diabatic effects on the eddy enstrophy budget at the 400-hPa level, and found that condensational heating generally acts as a source of upper-tropospheric enstrophy over the storm track regions. Over the North Pacific, this contribution is locally of the same order as the conversion from the mean flow.

The rate of generation of transient eddy available potential energy (EAPE) based on the midwinter heating rates diagnosed from the NCEP–NCAR reanalysis provides an alternative perspective on the role of diabatic processes to storm track dynamics. EAPE generation is proportional to the product of the eddy temperature perturbation and diabatic heating rate; for the purposes here, the heating rates, derived from the NCEP–NCAR reanalysis products, are averages from 6-h forecasts starting from the reanalysis grids. Given the inevitable uncertainty with such a procedure, the results here provide only a qualitative representation of the role of diabatic heating. However, an estimation of the total EAPE generation rates based on EAPE budget residuals (not shown) gives similar results.

For interpretation, the diabatic heating is separated into three processes: moist heating, including large-scale condensation and convective heating; sensible heating associated with surface sensible heat fluxes; and radiative heating. Figure 8a shows that moist heating is maximum along the storm track in the Pacific and Atlantic, with maximum generation rates as large as 40 m^2 s^{-3} over the Atlantic storm track entrance region. This heating is dominated by large-scale condensation in the warm sector of incipient cyclones, with deep convection actually giving a negative contribution, as it generally occurs in cold air trailing the cold front. The locations of this EAPE source agree quite well with the enstrophy source due to latent heating by Black (1998). Surface sensible heat fluxes, shown in Fig. 8b, provide a strongly negative contribution along the continental east coasts, consistent with strong thermal damping of cold continental air by the underlying ocean surface. The contribution is also negative along a band over the upstream portion of the storm tracks, basically canceling the positive contribution from moist heating over those regions. EAPE generation due to radiative heating (not shown), is an order of magnitude smaller than those due to moist and sensible heating.

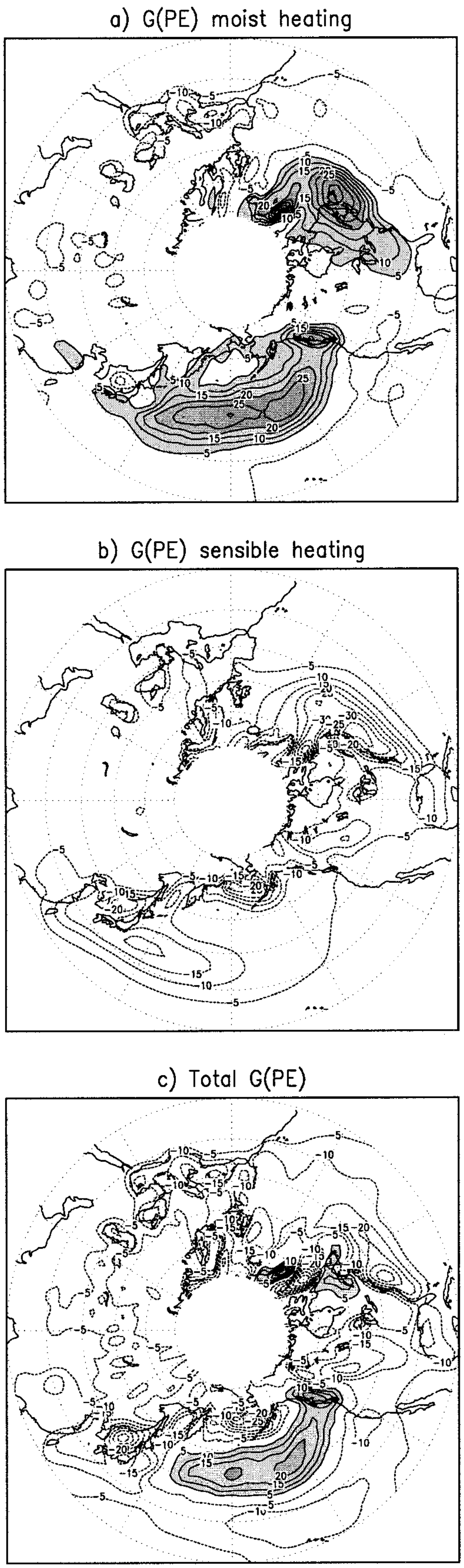

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FIG. 8. Vertically averaged rate of generation of EAPE [G(PE)] due to (a) moist heating, (b) sensible heating, and (c) total (moist plus sensible plus radiative) heating (contour 5 m^2 s^{-2} day^{-1}). The heating rates are from the NCEP–NCAR reanalysis product for the period of Jan 1980–1993. Regions over which G(PE) is greater than 5 and 20 m^2 s^{-2} day^{-1} are shaded.

