Eddy Shedding from the Upper-Tropospheric Asian Monsoon Anticyclone

JELENA M. POPOVIC AND R. ALAN PLUMB

Program in Atmospheres, Oceans and Climate, Massachusetts Institute of Technology, Cambridge, Massachusetts

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ABSTRACT

The authors investigate the transient behavior of the Asian monsoon anticyclone in the summertime upper troposphere for the four northern summers 1987–90. The evolution of potential vorticity near the tropopause shows the development of westward migrating anticyclones breaking off from the main anticyclone a few times each summer. These disturbances are relatively shallow, being confined to the upper troposphere.

1. Introduction

In shallow water simulations of a large-amplitude anticyclone forced by a localized mass source on a beta plane, Hsu and Plumb (2000) found secondary anticyclones being shed westward periodically as a result of the instability of the parent anticyclone. Further, they showed two examples, in July 1990, of apparently similar westward shedding of anticyclones, in the upper troposphere, from the Tibetan anticyclone. Such splitting of the anticyclone has been noted by He et al. (1987) and Dunkerton (1995), and the consequent generation of westward propagating anticyclones by Postel and Hitchmann.[1] In this paper, an analysis is presented of 4 northern summers; as will be shown below, similar behavior is evident in each of these years. In potential vorticity (PV), the monsoon anticyclone is manifested primarily as a ''doming'' of the tropopause over the Indian monsoon region [see Figs. 1b and 4 of Dethof et al. (1999); note the slight doming of the 380 K surface near 20°N in Fig. 1, below]; parts of this dome break off at intervals of 1–3 weeks and migrate westward, as far as about 30°E. In terms of Montgomery streamfunction, the eddies extend a little farther downward, but no farther than the midtroposphere.

The data used in this study are the National Centers for Environmental Prediction–National Center for Atmospheric Research reanalysis data (Kalnay et al. 1996). This dataset is global on a horizontal grid of 2.5° latitude by 2.5° longitude (73 by 144 grid points), available 6-hourly on 17 standard pressure levels. We use here 4 yr of data from 1987 to 1990. July monthly mean fields averaged over these 4 yr are used to represent the July (summer) climatology.

For the region of interest (~30°N), the isentropic surfaces that correspond to the upper troposphere (between roughly 300 and 100 mb) are 350 to 390 K. This may be seen from Fig. 1, where the time mean potential temperature (θ) field for July is zonally averaged between 40° and 100°E. In what follows, we chose the 370 K isentropic surface as the representative isentropic level to illustrate upper-troposphere behavior.

The mean July PV field at 370 K is shown in Fig. 2. Potential vorticity $P = -g\zeta_{a\theta}/(\partial p/\partial\theta)$, where $\zeta_{a\theta}$ is the vertical component (in θ coordinates) of absolute vorticity, g is gravitational acceleration, and p is pressure, is given in potential vorticity units (PVUs), where 1 PVU $= 10^{-6}$ K kg^{-1} m^2 s^{-1}. Tropospheric values of PV are generally below about 1.5 PVU. From Fig. 2 we see that the upper-tropospheric anticyclone over south and southwest Asia is characterized by low PV, considerably lower than at other places at the same latitude on this isentropic surface. Elsewhere, such low PV values are found only within a few degrees of the equator. As is also evident in Fig. 1, this anticyclone is manifested by a dome in the tropopause over the monsoon region.

The Montgomery streamfunction $M = c_pT + \Phi$, where c_p =1004 J kg^{-1} K^{-1} is the specific heat at constant pressure, T is temperature, and $\Phi = gz$ is the geopotential, is also shown in Fig. 2. The M is dominated by the large-scale anticyclonic circulation in the region of the Asian monsoon, with a weak secondary anticyclonic maximum west of main anticyclone (cf. Dunkerton 1995). In what follows we analyze the characteristics of this anticyclonic circulation in more detail.

[1] G. Postel and M. Hitchman, unpublished paper presented at the Amer. Meteor. Soc.'s *10th Conf. on Atmospheric and Oceanic Waves and Stability,* Big Sky, MT, June 1995.

Corresponding author address: Dr. Jelena M. Popovic, Atmospheric Studies Group, Earth Tech, 196 Baker Avenue, Concord, MA 01742.
E-mail: jelena@alum.mit.edu

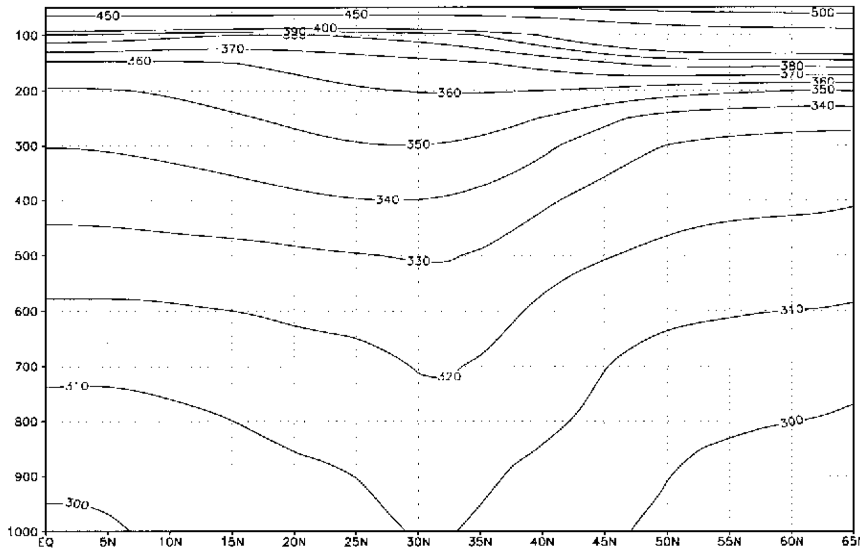

FIG. 1. Jul 1987–90 mean potential temperature zonally averaged from 40° to 100°E.

