Is Midlatitude Convection an Active or a Passive Player in Producing Global Circulation Patterns?

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ABSTRACT

The ability of persistent midlatitude convective regions to influence hemispheric circulation patterns during the Northern Hemisphere summer is investigated. Global rainfall data over a 15-yr period indicate anomalously large July total rainfalls occurred over mesoscale-sized, midlatitude regions of North America and/or Southeast Asia during the years of 1987, 1991, 1992, and 1993. The anomalous 200-hPa vorticity patterns for these same years are suggestive of Rossby wave trains emanating from the regions of anomalous rainfall in the midlatitudes.

Results from an analysis of an 11-yr mean monthly 200-hPa July wind field indicate that, in the climatological mean, Rossby waveguides are present that could assist in developing a large-scale response from mesoscale-sized regions of persistent convection in the midlatitudes. This hypothesis is tested using a barotropic model linearized about the 200-hPa July time-mean flow and forced by the observed divergence anomalies. The model results are in qualitative agreement in the observed July vorticity anomalies for the four years investigated. Model results forced by observed tropical forcings for the same years do not demonstrate any significant influence on the midlatitude circulation. It is argued that persistent midlatitude convective regions may play a role in the development, maintenance, and dissipation of the large-scale circulations that help to support the convective regions.

1. Introduction

During the summer of 1993 the development of a region of prolonged precipitation over the northern Mississippi watershed produced one of the largest rainfall totals this century over the north-central United States. Kunkel et al. (1994) show that the monthly rainfall totals during June, July, and August 1993 over the upper Mississippi River basin were between 137% and 211% of normal. They further illustrate that although the northern Mississippi watershed had above-normal rainfall beginning in the summer of 1992, leading to saturated soils, the severe flooding did not occur until the summer months of 1993.

The extreme nature of this flood event has led to a number of scientific investigations into the factors that may have contributed to the widespread, heavy rainfall. Bell and Janowiak (1995) indicate that an anomously intense trough prevails over the central North Pacific and an enhanced ridge prevails over the western United States during February–May 1993, prior to the flooding. This circulation is a reflection of the positive phase of the North Pacific teleconnection pattern (Barnston and Livezey 1987). Their analyses show that a trough replaces this ridge in the mean height fields during June and July. In conjunction with the disappearance of the ridge over the western United States, the anomalous variance of high-pass-filtered 700-hPa heights (periods $<$10 days) increases in this region, indicative of the more frequent passage of upper-air disturbances. These disturbances are found by Maddox (1983) and Cotton et al. (1989) to be related to the development of mesoscale convective systems (MCSs). Since it is well known that MCSs contribute a significant fraction of the warm season rainfall (Fritsch et al. 1986; Heideman and Fritsch 1988), Bell and Janowiak (1995) conclude that the large-scale circulation patterns controlled the onset and persistence of the floods by defining the regions traversed by upper-air disturbances.

In a related study, Mo et al. (1995) suggest that transient eddies with periods between 2.5 and 6 days can act both to strengthen and extend the Pacific jet eastward and also to establish an upper-level jet over the central United States. This upper-level jet over the United States then interacts with the Rocky Mountains to create a lee trough that is locked over the central United States during the summer, providing a favorable large-scale flow configuration for heavy rainfall.

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Mo et al. (1995) further indicate that the low-level jet (LLJ) is a common feature in many heavy rainfall events. Paegle et al. (1996) continue this investigation using a limited area numerical model to examine more closely the role of the LLJ in producing the persistent, heavy rainfall. Their results indicate that the LLJ provides much of the moisture source for the precipitation and may be responsible for the nocturnal rainfall maximum as suggested in previous studies (Means 1954; Pitchford and London 1962; Nicolini et al. 1993). This juxtaposition of a lee trough, upper-level jet stream, an LLJ, and sufficient low-level moisture is also recognized as a favorable configuration for the development of severe local convective storms (Newton 1963).

The heavy precipitation region during 1993 is also affected by upstream evaporation (Beljaars et al. 1996). This occurs predominantly through modifications to the boundary layer over elevated terrain that is then advected over regions of lower terrain, thereby altering the inversion strength being advected into the flood region. Beljaars et al. (1996) suggest that there is a pronounced feedback between the land surface hydrology and rainfall, for large spatial and long temporal scales, that is associated with the memory of the land surface boundary condition.

A common thread in all of these studies is the implicitly assumed passive nature of the convection in midlatitudes. Convection is viewed as responding to external forces, such as the large-scale circulation pattern, upper-air disturbances, the LLJ, and soil moisture, but has little direct effect on these features. Yet the influences of organized convective regions on the local large-scale environment have been known for many years. In particular, upper-level meso-α-scale anticyclones are associated with MCSs in the midlatitudes (Ninomiya 1971a,b; Maddox 1980; Fritsch and Maddox 1981) and have been associated with wind speed perturbations of over 20 m s^{-1} and 200-hPa height perturbations of over 80 m (Fritsch and Maddox 1981; Perkey and Maddox 1985; Smull and Augustine 1993). Organized convective activity also can enhance upper-level jet streaks (Keyser and Johnson 1984; Wolf and Johnson 1995a,b) and modify the large-scale environment to be more favorable for cyclogenesis (Zhang and Harvey 1995). Admittedly, these studies focused mainly upon individual MCSs, such that feedback on the large-scale appears to be spatially and temporally restricted to the space and timescales of the parent convection.

The analytic studies of Rossby (1945) and Yeh (1949) describe how stationary vorticity sources can yield a succession of ridges and troughs downstream of the source region through wave dispersion. Hoskins and Karoly (1981) further show that large-scale heating in both the Tropics and midlatitudes can lead to the development of wave trains. One demonstration of how a succession of MCSs in midlatitudes can quickly alter the large-scale flow pattern is documented by Stensrud (1996) using output from a mesoscale model. In that study, two simulations are produced on a grid that encompasses much of the Western Hemisphere north of the equator. One simulation incorporates both parameterized convection and grid-resolvable microphysics, while the other simulation does not allow any heating to occur owing to either the parameterized convection or the grid-resolvable precipitation processes, thereby removing the diabatic effect of the convection. Comparisons of these two runs shows that the persistent convective activity produces significant changes to the large-scale flow patterns. In low levels, these changes act to increase the LLJ and the flow of moisture toward the convective region. In upper levels, the convection acts as a Rossby wave source region and produces large upper-level perturbations that cover nearly one-quarter of the hemisphere by the end of the model simulations.

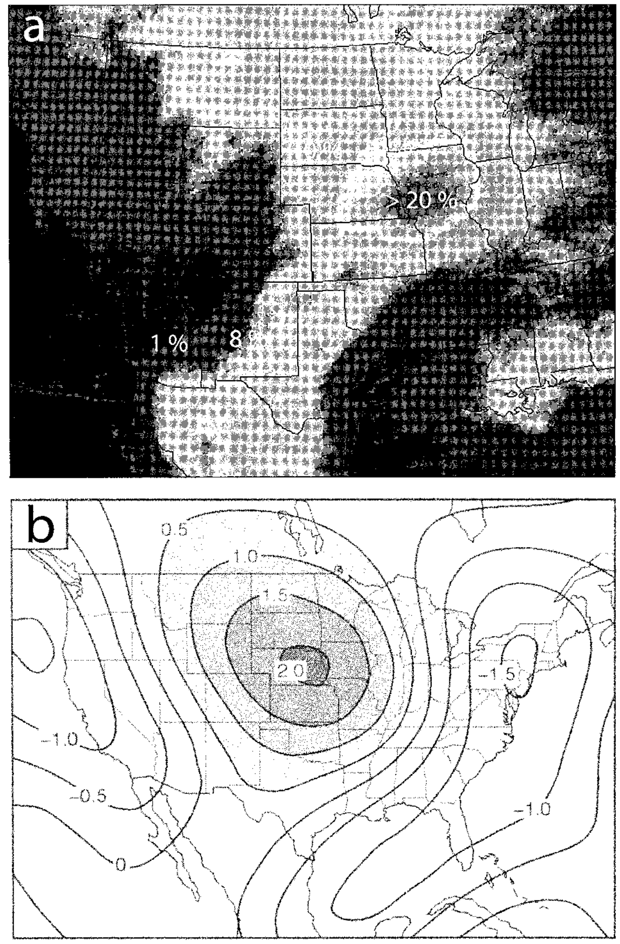

FIG. 1. (a) Frequencies of cloud-top temperatures less than −38°C derived from hourly infrared satellite imagery during Jul 1993. Black indicates less than 1% frequency, with other frequency values shown for various shadings. (b) The 200-hPa divergence anomaly from Jun and Jul 1993, with isolines every 0.5×10^{-6} s^{-1} (after Bell and Janowiak 1995).