Except over the middle and exit regions of the Pacific storm track, where EAPE generation due to condensational heating dominates, the total EAPE generation rate, shown in Fig. 8c, is dominated by the effects of the surface sensible heat fluxes. However, Chang (2001a) shows that this may not be true in the spring and fall, due to larger condensational heating during those seasons. This difference could be one factor contributing to the observed midwinter suppression of the Pacific storm track.

The above results suggest that condensational heating acts as a source of EAPE over the storm track regions. However, this may not be the only effect moisture has. Hayashi and Golder (1981) showed that given a fixed zonal mean state, eddies in a moist GCM experiment are much stronger than those in a dry GCM, not only because of an increase in EAPE generation due to diabatic heating, but primarily because baroclinic conversion is strongly enhanced in the presence of diabatic heating in the moist run.

However, the above results should not be taken as implying that eddy amplitudes on a dry earth would necessarily be lower. Held (1993) argues that the ability to transport latent heat makes eddies more efficient at transporting energy poleward, so weaker eddies are required to achieve equilibrium *for fixed forcing.* The GCM studies of Manabe et al. (1965) comparing simulations with and without a hydrological cycle subject to the same solar forcing suggest that eddies in the moist run are less intense than those in the dry run, with the meridional temperature gradient in the dry run larger than its moist counterpart. Since climate changes, past or future, involve changes in the amount of moisture in the atmosphere, the interaction between moisture and the dynamics of storm tracks is a topic ripe for further investigation.

f. The observed midwinter transient eddy energy budget

To place the aforementioned processes important to storm track dynamics in perspective, it is useful to consider the entire transient eddy energy budget of the midwinter (January) storm tracks. As pointed out by Plumb (1983), local energy budgets can be misleading due to the nonuniqueness of the flux and conversion terms. However, given the absence of a conserved wave action quantity for a time mean, zonally asymmetric, forced basic state, diagnosing storm tracks using the transient eddy energy budget at a minimum can be expected to provide certain insights. For example, comparison between the transient eddy energy budget to the budget of an approximately conserved wave action (Plumb 1986) for nonlinear baroclinic wave packets in the much more zonally symmetric situation of the Southern Hemisphere carried out by Chang (2001b) reveals that the results and interpretations from the two budgets are consistent. As long as it is interpreted with care, the local transient eddy energy budget is a useful interpretative tool.

Numerous studies have examined the eddy energy budgets of baroclinic waves and cyclones (e.g., Smith 1969; Kung 1977). Employing the form and interpretation of the budget suggested by Orlanski and Katzfey (1991) yields a budget of the form

$$\frac{\partial E}{\partial t} = \boldsymbol{\nabla} \cdot \overline{(\mathbf{v}E + \mathbf{v}'_a\phi')} + \frac{\alpha_m}{\Theta_m} \frac{\overline{\mathbf{v}'\theta'}}{(\partial\Theta/\partial p)} \cdot \boldsymbol{\nabla}\Theta - \overline{\mathbf{v}' \cdot (\mathbf{v}' \cdot \boldsymbol{\nabla})V_m} - \text{diss} + \text{diab}, \qquad (5)$$

where the subscript "m" denotes mean quantities and the primes indicate deviations therefrom. The notation is otherwise standard. The main difference between this and earlier versions of the transient eddy energy budget is the combination of an ageostrophic geopotential flux with the advective energy flux into a total energy flux (first term on the rhs). This flux is an indicator of eddy propagation (downstream development) by Chang and Orlanski (1994), who noted that in the WKB limit, this flux reduces to the product of the total eddy energy and the group velocity (see Pedlosky 1987; Yeh 1949). The second term on the rhs is baroclinic generation, and the third term is barotropic conversion. The fourth term is mechanical dissipation, and is diagnosed as a residual from the eddy kinetic energy budget. The last term is diabatic generation, which can either be diagnosed as a residual from the EAPE budget, or computed directly if the diabatic heating rates are known (Fig. 8c).