2. Subseasonal evolution

In order to illustrate the time evolution, we present in Figs. 3–6 Hovmöller diagrams of PV and *M* at 370 K for the latitudinal band between 25° and 35°N from 1 June to 1 September of the 4 analyzed years. Years 1987 and 1988 are shown in Figs. 3 and 5, while years 1989 and 1990 are shown in Figs. 4 and 6. These plots show the overall pattern of PV and *M* fields throughout the summer on the selected isentropic level of 370 K. Time is on the ordinate, increasing downward, and longitude, from 20°W to 180°, is on the abscissa. In the PV figures, PV on the 370-K surface is plotted such that only the tropospheric values between 0 and 1.5 PVU are color shaded. Higher values are not shaded so that white areas on the plots represent high PV stratospheric air.

From Figs. 3 and 4 we may immediately notice the following: from June to August, at the isentropic level of 370 K, the bulk of the low PV (tropospheric) air in the 25°–35°N latitudinal band is confined between the longitudes of 20° and 120°E. There are periods of time when the low PV strip remained stationary, indicating the presence of the anticyclonic vortex at almost the same position for several days. However, we also see a

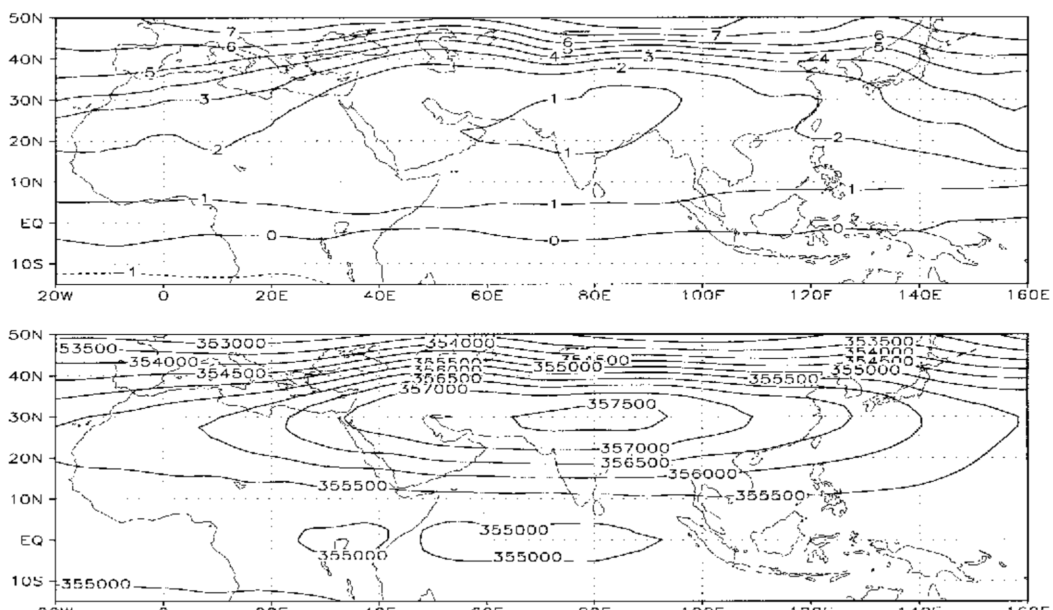

FIG. 2. Jul 1987–90 mean PV (upper) and Montgomery streamfunction (lower) at 370 K.