The influence of diabatic heating on the large-scale circulation patterns is also seen in a global modeling study. Liu et al. (1998) calculate the transient vorticity,

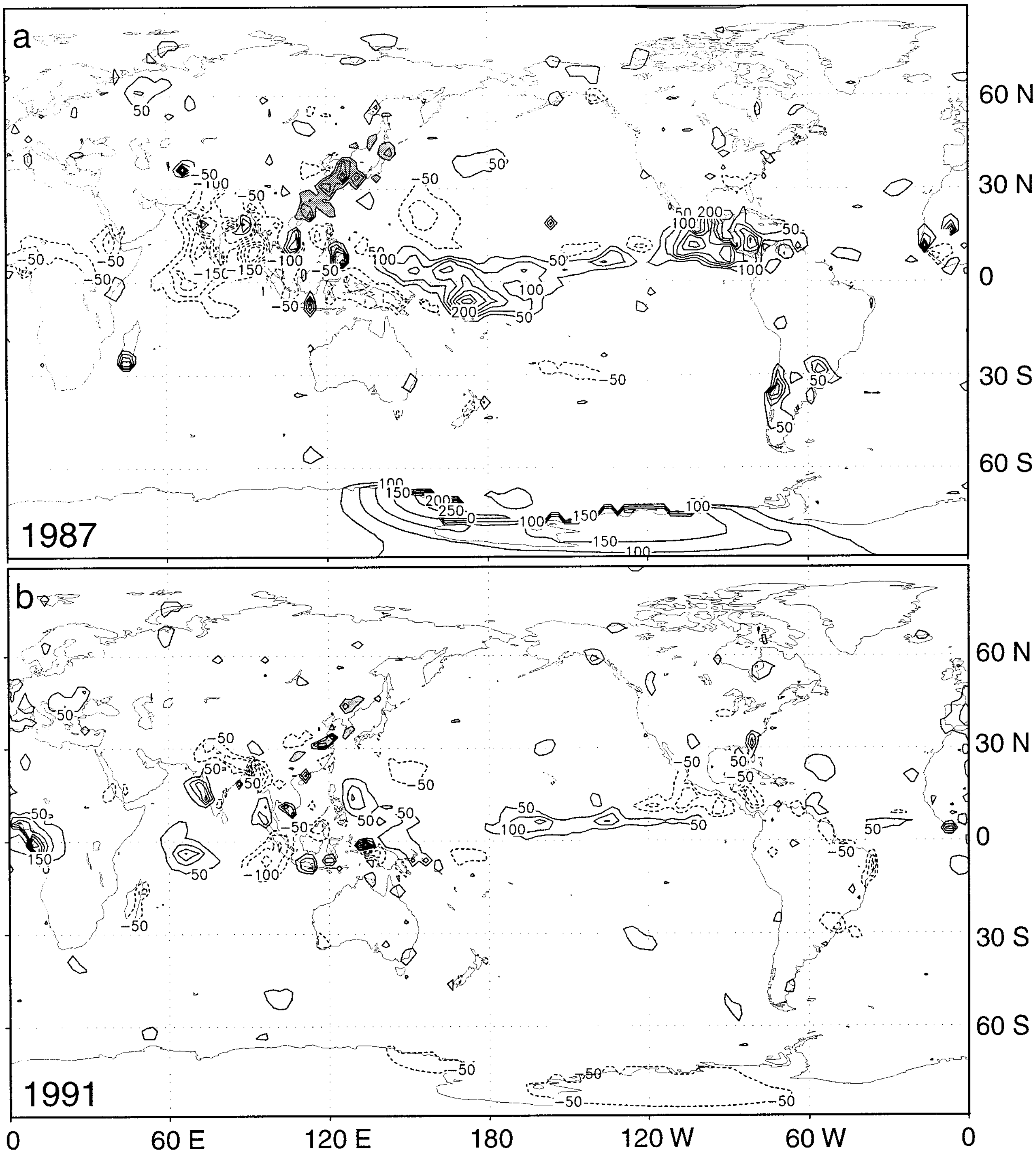

FIG. 2. Anomalous monthly rainfall (mm) for the month of Jul during (a) 1987, (b) 1991, (c) 1992, and (d) 1993 diagnosed from the CAMS/OPI dataset. Contour interval is 50 mm with negative contours dashed. Shading indicates the regions of anomalously large monthly rainfalls in the midlatitudes discussed in the text. The zero contour is not shown.

thermal forcing, and diabatic heating from global analyses over a 3-month period in order to force a linear stationary wave model. Results from this model indicate that the effects of diabatic heating on the large-scale circulation can be large. The potential influence of long-lived, mesoscale convective regions on the hemispheric circulation is also supported by Chen and Newman (1998), who indicate that the drought in the United States during 1988 is best viewed not as a single seasonal event, but rather as a succession of individual

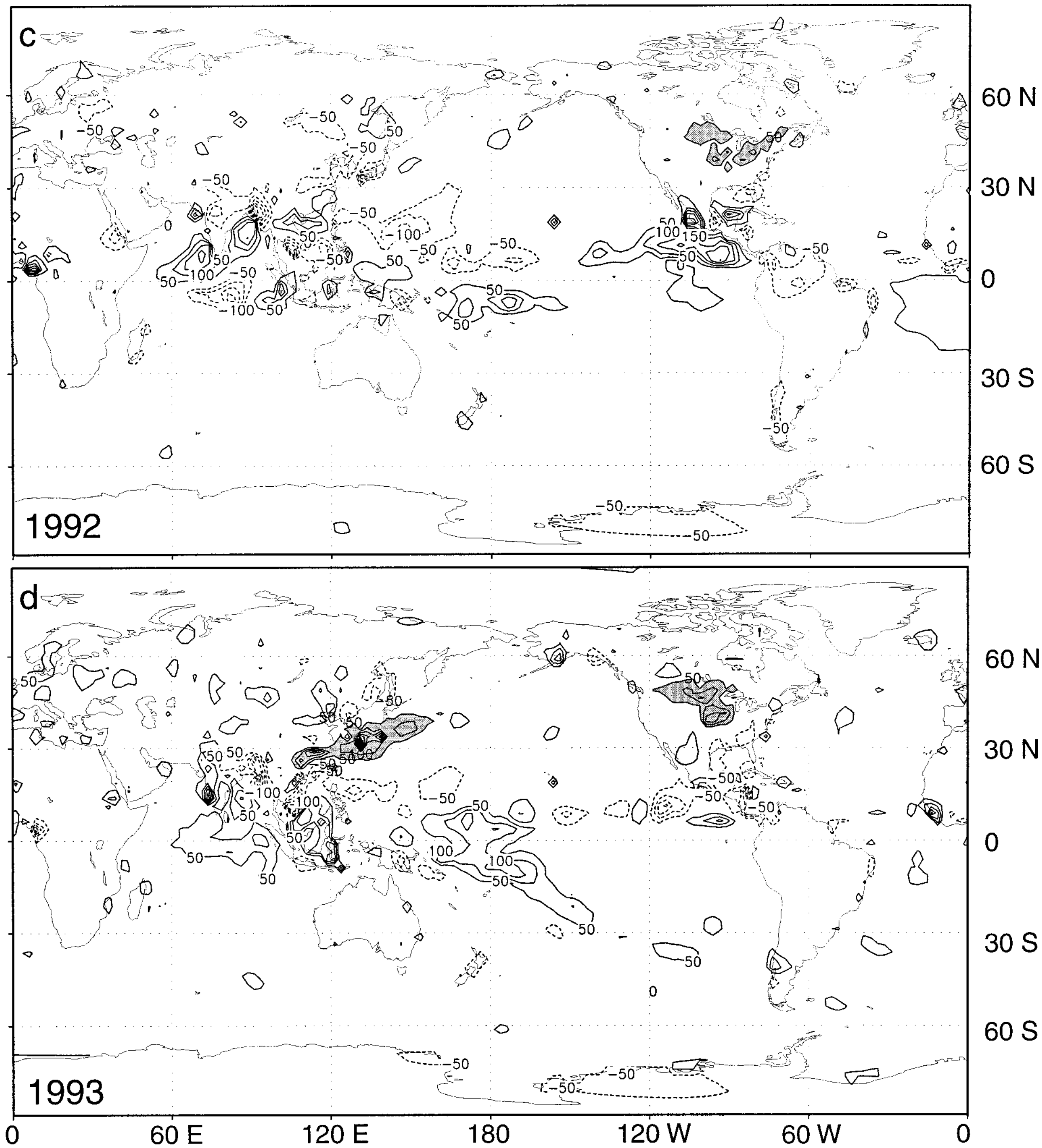

FIG. 2. (*Continued*)

events that together produce a seasonal anomaly. This viewpoint is consistent with the likely effects of successive MCSs occurring over the same general location.

The nearly continuous succession of MCSs over the central United States during summer 1993 is one of the major differences between the earlier studies of the upscale feedback of individual MCSs and the events that occurred during the Midwestern floods. The net effect of all these MCS events might be viewed as a persistent divergence source in the upper levels. This viewpoint is supported by an analysis of cloud-top temperatures as observed by satellite (Fig. 1a), indicating that over the central United States cloud-top temperatures of less than −38°C were observed over 20% of the time during July 1993, similar to the frequencies seen in the Tropics.

Bell and Janowiak (1995) further show that a 200-hPa divergence anomaly is centered over eastern Nebraska during June and July 1993, nearly collocated with the center of the highest frequency of cold cloud-top temperatures (Fig. 1b). This circumstantial evidence linking the upper-level anomalous divergence to the most frequent region of convective activity during July suggests that the persistent convection produced during the Midwestern floods could have produced a significant alteration to the hemispheric flow pattern, as suggested by Stensrud (1996).