Figure 9a shows the distribution of vertically averaged transient eddy energy [eddy kinetic energy (EKE) plus EAPE], while the distribution of EKE is shown in Fig. 9b. Here, transients are defined as deviations from the monthly mean, as separation into different frequency bands creates many more transfer terms and make the results ambiguous. Nevertheless, the peaks in transient eddy energy are clearly located over the two storm track regions. Baroclinic conversion (Fig. 9c) is located well upstream of the peak eddy energy in the Pacific, but just slightly upstream over the Atlantic, and is much more zonally localized than the eddy energy distribution itself. Barotropic conversions (Fig. 9d) are generally positive over the storm track entrance regions, and negative over the exit regions, with magnitudes generally smaller than the baroclinic conversion but still locally significant. Mechanical dissipation, diagnosed as an EKE budget residual, is negative almost everywhere, and is strongest over the continents.

The contribution from the divergence of the total energy flux, shown in Fig. 9e, is an energy sink over the entrance regions of both the Pacific and Atlantic storm tracks, largely balancing the strong baroclinic generation in those regions. In contrast, this flux divergence is a strong energy source over western North America, the eastern Atlantic, Europe, and parts of Asia. The energy flux clearly acts to redistribute energy from the regions where it is generated into downstream regions

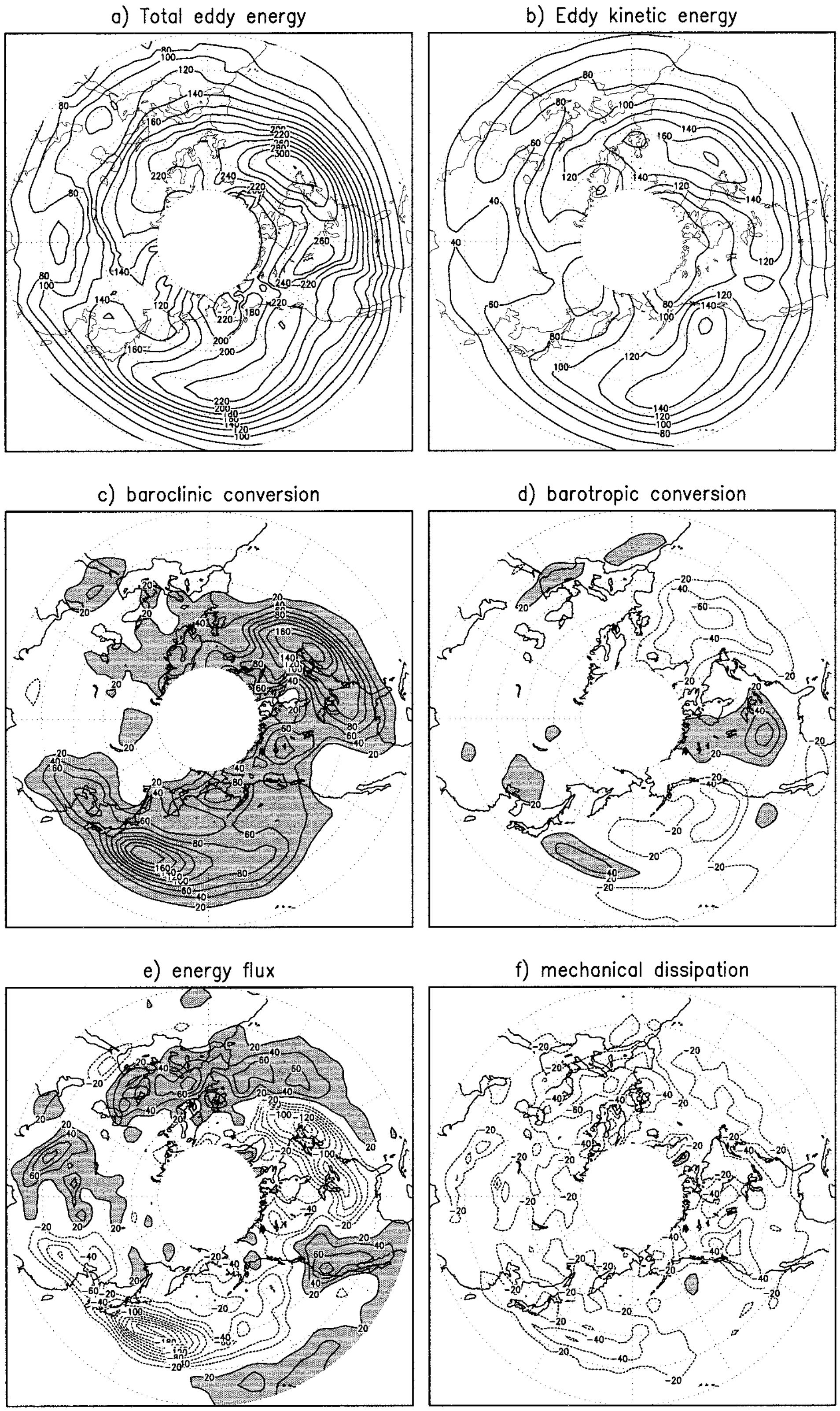

FIG. 9. Vertically averaged distributions of (a) EAPE + EKE, (b) EKE, (c) baroclinic conversion, (d) barotropic conversion, (e) convergence of total energy flux, and (f) mechanical dissipation (computed as a residual in the EKE budget). Contour intervals are 20 $m^2 s^{-2}$ in (a) and (b), and 20 $m^2 s^{-2} day^{-1}$ in (c)–(f). The shading in (c)–(f) denotes regions where the energy conversion rate is greater than 20 $m^2 s^{-2} day^{-1}$.

where baroclinicity is weak, extending the storm track in the zonal direction as previously discussed. Being a flux term, this term averages to nearly zero when a zonal mean is taken, and in the zonal mean, the energetics is dominated by baroclinic growth balanced by barotropic decay and dissipation.