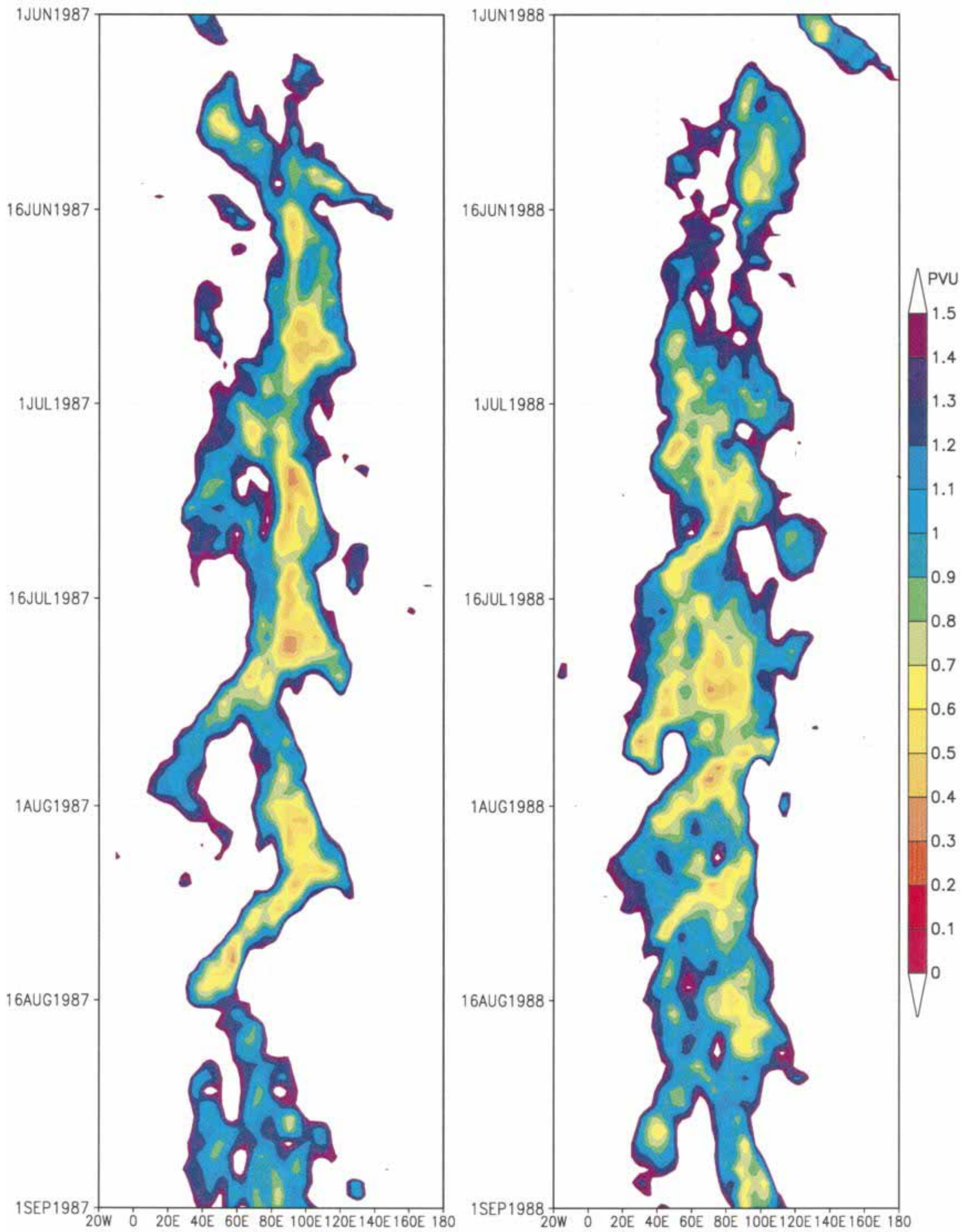

FIG. 3. Hovmöller diagrams of daily mean PV at 370 K averaged over the latitudinal band between 25° and 35°N, from 1 Jun to 1 Sep of 1987 (left) and 1988 (right).

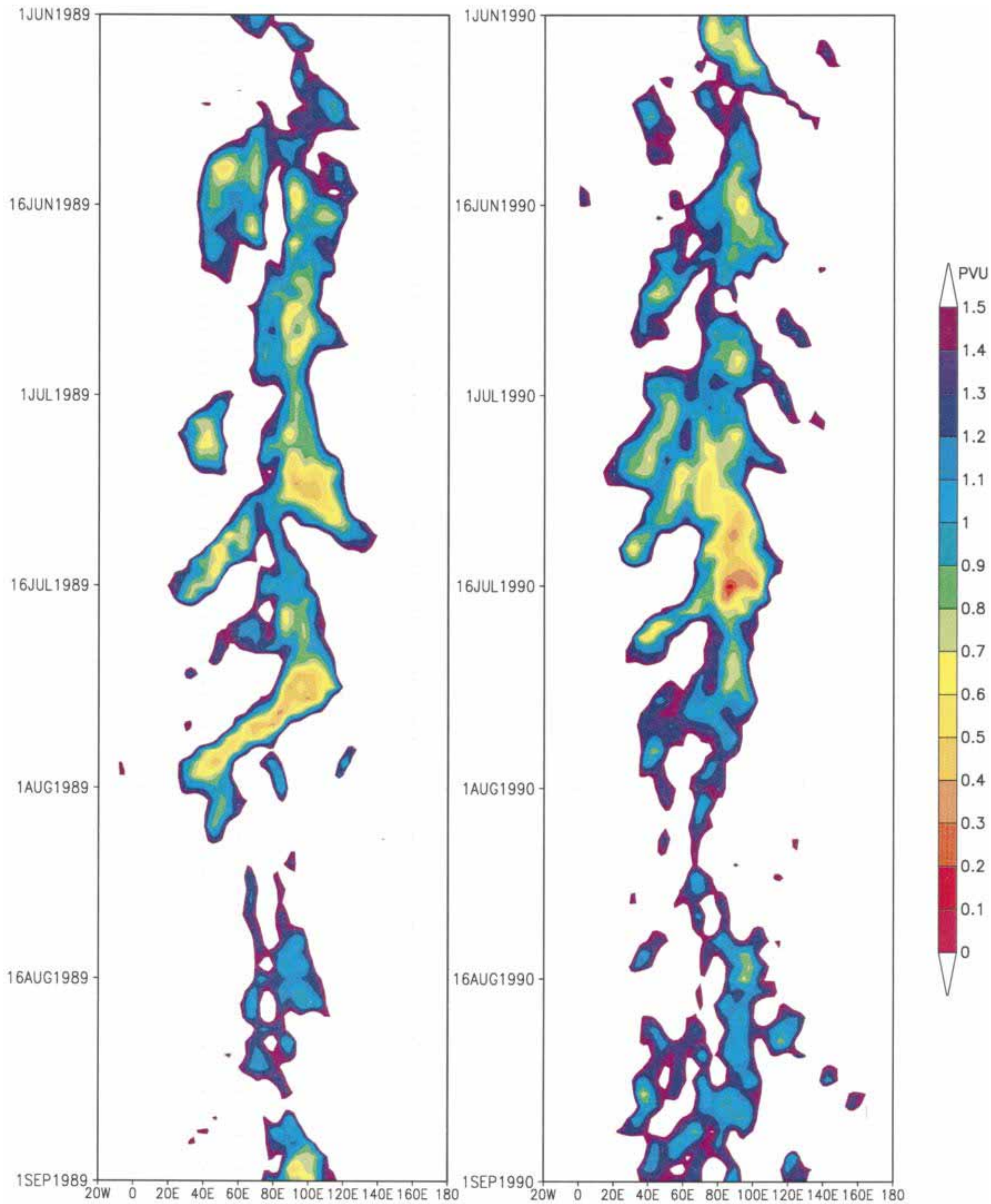

FIG. 4. Hovmöller diagrams of daily mean PV at 370 K averaged over the latitudinal band between 25° and 35°N, from 1 Jun to 1 Sep of 1989 (left) and 1990 (right).

number of prominent westward-propagating features in each summer, and a few eastward-propagating features. Most of the westward moving features coexisted with the almost stationary low PV body, and these were actually the eddy shedding events. The eddies that were shed had widths of typically 1500 km. There were also two westward moving features that did not have a low PV body counterpart on the east (in early August 1987 and late July 1989). In these cases, the anticyclone did not break up but apparently migrated westward as a whole.

The same features are evident in the Montgomery streamfunction (Figs. 5 and 6). Notice that the westward propagating peak sometimes stalls near 50°–60°E, which could explain the secondary peak in the monthly averaged *M* (Fig. 2b; see also Dunkerton 1995).