In order to further investigate the possible upscale influences of persistent, midlatitude convective activity, two items are needed: 1) several cases of anomalously large monthly rainfall totals over mesoscale-sized regions in the midlatitudes during the Northern Hemisphere summer, and 2) a longer-term integration of a global model. Hoskins and Ambrizzi (1993), Ambrizzi et al. (1995), and Ambrizzi and Hoskins (1997) show that simple barotropic models, linearized about climatological flow patterns, produce features that are remarkably consistent with observational teleconnection studies. Therefore, as a first-order test of the hypothesis that midlatitude convection can produce an upscale feedback to the large-scale circulation patterns, a global linear barotropic model is used to examine the responses to upper-level divergence sources and compared with observations. Relevant observations, available over a 15-yr period, are described in section 2. Comparisons of the observations with results from the barotopic model are made in section 3. A final discussion is found in section 4.

2. Observations

The data used in the present study are the 2.5° × 2.5° 200-hPa wind data from the reanalysis global atmospheric data archive, created by the European Centre for Medium-Range Weather Forecasts (ECMWF), and the 2.5° × 2.5° global rainfall data archive, created by the Climate Prediction Center (CPC) of the National Centers for Environmental Prediction (NCEP). The global rainfall data incorporate rain gauge data from the Climate Anomaly and Monitoring System (CAMS), and satellite-derived rainfall estimates from the outgoing longwave radiation (OLR) precipitation index (OPI) technique of Xie and Arkin (1998). These data are merged at the CPC using a method adapted from Xie and Arkin (1996) to produce the CAMS/OPI data used in this study. The ECMWF data are available at daily intervals, whereas the CAMS/OPI rainfall data are monthly mean values. These data are used to define the departures from a mean base period of the large-scale patterns and rainfall amounts.

Maps of the CAMS/OPI July anomalous rainfall, defined as the departure from a 1979–1995 base period, show mesoscale-sized regions (5° × 5° or larger) of above-normal rainfall (>50 mm with peak value of 100 mm or greater) in the Northern Hemisphere above 30°N for four separate years (Fig. 2). There are two main regions of anomalously large midlatitude July total rainfall: central North America and Southeast Asia. Both regions are associated with the frequent occurrence of large MCSs as shown by Laing and Fritsch (1997). However, Southeast Asia is also influenced by tropical cyclones during July and it is likely that some of the anomalously high monthly rainfalls are produced by tropical cyclones. Other regions of anomalously large rainfall are evident south of 30°N and are often within regions of anomalously warm sea surface temperatures (not shown). While there is some evidence of an impact of tropical sea surface temperature forcing on the extratropical circulation and precipitation over the central United States during extremely wet years, the signal is limited during spring (Bates et al. 2001) and extremely weak during summer (Kumar and Hoerling 1998). Therefore, we focus our attention on the possible role played by the midlatitude precipitation anomalies only.

Examination of tropical cyclone track data for these years indicates that an average of 1.3 tropical cyclones occur during July in the region between 30° and 40°N and 120° and 150°E per year. For the three years of interest, the track data indicate that July 1987 had two tropical cyclones in this region, July 1991 had one tropical cyclone, and July 1993 had three tropical cyclones. This suggests that the years 1987 and 1991 are not anomalous with respect to the number of tropical cyclones, but July 1993 likely is anomalous owing to the three tropical cyclones that moved over Southeast Asia. The track data are not sufficient to discriminate between rainfall produced by MCSs versus produced by tropical cyclones, so it is not possible to determine the rainfall totals produced by persistent mesoscale convective activity only. However, the location and orientation of both the rainfall and the 200-hPa divergence anomalies (not shown) are not aligned with the tropical cyclone paths for any of the years. This suggests that it is reasonable to assume that MCSs produced much of the rainfall over Southeast Asia during July 1987 and 1991, with July 1993 being more problematic.

To further investigate the potential role of these persistent convective regions, the ECMWF 200-hPa reanalysis wind data are used to calculate a mean July flow pattern over the years 1979–93. Data from the four years with anomalously large rainfall totals are not included in this analysis. The theory of Rossby wave propagation, as discussed in Hoskins and Ambrizzi (1993) and Ambrizzi et al. (1995), is used to identify preferred waveguides and typical wave behavior. They show that the Rossby wave ray paths have clockwise curvature if K_S, the stationary wavenumber,[1] increases toward the equator and have anticlockwise curvature if K_S increases

[1] Here $K_S = (\beta^*/U)^{1/2}$, where β^* is the poleward gradient of absolute vorticity and U is the westerly wind.

toward the pole. This curvature increases as the meridional gradient in K_S^{-1} increases. Using these basic theoretical results, it can be shown that strong westerly jets can act as Rossby waveguides.

The mean u-component 200-hPa July flow shows the Asian, North American, and Pacific jet streams in the Northern Hemisphere (Fig. 3a). With easterly wind throughout the tropical region, there are no favorable regions for cross-equatorial wave propagation (Webster and Holton 1982; Branstator 1983). In the middle of the Asian, North American, and Pacific jet streams, the meridional gradient of absolute vorticity (β^*) has maximum values, with minima located to both the north and south (Fig. 3b). Hoskins and Ambrizzi (1993) have shown that upstream Rossby wave propagation can occur in regions of large β^*, with the speed of the westward propagation directly proportional to the magnitude of β^*. This result, when applied to the mean July flow conditions, suggests that a wave source over North America could conceivably influence the Pacific region, as there are two zones of higher β^* values stretching into the Pacific from North America (Fig. 3b). In particular, the position and orientation of the southern zone of higher β^* (stretching from 30°N, 130°W to 10°N, 160°E as shown by the arrow in Fig. 3b) would produce upstream wave propagation southwestward from the west coast of North America into the south-central Pacific.

The geographical distribution of the stationary wavenumber K_S shows a zone of uniform values, with typical wavenumber 10, stretching from North America northeastward into the North Atlantic that would act as a Rossby waveguide (shown by the arrows in Fig. 3c). For wave source regions over Southeast Asia, another waveguide exists for wavenumber 10 that stretches eastward from Korea to Japan and northeastward into the northern Pacific (shown by the arrows in Fig. 3c). Waves propagating through this region would be directed toward the Gulf of Alaska where a critical line of $U = 0$ exists and wave absorption would occur (Hoskins and Karoly 1981), although the realism of this behavior, seen in linear dissipative models, is uncertain (Tung 1979).

These analyses suggest three ways in which persistent, midlatitude convective regions in North America and Southeast Asia could potentially influence the large-scale circulation patterns. The first is that midlatitude convection over North America produces Rossby waves that propagate northeastward into the North Atlantic region (Fig. 3c). However, this same area of convection may also produce upstream Rossby wave propagation from the convective region into the south-central Pacific along the zone of high β^* values (Fig. 3b). Finally, persistent convection over southeastern Asia produces Rossby waves that propagate into the western and central Pacific region (Fig. 3c). To determine if any of these patterns are apparent from observations, further analysis of the reanalysis wind data is conducted.

The ECMWF 200-hPa wind data are used to calculate the July vorticity anomalies for the four anomalous years to determine whether or not waves are suggested in the anomaly fields. Patterns of positive/negative vorticity anomalies occurring near the waveguides shown by the arrows in Fig. 3c would suggest the possibility of Rossby wave propagation from a midlatitude convective region. Results suggest that waves are observed in the western Pacific for the two years (1987, 1991) with persistent mesoscale-sized convective areas during July only occurring over Southeast Asia (Fig. 4). These waves propagate northeastward from their source region into the north-central Pacific area. The wave propagation path appears to turn equatorward near the date line, similar to that seen in Ambrizzi et al. (1995). They argue that the wave activity is able to tunnel through the low (in their case, negative) K_S region of U and β^* maxima found in the central Pacific (associated with the jet maxima near 20°N, 170°W in Fig. 3a), although this behavior has not been suggested by lag correlation analyses (Ambrizzi et al. 1995).

In contrast, the calculated July vorticity anomaly for 1992, in which persistent convection occurs over North America, is suggestive of both downstream and upstream wave propagation (Fig. 5). Downstream wave propagation is suggested by the positive vorticity anomaly in the North Atlantic, while the upstream propagation is suggested by the zone of positive vorticity extending southwestward into the eastern Pacific from the proposed source region over central North America. This zone of positive vorticity is near the axis of high β^* values (Fig. 3b), consistent with the expected outcome for a North American source region.

A combination of both Southeast Asian and North American source regions occurs during July 1993. The calculated July vorticity anomaly for this year shows indications of the behaviors seen in the earlier years (Fig. 6). There is a clear positive vorticity anomaly over the North Atlantic, suggestive of wave propagation from persistent convective activity in central North America. There is also a zonally elongated, positive vorticity anomaly in the western Pacific, suggestive of wave propagation from the Southeast Asian source region. However, this picture is clearly not a simple linear addition of the two responses previously described and includes several features, such as the negative vorticity anomaly off the west coast of the United States, that are not observed in the other years. Thus, it appears that July 1993 represents the most complex of the four scenarios available to examine.