The magnitude of the diabatic energy generation rate in general is much weaker than the baroclinic source (cf. Figs. 8c and 9c), with condensational heating adding about 20% to the baroclinic generation rate over the storm track entrance regions. Similarly, energy dissipation due to surface sensible heat fluxes is dominated by that due to mechanical dissipation, which is primarily due to surface friction.[1]

The eddy energy budget illustrates the importance of all the aforementioned mechanisms; the importance of baroclinic generation of energy over the storm track entrance regions, with much of that energy transported downstream via the process of downstream development. Eddy energy is dissipated over the downstream portion of both storm tracks via barotropic conversion back to mean flow kinetic energy, as well as by surface friction over the continents. The number of important, interacting processes that lead to the observed storm track structures highlights the difficulties involved in modeling storm tracks.

g. *Feedback upon the mean flow*

By virtue of the ability of storm track transients to transport heat and momentum, it is not surprising that the storm tracks are also sites of active eddy–mean flow interaction. This is certainly the case regarding the maintenance of the zonal and/or time mean tropospheric state. For example, insofar as high-pass upgradient momentum fluxes are concentrated in the Northern Hemisphere storm track exit regions, these storm tracks naturally play a vital role in the maintenance of the extratropical westerlies against disspation at the surface (e.g., Held 1975). Regarding the time mean flow, this interaction is most easily understood with reference to the geopotential height tendency exerted by the transients on the planetary-scale flow (Lau and Holopainen 1984; Holopainen 1990). This tendency is primarily forced by the convergence of (i) transient heat fluxes, which are largest in the lower-tropospheric steering level for baroclinic transients (roughly 700 hPa), but whose influence is felt throughout the depth of the troposphere by eddy-induced secondary circulations, locally more intense at the end of the storm tracks (e.g., Blackmon et al. 1977), and (ii) transient vorticity fluxes, typically largest at or near the tropopause.