3. Vertical structure

The phenomenon is relatively shallow. Figure 7 shows the vertical structure of the deviation from the zonal mean of the PV and *M* fields at 0000 UTC on 13 July 1990 [the eddy shedding event from Hsu and Plumb (2000); see their Figs. 14 and 15]. In this figure, vertical cross sections are latitudinally averaged between 25° and 35°N. The negative values of the deviation of PV from the zonal mean, in Fig. 7 arise because the potential vorticity field has much lower values in the region of interest compared to other longitudes. To stress the important areas, the values smaller than -0.5 PVU are shaded. The strongest deviation from the zonal mean is found close to the 370-K level. The magnitude of the deviation from the zonal mean decreases downward, but at the same time the absolute value of the PV field itself decreases as well.

From Fig. 7, eddy shedding in the PV field is confined to the layer between the tropopause ($\sim$380 K) and the 340-K surface. The structure in the *M* field is similar, but extends somewhat deeper. The Montgomery streamfunction has large values in the monsoon anticyclone, so that the deviation from the zonal mean is positive in this area, exceeding 1500 $m^2\ s^{-2}$ near the 370-K surface. Although still shallow, the deeper structure in the *M* field than in PV is to be expected from the invertibility principle. Given a length scale L, the expected vertical scale of the *M* signal associated with a vertically localized PV anomaly is $H \sim fL/N$, where N is the buoyancy frequency and f the Coriolis parameter. In terms of the Rossby radius of deformation L_R, based on tropopause height H_T, given by $L_R = NH_T/f$, we have $H \sim H_T(L/L_R)$. With $H_T = 15$ km, $L_R \approx 3000$ km at latitude 30°N; hence, for motions of scale $L \approx 1500$ km, $H \approx \frac{1}{2}H_T = 7.5$ km.

4. Individual cases

The shedding event of late July 1990 presented in Hsu and Plumb (2000) was perhaps the clearest example of this phenomenon in the 4 years we analyzed. It occurred when the primary vortex achieved an elongated elliptical shape and, presumably, became unstable. The separation of the westward migrating eddy seen in PV maps was confirmed by contour advection calculations. All other events that we found and examined in 4 summers from 1987 to 1990 have many similarities with the 1990 event but some differences as well.

July 1987 was marked by two eddy shedding events. The first one lasted from 5 until 13 July. The second shedding event started on 26 July. From the time sequence of PV and *M* fields at 370 K, with a 24-h interval shown in Fig. 8, we see that the elongated anticyclonic cell centered around 60°E and 30°N on 25 July (upper plot) gradually broke into two entities (26 July—second plot). On 27 and 28 July, the mother vortex remained on the eastern side, building up with time while the daughter vortex moved westward and weakened. This eddy shedding event ended 3 days after the last time plot shown (bottom plot), after which time the low PV of the weakened daughter vortex was advected by the westerly jet and reassimilated into the mother vortex.

In summer 1988, the upper-tropospheric circulation was much stronger than in the other 3 analyzed years (see Fig. 5). For the first 5 days of July, the flow had a strong anticyclonic circulation centered at 60°E and 30°N and also had rather low eccentricity. After 5 July, the anticyclone became weaker. The center itself gradually moved eastward to around 80°E, and the whole system became zonally elongated with a long tongue of anticyclonic PV stretched out eastward in the westerlies on the northern flank of the anticyclone (Fig. 9, upper plot). Subsequently, the main anticyclonic cell remained almost at the same position, losing much of its eccentricity by 12 July (bottom plot). At the same time, on the northeastern flank of the large-scale anticyclonic circulation, the long filament of low PV rolled anticyclonically into a secondary vortex by 11 July (third plot). This eastward filamentation event has been discussed by Dethof et al. (1999), who ascribed the observed behavior to the influence of a midlatitude cyclone immediately to the north of the subtropical jet.

After this event, by mid-July the anticyclone regained its (unusual) strength and large eccentricity and remained like that for the following 10 days. It was particularly strong around 20 July, before another, westward, eddy shedding event occurred around 25 July. At that time, the center of the large-scale anticyclone shifted slightly northeastward and shed a daughter vortex westward.

July 1989 began with a shedding event (see Figs. 4 and 6). The end of June was marked with a weak and eccentric anticyclone that was followed by the eddy shedding. The whole event lasted 1 week. In this case the anticyclone was centered around 80°E before the shedding commenced. When the shedding happened the mother vortex remained on the eastern side with its center as far east as 100°E, while the daughter vortex