If the July reanalysis data are used to calculate the vorticity anomalies for years in which there is no significant, positive, midlatitude rainfall anomalies, then we find that the anomalies are more suggestive of a meridional shift in the jet stream position (Fig. 7). The two examples both have regions of 60° in longitude that have consistent positive or negative vorticity anomalies over nearly the same latitude bands. These patterns do not appear to be consistent with wave propagation from

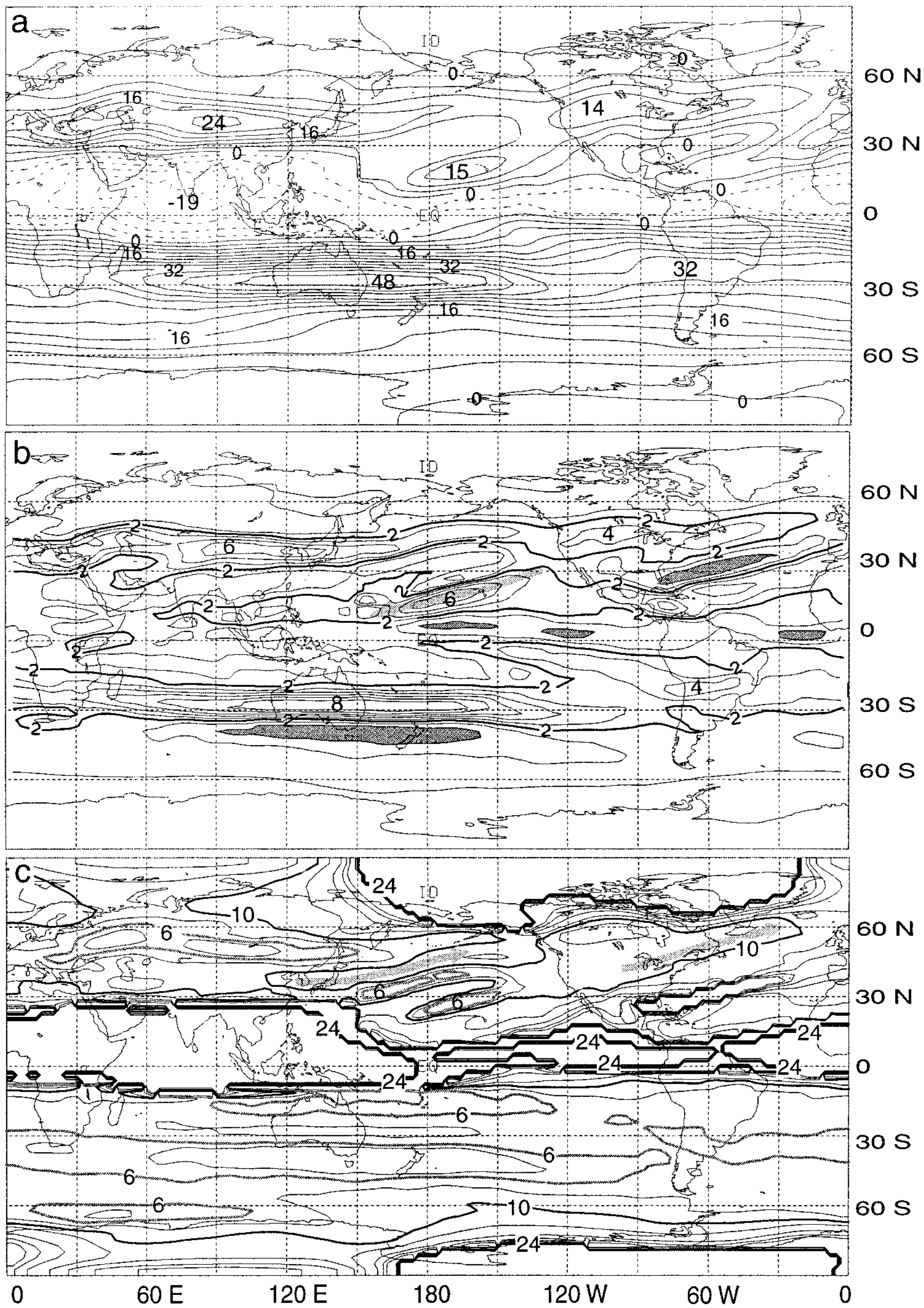

FIG. 3. The mean Jul 200-hPa flow based upon ECMWF data from 1979 to 1992. (a) Westerly *u* component of the wind (m s^{-1}) with a contour interval of 4 m s^{-1}. (b) Meridional gradient of the absolute vorticity β^* with a contour interval of 1×10^{-11} s^{-1} m^{-1}, with negative values shaded (zero contour not drawn) and a thicker line used for the 2×10^{-11} s^{-1} m^{-1} contour. (c) Stationary wavenumber K_S for β^* positive and westerly winds. Wavenumbers 6, 8, 10, 12, 14, and 16–24 at intervals of 4 are shown, with a gray line for wavenumber 6, and a thicker line used for wavenumbers 10 and 24. The zone of possible upstream propagation of Rossby waves from a North American source region is indicated by a gray arrow in (b), while Rossby waveguides for upper-level divergence sources over Southeast Asia and North America are indicated by gray arrows in (c). Negative contours are dashed in all panels.

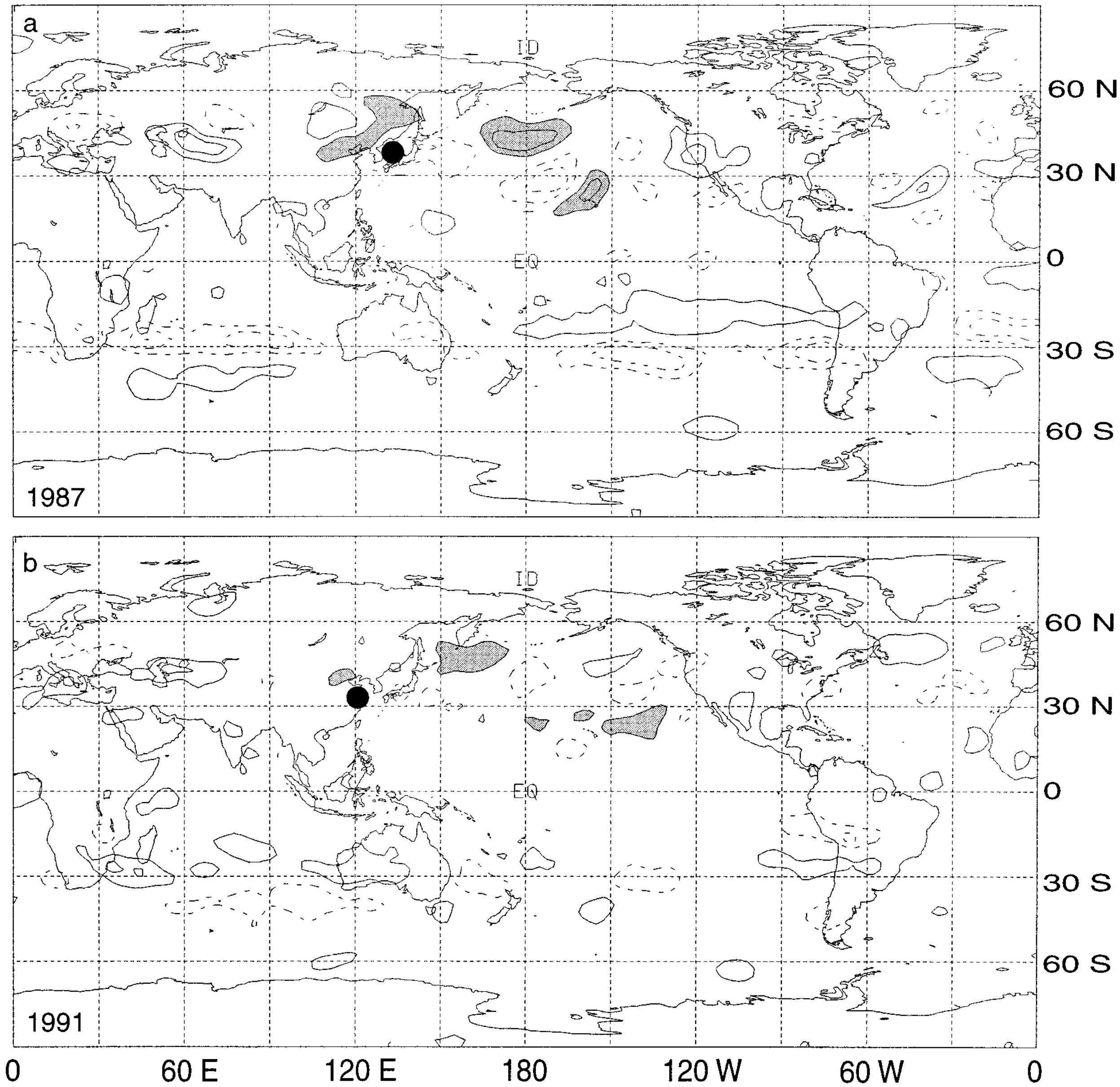

FIG. 4. The analyzed Jul vorticity anomalies for (a) 1987 and (b) 1991 calculated from the ECMWF reanalysis data. The contour interval is 1×10^{-5} s^{-1} and negative contours are dashed. The zero contour is not shown. The black circle indicates the observed region of anomalously large monthly rainfall totals. Positive vorticity anomalies that occur in regions of Rossby waveguides are shaded.

midlatitude or tropical source regions (see Fig. 3). The positive vorticity anomaly during 1981 that stretches along 30°N from 100° to 180°E is in a region of minimum β^* values (Fig. 3b) and has no collocated waveguide (Fig. 3c), indicating that these anomalies do not conform to the behavior of Rossby waves. In contrast, the positive vorticity anomaly during 1988 that stretches east–west across the North Pacific is within a region of high β^* values. However, 1988 was a drought year in the United States (Trenberth and Guillemot 1996), with no North American source for the upstream propagation of Rossby waves from persistent convective activity (although other types of anomalous forcing cannot be ruled out immediately).