Lau and Nath (1991) provide the most complete diagnostic of observed storm track eddy–mean flow interactions. In the upper troposphere, they find that while the geopotential tendency due to storm track heat flux anomalies tends to damp a concurrent monthly mean height anomaly, it is overcome by the opposite tendency induced by the anomalies in the vorticity fluxes, so that the total eddy tendency tends to reinforce the observed monthly mean height anomaly. This suggests that at the very least storm track anomalies help to maintain the low-frequency flow anomalies concurrent with the storm track anomalies. However, diagnosing eddy–mean flow interactions using tendencies has its dangers; as noted by Pierrehumbert (1985), the physical processes against which the tendencies act to maintain the flow (e.g., boundary layer friction, radiative damping, etc.) in general are as important to the nature of the response as the tendencies themselves. As such, the effects of these physical processes should be taken into account to properly interpret eddy feedback onto the mean flow.

While many of the effects arising from this vigorous storm track eddy–mean flow interaction lie outside the scope of this review, most notably a detailed description of the role of such transients in the extratropical response to El Niño (e.g., Trenberth et al. 1998), one unifying concept underlying this interaction merits attention. Cai and Mak (1990) and Robinson (1991) both note that there is a ‘‘symbiotic’’ link between storm track anomalies and anomalies in the planetary-scale flow, in that planetary-scale flow anomalies do not occur in isolation, but rather are accompanied by anomalies in synoptic transients and their fluxes. While the turbulent, upscale cascade of energy from deformation-scale synoptic transients to the planetary-scale flow in principle could excite any of a number of planetary-scale flow patterns, Branstator (1995) notes that patterns that organize storm track activity in such a way that they induce a positive feedback onto the large-scale anomaly should be preferred. From the synoptic perspective, such two-way interaction has been widely examined within the context of the formation and maintenance of blocking highs, as pioneered by Berggren et al. (1949) and Rex (1950) (see Colucci 1985; Dole 1986; Nakamura and Wallace 1990, 1993 for a more recent perspective).

As a result, the study of low-frequency variability cannot be divorced from the study of storm tracks; the two scales are inseparable, greatly complicating efforts to construct simple models of the extratropical response to El Niño, increasing greenhouse gas concentrations, and a host of other important problems. This inseparability causes a high premium to be placed on model fidelity across a variety of scales to capture climate variability; this subject is the topic of the next section.

4. Modeling storm tracks

As stated in the introduction, perhaps the ultimate test of understanding storm tracks is whether these structures can be successfully modeled, including successful re-

[1] A momentum damping timescale of 1/2 to 1 day in the planetary boundary layer (e.g., Klinker and Sardeshmukh 1992) is sufficient to entirely account for the magnitude of the residue shown in Fig. 9f.

production of the annual to interannual variability outlined in section 2. There are (at least) two approaches to this problem: full atmospheric general circulation models (AGCMs) that incorporate all physical processes, and more minimalist approaches that attempt to incorporate only those processes viewed as essential to storm track dynamics. In this sense, the linear theory and barotropic modulation, as discussed in section 3, can also be viewed as modeling. With this perspective in mind, in this section, we seek to review some of the accomplishments in storm track modeling that we did not yet touch upon. We will also be highlighting those areas for which substantial work still remains.

a. GCM simulations

AGCMs attempt to incorporate all of the relevant physical processes in the models, hence all the processes discussed in section 3 are present in AGCM simulations. As such, they represent the most comprehensive models of the storm tracks. Generally speaking, storm tracks in the current generation of AGCMs are quite realistic (e.g., Lau and Nath 1987; Kageyama et al. 1999), as the weak storm tracks in earlier studies in part were the result of insufficient horizontal resolution (e.g., Boville 1991; Kageyama et al. 1999). Apart from the distribution in eddy statistics, Chang (1999) showed that the characteristics and propagation of eddies over the storm track regions are also fairly well captured in AGCM simulations. However, most comparisons between AGCM simulations and observations have been based on comparisons of the climatological storm tracks, and it is not clear that the variations in storm tracks associated with variations in the low-frequency flow component are well captured by AGCMs. This must be carefully assessed before modeled storm track changes in response to different climate change scenarios (e.g., Hall et al. 1994; Stephenson and Held 1993; Kageyama et al. 1999; Hall et al. 1996; Dong and Valdes 2000; and others) can be interpreted properly.