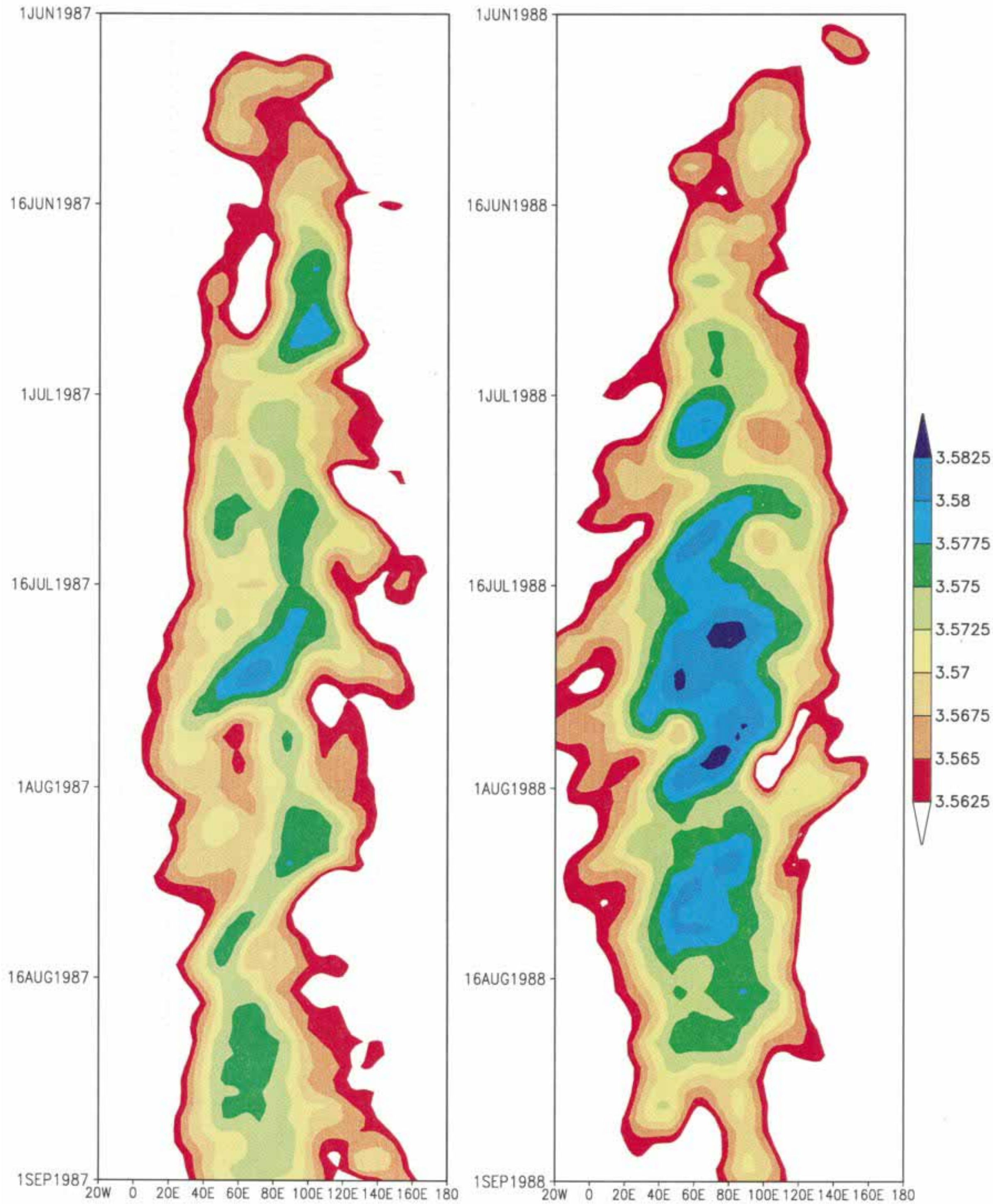

FIG. 5. Hovmöller diagrams of daily mean Montgomery streamfunction at 370 K averaged over the latitudinal band between 25° and 35°N, from 1 Jun to 1 Sep of 1987 (left) and 1988 (right). Units are 10^5 m^2 s^{-2}.

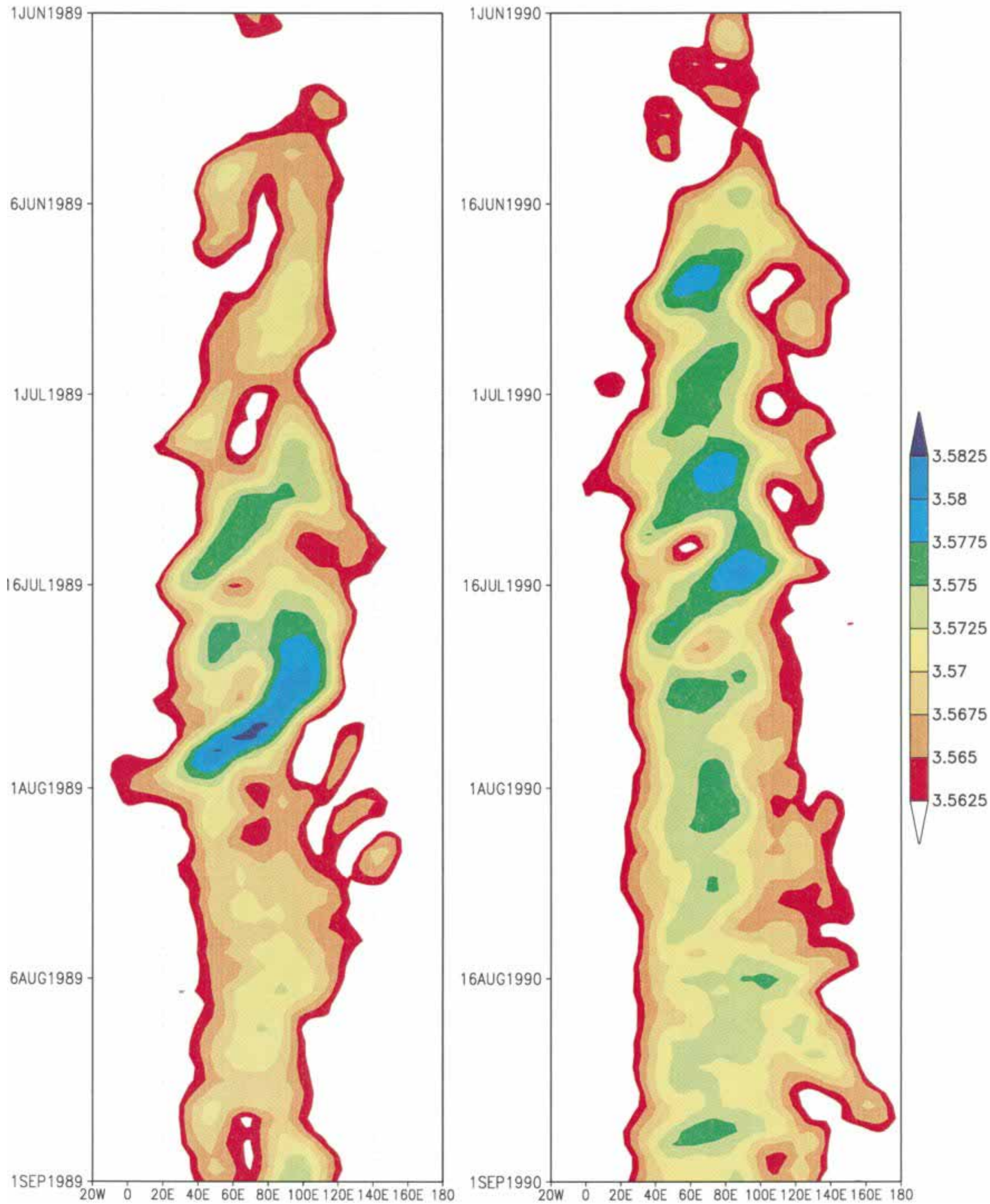

FIG. 6. Hovmöller diagrams of daily mean Montgomery streamfunction at 370 K averaged over the latitudinal band between 25° and 35°N, from 1 Jun to 1 Sep of 1989 (left) and 1990 (right). Units are 10^5 m^2 s^{-2}.

detached on the western side and moved northwestward with its center reaching 40°E. This was clearly seen in both PV and M fields, which were very well spatially correlated in this case.