We conclude that the four years in which anomalously heavy rainfall occurred in the midlatitudes north of 30°N are sufficiently unique to warrant further investigation. Numerical simulations with a barotopic model are used to explore further the possibility that the observed regions of persistent convection, over southeastern Asia and central North America, can influence the large-scale flow patterns.

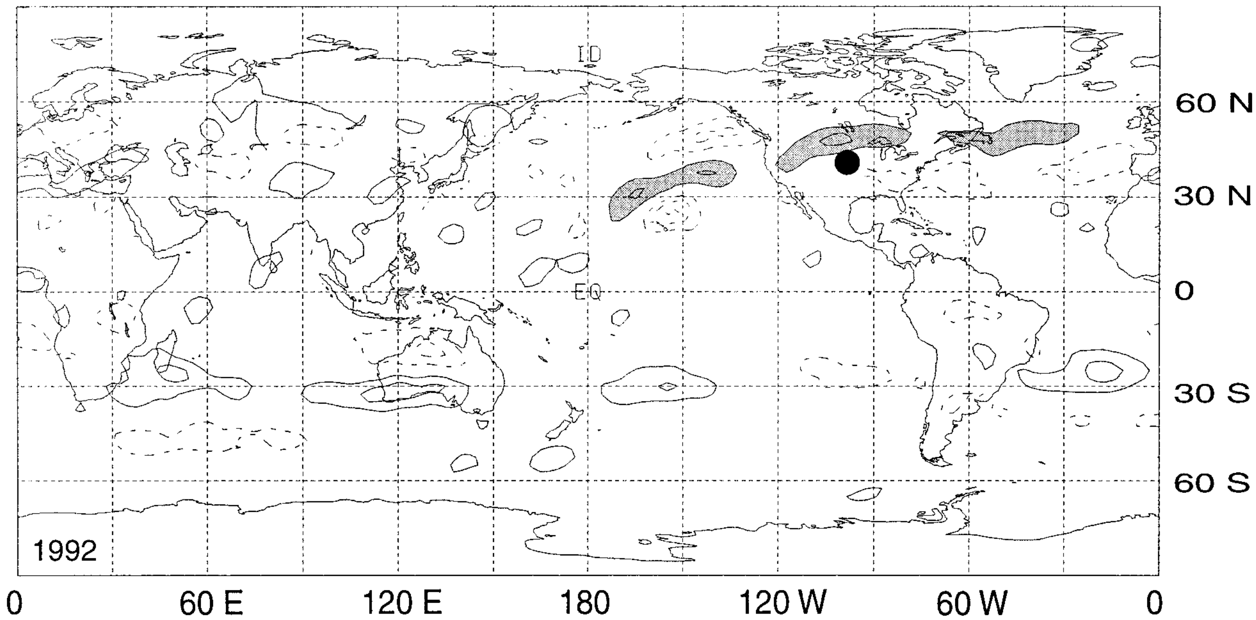

FIG. 5. The analyzed Jul vorticity anomalies for 1992 calculated from the ECMWF reanalysis data. The contour interval is 1×10^{-5} s^{-1} and negative contours are dashed. The zero contour is not shown. The black circle indicates the observed region of anomalously large monthly rainfall totals. Positive vorticity anomalies that occur in regions of Rossby waveguides are shaded.

3. Barotropic model

The model used is based on the damped barotropic vorticity equation:

$$\left(\frac{\partial}{\partial t} + \mathbf{V}\cdot\boldsymbol{\nabla}\right)\zeta = F - \lambda\zeta - \mu\nabla^4\zeta, \qquad (1)$$

where $\mathbf{V} = \overline{\mathbf{V}} + \mathbf{V}'$ is the nondivergent barotropic velocity, $\zeta = \overline{\zeta} + \zeta'$ is the absolute vorticity, $F = \overline{F} + F'$ is a prescribed time-constant forcing, λ is a linear damping coefficient with a timescale of 4 days, and μ is a diffusion coefficient set so that the smallest resolved wave is diffused with a timescale of 10 days. The overbar represents a horizontally varying, temporally constant basic-state quantity and the prime a departure from the basic state. The basic forcing $\overline{F}$ is selected so that the basic-state quantities are an exact solution of (1):

$$\overline{F} = \left(\frac{\partial}{\partial t} + \overline{\mathbf{V}}\cdot\boldsymbol{\nabla}\right)\overline{\zeta} + \lambda\overline{\zeta} + \mu\nabla^4\overline{\zeta}. \qquad (2)$$

An additional forcing, F', which represents the local forcing associated with anomalous precipitation events, is defined as

$$F' = \begin{cases} A\omega\cos^2\left(\dfrac{\pi\Delta x}{2r_x}\right)\cos^2\left(\dfrac{\pi\Delta y}{2r_y}\right)\sin(\varphi), & \Delta x \le r_x, \quad \Delta y \le r_y \\ 0, & \Delta x > r_x, \quad \Delta y > r_y, \end{cases} \qquad (3)$$

where A is the amplitude of the perturbation, ω is the earth's rotation, Δx and Δy are the distances from the center of the perturbation in the x and y directions, respectively, r_x and r_y are the radii in the x and y directions, respectively, and φ is latitude. The value of A is chosen to match the value of the divergence anomaly calculated from the ECMWF reanalysis data for the month of interest over the specified source region. Values of A vary from 0.5×10^{-5} s^{-1} to 1.4×10^{-5} s^{-1}. The model (1) is solved using a spectral transform technique as in Hoskins and Ambrizzi (1993) with a triangular trunction at wavenumber 42 (T42). The source region has a cosine squared amplitude over an elliptical or circular region, designed to approximate the size and shape of the observed divergence anomaly (see Figs. 4–6 for the locations of these perturbations). Tropical sources are also examined, using a typical 1.0×10^{-5} s^{-1} value for A, to fit the approximate size and shape of the anomalous regions of mean monthly rainfall diagnosed using the CAMS/OPI data. A 30° by 20° ellipse is used as the

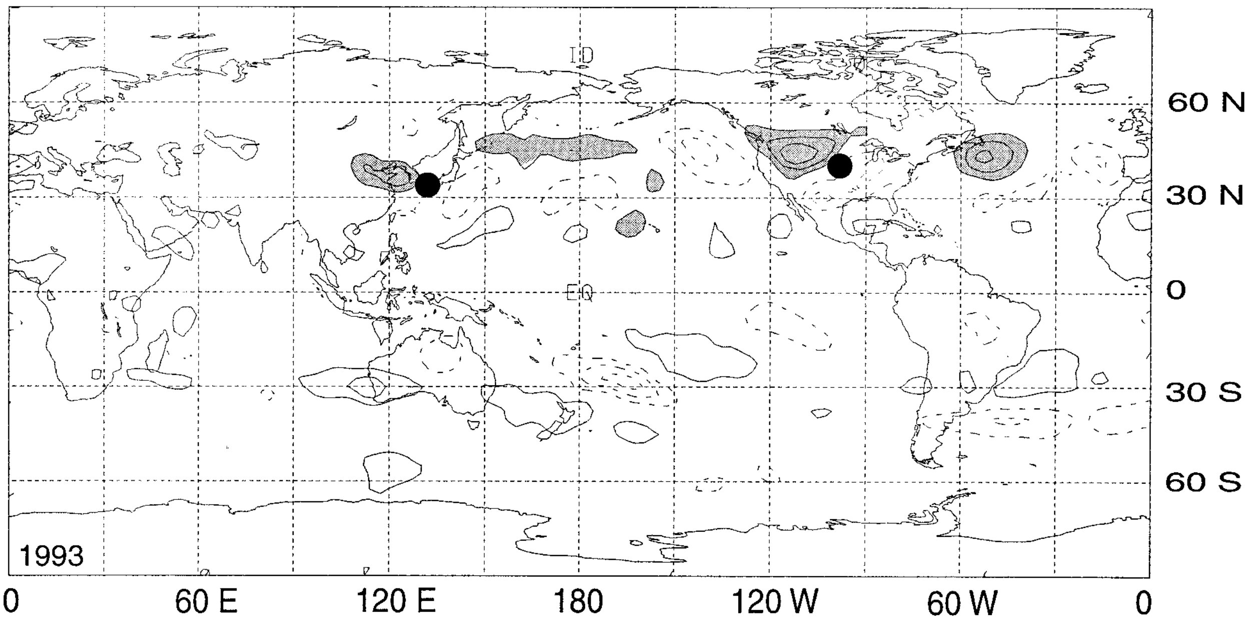

FIG. 6. As in Fig. 5, but for Jul 1993.

source region shape for the simulations with tropical source regions.