Apart from the climatological storm tracks, the seasonal variations in the Pacific storm track—the midwinter suppression discussed above—are captured in ECHAM4 (Christoph et al. 1997) and GFDL (Zhang and Held 1999) GCM simulations. In the GFDL GCM, Chang (2001a) found that diabatic heating strongly damps transients in January, but is nearly neutral in the spring and fall. While this may explain the GCM's midwinter suppression, it should be noted that the GCM's seasonal cycle in diabatic heating's contribution to eddy generation was enhanced vis-à-vis the reanalysis, so extension of this result to the actual atmosphere must be viewed with caution.

For interannual timescales, Straus and Shukla (1997), Zhang and Held (1999), and Carillo et al. (2000) showed that observed storm track variabilities associated with ENSO can be simulated by the Center for Ocean–Land–Atmosphere (COLA), GFDL, and ECHAM4 model simulations, respectively. Notably, there are indications that the ENSO signal is the only midlatitude interannual signal captured by ensemble integrations of GCMs forced by observed SST fields (Carillo et al. 2000; M. Hoerling 2000, personal communication).

While it is clear that the midlatitude atmospheric circulation does respond to tropical SST variations, the relationship between atmospheric circulation and midlatitude SST variations is not transparent (e.g., Lau 1997). While the midlatitude oceans certainly respond to changes in the atmosphere via both surface fluxes and wind stress anomalies (e.g., Delworth 1996; Blade 1997; Seager et al. 2000), both the extent and the structure of the atmospheric response to imposed SST anomalies remains unresolved (e.g., Palmer and Sun 1985; Latif and Barnett 1996; Ferranti et al. 1994; Kushnir and Held 1996; see the review by Robinson 2000). Storm tracks do appear to play a vital role in that response, however; Peng et al. (1997) and Peng and Whitaker (1999) have suggested that whether a model can successfully capture the correct atmospheric response to prescribed midlatitude SSTA may hinge on whether it succeeds in simulating correctly the climatological location as well as changes in the structure and amplitude of the storm tracks.

b. Storm track modeling

While AGCMs are clearly the most comprehensive tool to be used for modeling storm tracks, the complexity involved in the fully nonlinear interactions between the storm tracks and the low-frequency flow component make it difficult to understand circulation changes in a mechanistic or causal sense. Attempts have to be made to somehow separate out the two-way interactions into two distinct pieces: the response of storm track transients to changes in the planetary-scale flow, and consequent transient feedback onto that planetary-scale flow itself. For the former, theoretical considerations lead us to expect that storm track eddies should respond to changes in the jet location, baroclinicity, as well as deformation. But for given changes in basic-state flow, a storm track model is needed to quantify how the storm track responds to such changes.

While earlier attempts at modeling how changes in storm track structure focused on linear normal modes as discussed in section 3b, the inability to describe storm tracks as a single local mode quickly leads to questions regarding how changes observed in different modes due to changes in the basic state translate into changes in the storm track structure. In a clever detour around these issues, Branstator (1995) instead modeled storm track structure based on an initial value approach. Using a baroclinic model linearized about an observed (unstable) basic state, instead of solving for the most unstable normal modes he conducted an ensemble of integrations starting from random initial conditions, and used the variation among ensemble members at day 5 to represent the storm track. In this way, growth from all possible modes due to modal or nonmodal processes can be tapped. While the dependence of Branstator's model on the arbitrary choice of the time to accumulate statistics is controversial, his results do indicate that storm track anomalies accompany changes in the planetary-scale flow. Coupled with the fact that observed and modeled storm track anomalies appear to feed back positively onto concurrent planetary-scale flow anomalies (e.g., Lau and Nath 1991; Branstator 1992; Ting and Lau 1993; Hoerling and Ting 1994), Branstator made the intriguing hypothesis that there may be a natural selection process where the most frequently observed planetary-scale flow anomalies are those that are associated with the strongest positive feedbacks from concurrent storm track anomalies.