The strong anticyclone usually starts to build up in the first half of June. It is strongest in July and first part of August, and then starts to gradually weaken in the second part of August. We found one shedding event in the month of August, starting on 25 August 1988. Preceding the shedding, the anticyclone was again

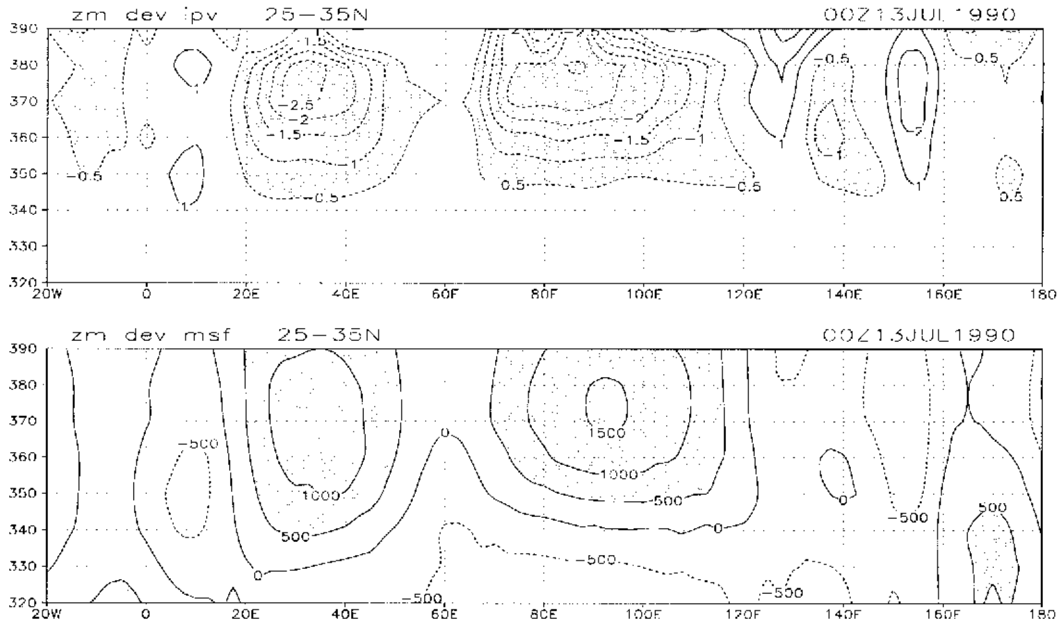

FIG. 7. Vertical cross sections of the perturbation from the zonal mean of PV (upper) and Montgomery streamfunction (lower), latitudinally averaged between 25° and 35°N, at 0000 UTC 13 Jul 1990. The values smaller than −0.5 PVU and greater than 500 $m^2 s^{-2}$ are shaded.

strong and elongated. When the shedding started and as the western cell was moving westward, the whole system weakened and by the end of August, daughter vortex was no longer evident.

One other interesting event happened during August. In August 1987, there was a ten day period when the anticyclone moved westward with its center moving from ~ 90°E on 6 August to ~50°E on 16 August (Fig. 10). The low values of PV were confined to the area of the maximum values of the streamfunction. This case was similar to the one that happened in late July 1989 (Figs. 4 and 6). After the two cells coexisted together for 10 days from mid-July 1989, with the western one gradually moving northward and weakening, by 26 July, the eastern cell remained to dominate the field. This remaining strong anticyclone moved westward, losing much of its strength by the end of July.

5. Conclusions

The upper-tropospheric divergent flow in the area of the Asian summer monsoon is present during the whole summer, but it varies greatly in form and strength. The most dramatic changes are seen during eddy shedding events, and during the westward propagation of the anticyclone.

Westward eddy shedding from the Tibetan anticyclone seems to be a common phenomenon. It is found every summer (at least in the 1987–1990 data studied here as well as in year 1996 not discussed here). We do not find a regular periodicity; events occur a few (2–4) times per summer. The duration of each event is typically 4–8 days. In most cases when the anticyclone strengthens and elongates the shedding happens 2–3 days later.

Synoptic conditions just prior to westward eddy shedding were quite compatible with those in the shallow water model of Hsu and Plumb (2000). Both in the modeled flow and in the atmosphere, when the low PV air of the anticyclone becomes rather narrow and long enough in the zonal direction, it becomes unstable and sheds eddies (for a very elongated eddy, one would expect parallel flow theory to be valid and thus flow to become dynamically unstable due to the change of sign in PV gradient). Thus, following Hsu and Plumb (2000), we associate the westward shedding with the internal dynamics of the monsoon anticyclone. This contrasts with other events—which are less intense, in the periods we have analyzed—in which low PV air is extruded eastward, along the subtropical westerly jet; the case of July 1988 (Fig. 9) was the clearest example we found of this behavior. Dethof et al. (1999) ascribed this behavior to the interaction of the anticyclone PV with eastward traveling midlatitude disturbances.

For the most part, we have seen that these disturbances are confined to the upper troposphere ($\theta > 340$ K). It is not clear whether they have any impact on the monsoon circulation as a whole. The shedding events could conceivably influence the deep monsoon circulation through the divergent flow; however, in a preliminary study we could find no clear relationship between these events and lower tropospheric behavior.