Sensitivity tests using various values of μ and λ and the 200-hPa July mean flow indicate that the barotropic model results are not qualitatively sensitive to a factor of 2 change in the values chosen for the damping and diffusion coefficients, with the general patterns of the wave trains unmodified (not shown). However, the *amplitude* of the model perturbation vorticity is affected by even small changes in the diffusion and linear damping coefficients. This suggests that the anomalous vorticity patterns produced by the model are more robust than the vorticity magnitudes. In contrast, the model is not particularly sensitive to the exact location of the local Rossby wave source region for shifts $\pm 5°$ in any direction.

Admittedly, the barotropic model results are not able to explore all the possible interactions between persistent, deep convection and the large-scale environment. The model also is not able to simulate the Madden–Julian oscillation that is known to influence tropical convection (Madden and Julian 1994). However, this very simple model should be sufficient to investigate our question about the role that persistent midlatitude convective regions may play in favorably altering the large-scale circulation patterns to support further convective development. Reasonable agreement between observations and model simulations of the vorticity anomalies is viewed as supporting evidence for this hypothesis, although a complete examination of cause and effect is not possible.

Vorticity anomalies from day 20 of the barotropic model with forcing over southeastern Asia are qualitatively similar to those analyzed during 1987, showing that a Rossby wave train emanates from the source region over southern Japan (Fig. 8) with a wavelength close to that observed. The wave train appears to turn equatorward near the date line, in agreement with the analyses (Fig. 4a). While the model does not reproduce the precise location of the vorticity anomalies, the general behavior is quite similar to the observations. Since the mean 200-hPa background flow used in the model is from an 11-yr monthly mean, these differences between the simulated and observed anomaly patterns from a single month are not surprising. Thus, while the model cannot explain the analyzed positive vorticity anomaly along the west coast of the United States, it does replicate many of the other features seen, including the positive/negative vorticity couplet in the mid-Atlantic (along 30°N).

The calculated divergence source during 1991 is weaker than that calculated during 1987, and produces a weaker response in the model (Fig. 9). The source region for 1991 is also located farther to the southwest, over southeastern China, providing for a more northeastern wave propagation. However, this behavior is again supported by the analyzed vorticity anomaly (Fig. 4b). The positive vorticity anomaly located just to the west of the date line is in a similar position in both the model and the analysis. There is a small hint of the equatorward turning of the wave train in this model simulation, as indicated by the small positive vorticity anomaly near 25°N and 150°W. As with the 1987 results, the model replicates several of the observed features, but it is, not surprisingly, unable to account for all the anomalies seen in the analyses. Yet the model behavior

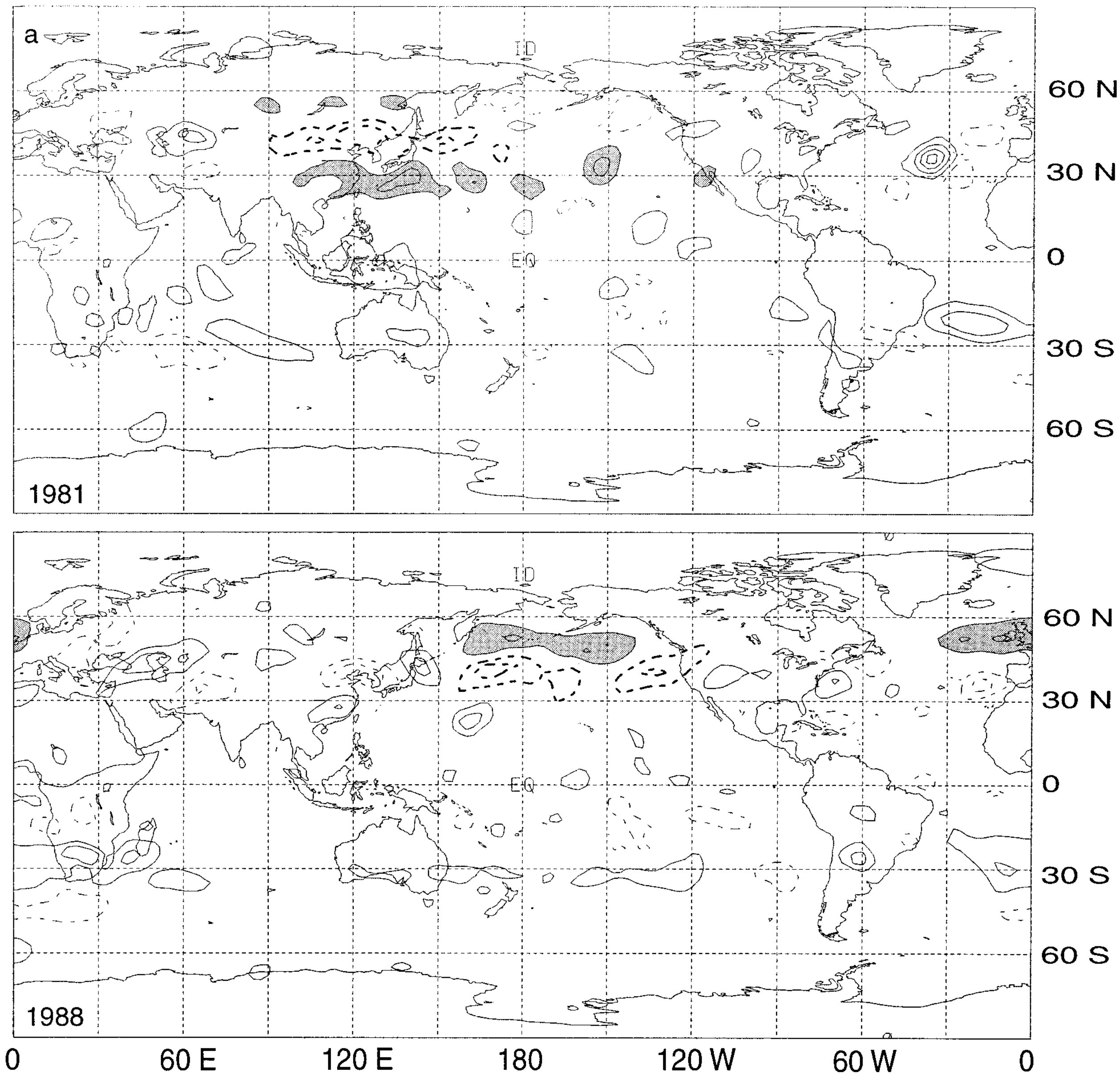

FIG. 7. The analyzed Jul vorticity anomalies for (a) 1981 and (b) 1988 calculated from the ECMWF reanalysis data. The contour interval is $1 \times 10^{-5}\ s^{-1}$ and negative contours are dashed. The zero contour is not shown. Large positive vorticity anomalies that stretch across more than 60° in longitude are shaded.

is consistent with the analyses over the region that theory suggests might be influenced by a midlatitude region of persistent convection.

The rainfall anomaly during 1992 is over the central United States, and the barotropic model produces a very different response for this source region (Fig. 10). One wave train emanates from the central United States and propagates northeastward into the North Atlantic along the waveguide suggested by theory (arrows in Fig. 3c). However, there is also an upstream (westward) extension of the positive vorticity anomaly from the source region in central North America, southwestward into the central Pacific. This behavior qualitatively agrees well with the analyzed vorticity anomaly for 1992 (Fig. 5) for both the upstream and downstream propagation.

As mentioned earlier, the analyzed vorticity anomaly patterns for 1993 do not easily fit into the simple patterns suggested by wave source regions located over either Southeast Asia or North America. Model results for the two source regions analyzed during 1993 show several features that resemble the analyzed patterns (cf. Figs. 11 and 6), but miss many others. Again, the model reproduces the wave train that emanates from the source region over North America and propagates northeast-

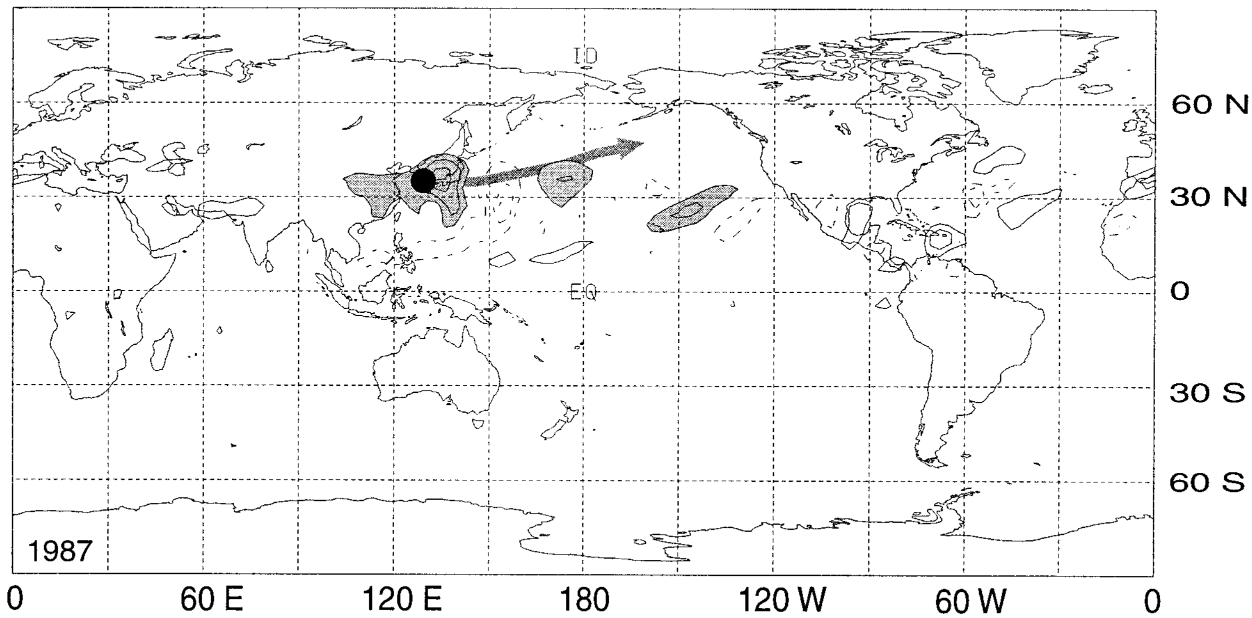

FIG. 8. Model vorticity anomalies from day 20 for the forcing centered over Southeast Asia for Jul 1987. The contour interval is 1×10^{-5} s^{-1} and negative contours are dashed. The zero contour is not shown. Positive vorticity anomalies that occur in regions of Rossby waveguides (indicated by gray arrows) are shaded.