A different approach based on treating eddies as stochastically forced disturbances evolving on a baroclinically stable background flow has been taken by Whitaker and Sardeshmukh (1998) and Zhang and Held (1999), based upon ideas advocated by Farrell and collaborators (e.g., Farrell and Ioannou 1994; Delsole and Farrell 1995). The main hypothesis is that nonlinear terms may be parameterized by linear damping plus stochastic excitation, with the damping sufficiently strong to stabilize the model linearized about an observed basic state. Both Whitaker and Sardeshmukh (1998) and Zhang and Held (1999) were able to obtain good fits to the climatological storm track distribution, as well as the distribution of the eddy fluxes. Zhang and Held used their model to successfully simulate storm track variations associated with ENSO. However, while Zhang and Held were also able to simulate the midwinter suppression in the Pacific storm track using their model with forcing of constant amplitudes, Whitaker and Sardeshmukh were unable to do so unless stronger excitations are invoked during the transition seasons. It is currently not clear why their results disagree with each other. As discussed above, Chang (2001a) suggested that changes in the effects of diabatic heating may be an important contributor to the midwinter suppression, and the results of Whitaker and Sardeshmukh are consistent with that, while the results of Zhang and Held would suggest that dry dynamics may be sufficient to explain much of the suppression.

Whitaker and Sardeshmukh also used their model to simulate interannual variability of the winter storm tracks. Their success was modest, with an average anomaly correlation between their modeled storm track bandpass eddy streamfunction variance anomalies and observed anomalies being 0.3. While this discrepancy between modeled and observed storm track anomalies might be due to the effects of subseasonal variabilities not included in the model, as suggested by Whitaker and Sardeshmukh, Delsole and Hou (1999) pointed out that while the stochastic linear modeling approach based on a linear operator derived empirically from observed eddy statistics appears to give a good fit to the dominant climate statistics of a GCM simulation, using an operator constructed by linearizing a nonlinear model and then adding on simple linear damping does not. There is no doubt that nonlinear processes that are not represented by the operator in the linearized model but are in the empirical linear operator, explain a significant part of the discrepancy. Thus, fruitful use of stochastic modeling calls for addressing 1) what these nonlinear processes are, and 2) how they are represented in the empirical linear operator.

5. Concluding remarks

Reflecting over the theoretical advancement of storm track dynamics, the evolution of the paradigm is apparent. Beginning with the monumental impact that normal mode baroclinic instability (Charney 1947; Eady 1949) had on the field of atmospheric dynamics as a whole, it is natural that there were high expectations that viewing storm tracks as local modes to zonally varying basic-state flows might explain the zonal localization and other features of observed storm tracks (Pierrehumbert 1984). However, it is now well established that preexisting, finite amplitude baroclinic eddies not composing a unique modal structure are modulated by the longitudinal variation in both the baroclinic and barotropic components, and that this modulation leads to the observed storm tracks. Other processes, most notably diabatic heating, play an important modifying role in this picture, and indeed may be important in causing the nonintuitive minima in storm track activity observed over the Pacific in midwinter.

Looking to the future, a large class of important problems involves establishing a causal relationship between the temporal variability of storm track eddies and that of the background flow. As touched upon in section 3, there undoubtably is some degree of positive feedback between storm track anomalies and more slowly varying, large-scale planetary flow anomalies. However, the fact that these anomalies do not grow indefinitely provides important evidence that other physical processes limit this feedback. An overall picture of the interaction between storm track eddies and such anomalies will involve careful study of the transient evolution of these anomalies and quantifying the relative importance of changes in storm track structure versus other dynamical processes.

Finally, the improved understanding of the physical processes vital to storm track dynamics has been driven by an improved faculty in both GCM and theoretical modeling, and it is undeniable that modeling will continue to drive the field forward. From the viewpoint of basic research, the existence of seasonal and interannual storm track variability provides nature's offering for a ''parameter study.'' As a successful storm track theory should account for phenomena observed over a wide range of parameter space, an understanding of the dynamical processes behind seasonal and interannual storm track variability provides a steep challenge for any such theory. While there are numerous GCM studies that examine storm track changes under various climate change scenarios, the context that would come from an in-depth understanding of observed seasonal and interannual variability over the past 50 years is clearly in its infancy. Given the importance of storm track structures to the overall climate of the terrestrial midlatitudes, the importance of continuing to explore storm track dynamics cannot be overemphasized.

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