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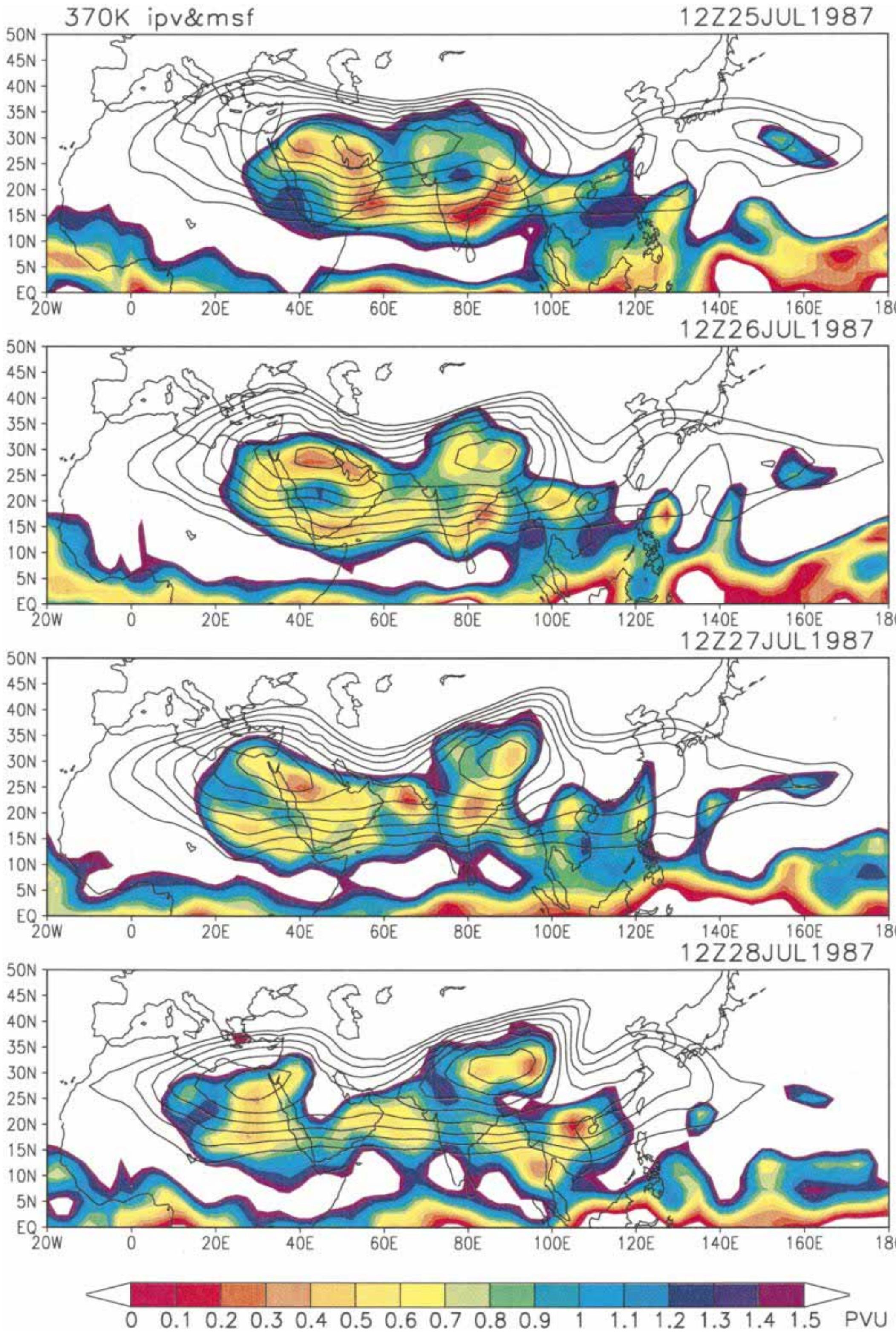

FIG. 8. Time sequence of PV (color shading) and Montgomery streamfunction (line contours) fields at 370 K given in successive 24-h intervals from 1200 UTC 25 Jul (top) to 1200 UTC 28 Jul 1987 (bottom). Minimum value of displayed contour lines is 356 000 m^2 s^{-2}. Contour line interval is 250 m^2 s^{-2}.