ward into the North Atlantic along the waveguide. The model also produces a wave train emanating from the source region over Southeast Asia that propagates into the central Pacific, although this is not as clearly seen in the analysis (Fig. 6). However, during 1993 there is no analyzed unbroken, upstream extension of the positive vorticity anomaly from the North America source region into the Pacific (Fig. 6) that is seen clearly in the model data (Fig. 11). There is at best a hint of an analyzed positive vorticity anomaly in the central Pacific (near 20°N, 160°W) that may be an indication of this behavior. However, the analyzed negative vorticity anomaly off the West Coast of the United States is not reproduced by the barotropic model.

While the analyzed and model results for 1993 are somewhat ambiguous, the qualitatively good agreement between the model and corresponding analyses for 1987, 1991, and 1992 suggests that the midlatitude con-

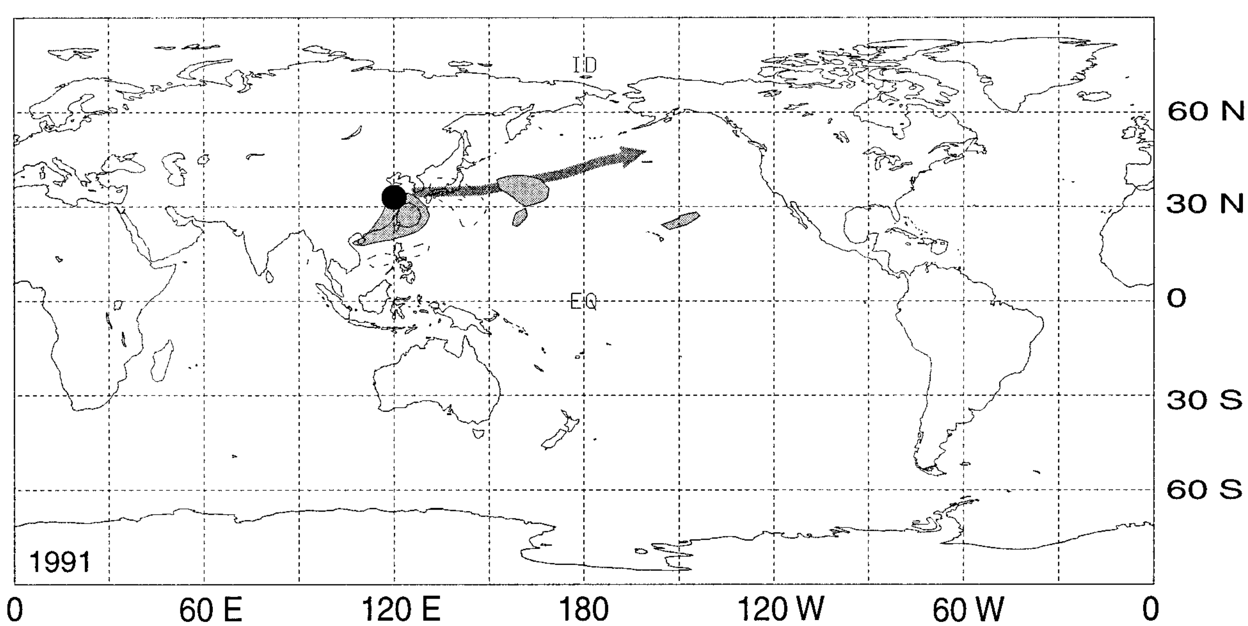

FIG. 9. As in Fig. 8, but for the forcing centered over Southeast Asia for Jul 1991.

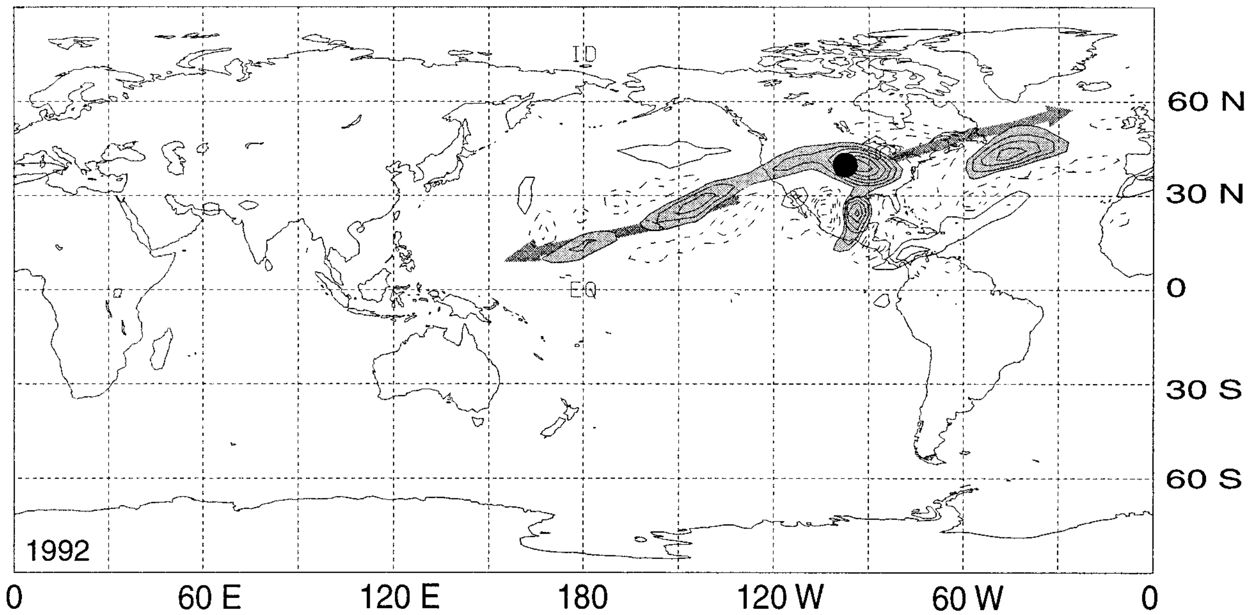

FIG. 10. As in Fig. 8, but for the forcing centered over central North America for Jul 1992.

vective regions may have played a role in their own longevity. The southwestward extension of the positive vorticity anomaly from the North American source region produces an enhancement to the westerly flow across the Pacific and the southern Rocky Mountains. Mo et al. (1995) argue that increased westerly flow over North America produces a lee trough, forced by the Rocky Mountains, that assists in sustaining the low-level flow into the convective region during 1993. A similar process may explain the persistence of the midlatitude convective region during July of 1992. The location of the high terrain associated with the Tibetan and Yun-Gui Plateaus in Asia suggests that a similar mechanism may be acting, although the upstream wave propagation over Asia is much less than over North America. Results from Stensrud (1996) indicate that the low-level flow is also sustained locally by the feedbacks from the convection, suggesting a multiscale, positive feedback mechanism at low levels. The midlatitude convection produces a large-scale, upper-level response that