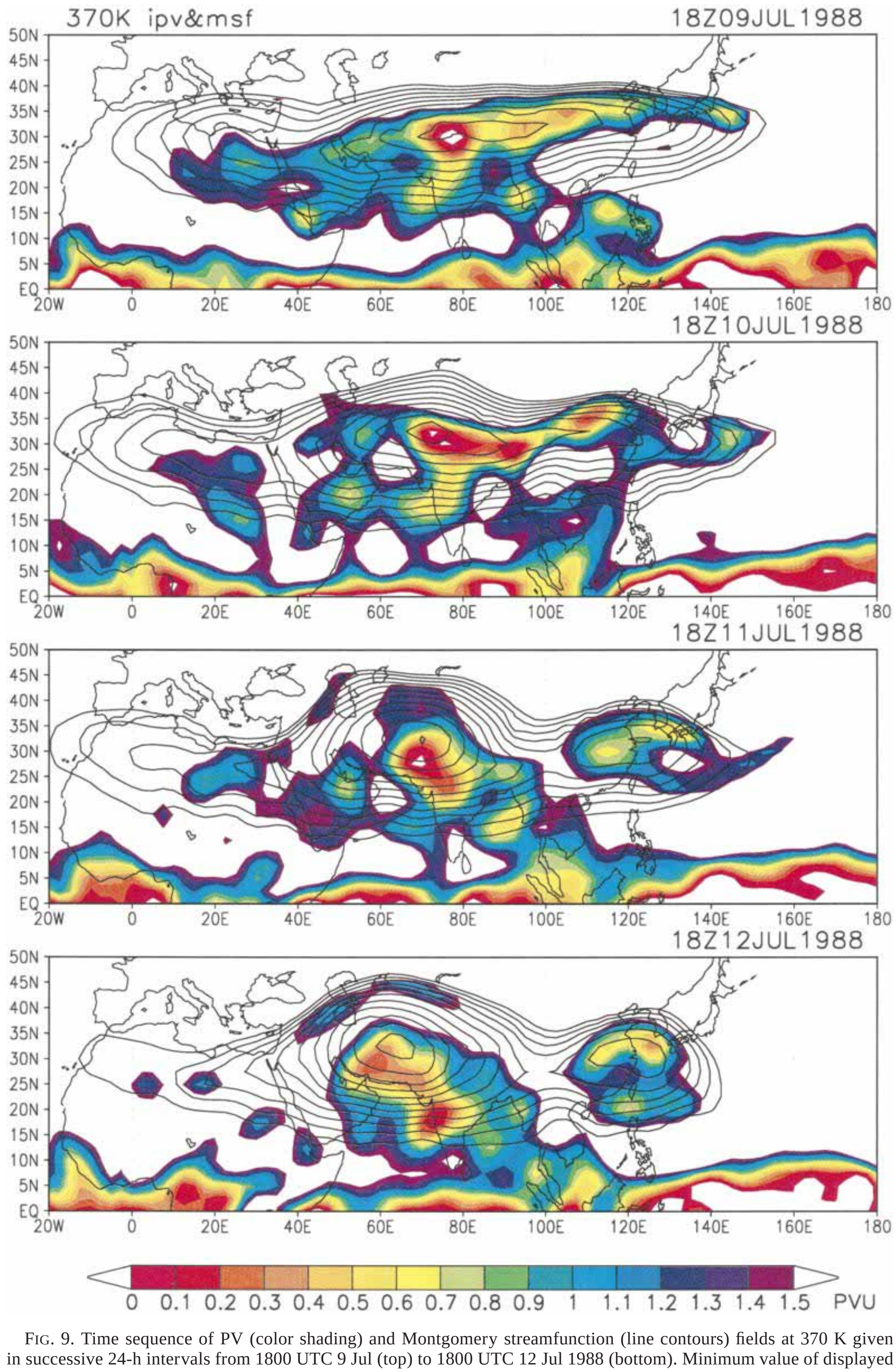

FIG. 9. Time sequence of PV (color shading) and Montgomery streamfunction (line contours) fields at 370 K given in successive 24-h intervals from 1800 UTC 9 Jul (top) to 1800 UTC 12 Jul 1988 (bottom). Minimum value of displayed contour lines is 356 000 $m^2 s^{-2}$. Contour line interval is 250 $m^2 s^{-2}$.

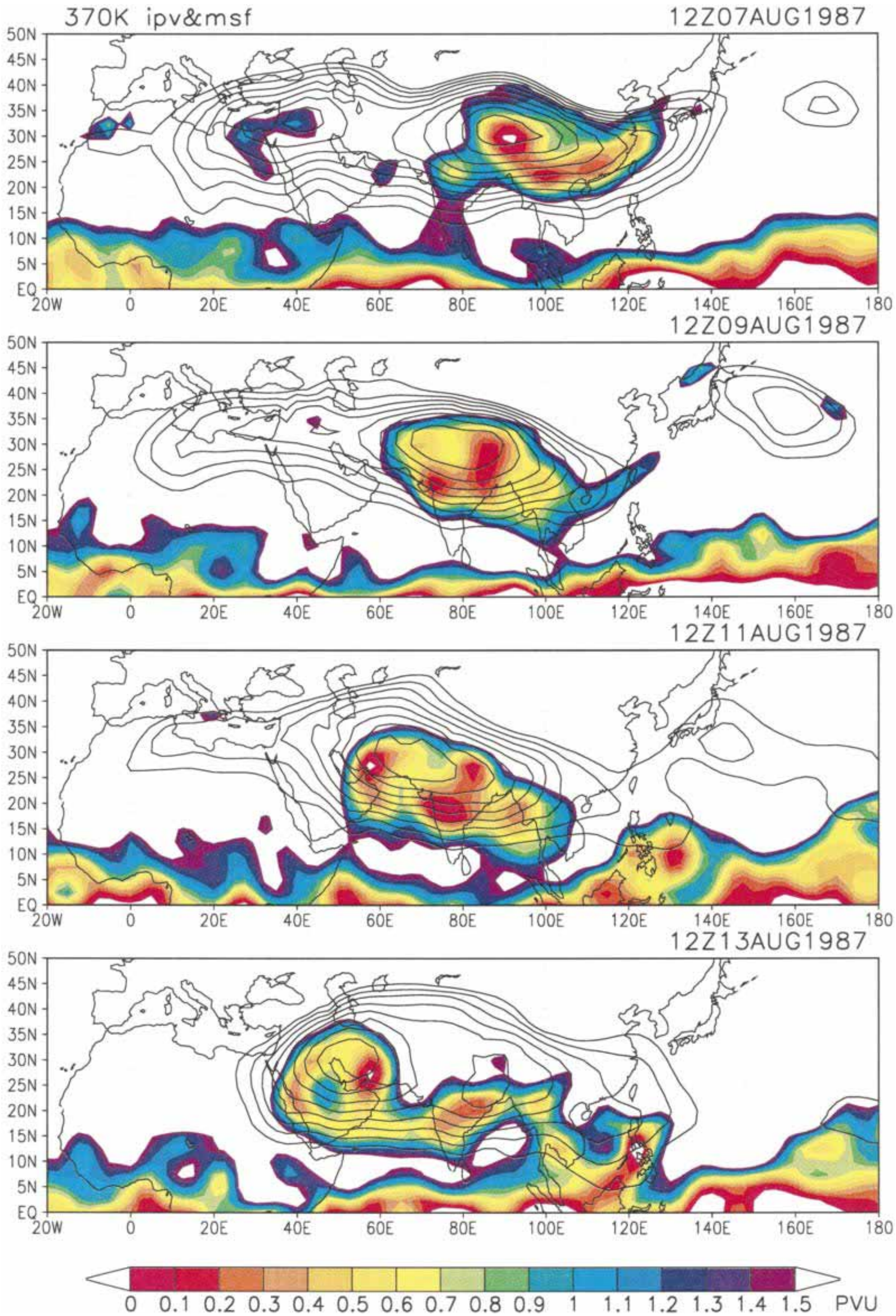

FIG. 10. Time sequence of PV (color shading) and Montgomery streamfunction (line contours) fields at 370 K given in successive 48-h intervals from 1200 UTC Aug (top) to 1200 UTC 13 Aug 1987 (bottom). Minimum value of displayed contour lines is 356 000 $m^2\ s^{-2}$. Contour line interval is 250 $m^2 s^{-2}$

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