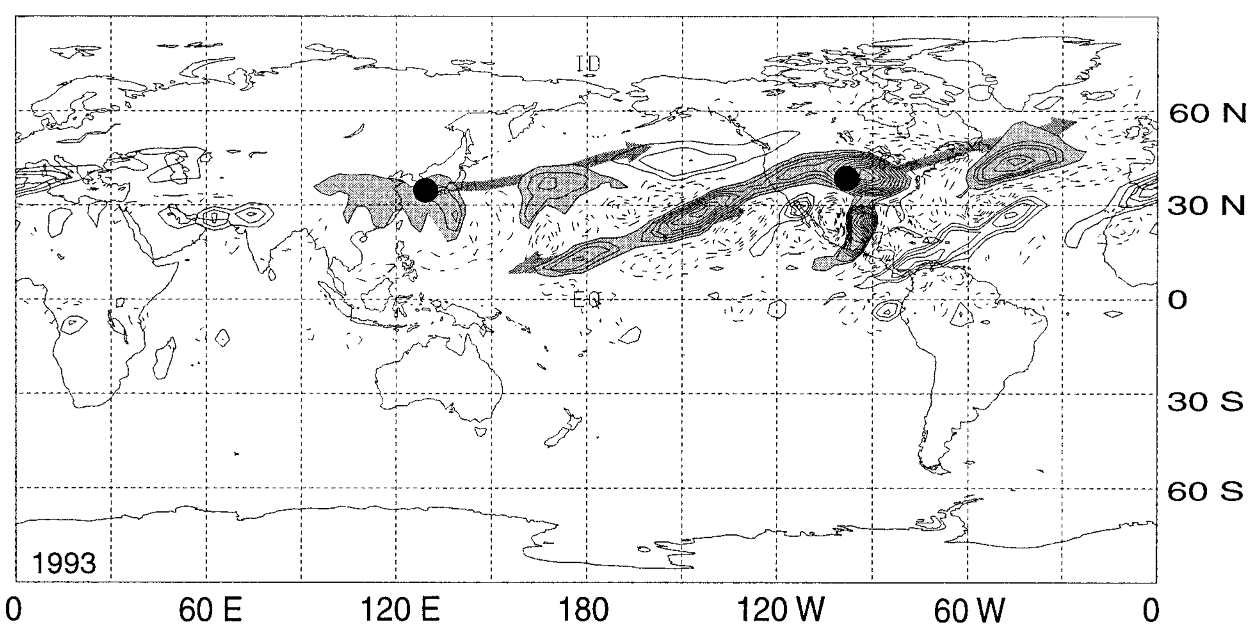

FIG. 11. As in Fig. 8, but for the forcing centered over both Southeast Asia and central North America for Jul 1993.

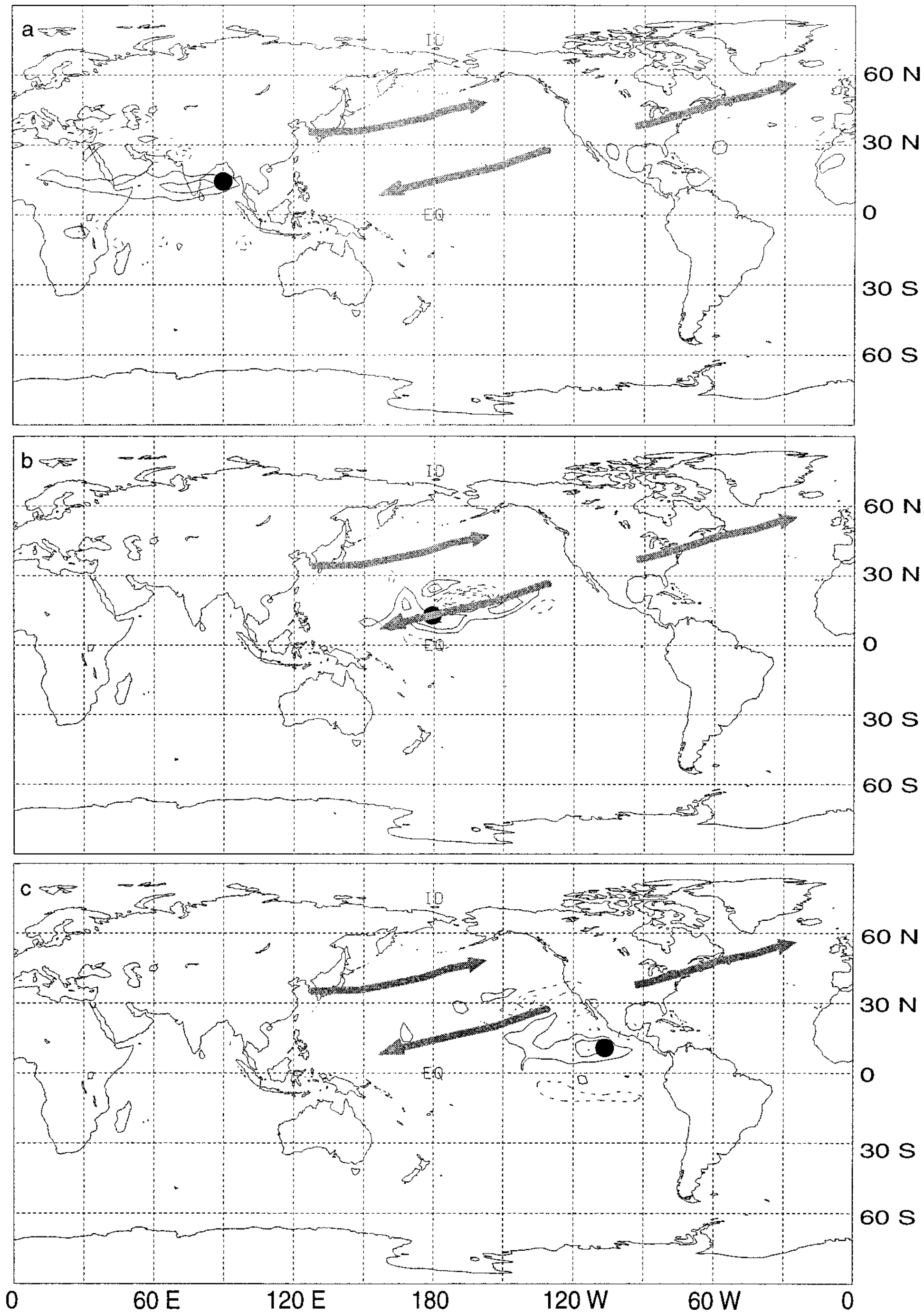

FIG. 12. As in Fig. 8, but for the forcing centered over (a) the Bay of Bengal, (b) the central Pacific, and (c) the eastern Pacific.

assists in sustaining a large-scale flow pattern that is conducive to further convective development.

It is well-known that, in winter, anomalous forcing in the Tropics is the most efficient way to generate an extratropical response. In order to investigate whether or not tropical anomalous forcing is an important factor in the summer cases being examined, three different tropical forcing anomalies are examined using the barotropic model. The regions of anomalous forcing are chosen to be consistent with the observed CAMS/OPI rainfall anomalies diagnosed for the four years being examined. These areas are over the Bay of Bengal, the central Pacific, and the eastern Pacific (see Fig. 2). Results indicate that the tropical sources do not produce significant midlatitude vorticity anomalies (Fig. 12). The source region with the largest response is the eastern Pacific, which produces a northwestward propagating wave train. While similar features can be seen in the analyses (Figs. 4–6), their importance to the global circulation and the regions of persistent, midlatitude convection is minimal. Hence, it appears from these simple model results that the influence of tropical convective anomalies on the July circulation patterns is not large. Nevertheless, tropical forcing could be one of the many factors that explain part of the inconsistency between the extratropically forced barotropic model results and the observations.

4. Discussion

Results from both analyses and a simple, one-level barotropic model suggest that persistent MCS activity over central North America and southeastern Asia during the Northern Hemisphere summer can produce source regions for Rossby waves that influence the hemispheric circulation pattern in midlatitudes. It is possible that these Rossby waves provide a positive feedback mechanism, through the enhancement of westerly flow over mountain ranges (Mo et al. 1995), for deep convection. While these results are not conclusive, owing to the simplicity of the model and the limited number of cases available for examination, they do suggest that long-lived convection in the midlatitudes do not play a passive role in the global circulation as is often implicitly assumed. While it is well-known that the most efficient way to generate a midlatitude response in a global model is to perturb the Tropics during the wintertime, tropical systems do not necessarily dominate the hemispheric flow patterns during the summertime owing to the presence of easterly winds in the lower latitudes. Several studies show evidence of an influence of tropical sea surface temperature forcing on the extratropical circulation and precipitation over the central United States during extremely wet years, but the signal is limited during spring (Bates et al. 2001) and extremely weak during summer (Kumar and Hoerling 1998).

A number of studies have examined the global distribution of large and long-lived MCSs, which show a distinct tendency to occur in specific areas (Laing and Fritsch 1997). Results from the present study suggest that it is possible that these favored MCS regions occur, in part, owing to the interaction of the MCSs with the large-scale circulation. In this scenario, the succession of MCSs produces an alteration to the large-scale flow through Rossby wave propagation that enhances the flow over the upstream mountain ranges. This enhanced flow acts to develop a persistent lee trough to the east of the mountains as shown in Mo et al. (1995). The MCSs also act to enhance the low-level flow of warm, moist air from the south and southeast toward the convective region. Therefore, the development of a long-lived, mesoscale-sized region of convection occurs through alterations to both the large-scale hemispheric circulation patterns and more localized low-level flow. Unfortunately, global MCS studies have only been conducted over a handful of years and the data are not sufficient to examine the month-to-month relationships between the MCS regions. An extension of these studies is highly desirable with the goal of relating these MCS regions to hemispheric circulation patterns.

These results suggest that studies with general circulation models, possibly with embedded mesoscale domains, are needed to evaluate the ability of midlatitude convective regions to interact over longer timescales. While the barotropic model results are very suggestive, the model is highly idealized and ignores the baroclinic processes that are important near the convective regions. The use of a more complex numerical model may produce different results, although the similarity between barotropic model results and the atmosphere has been compelling in the studies of Hoskins and Ambrizzi (1993) and Ambrizzi et al. (1995).

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