The Development of A Mei-Yu Disturbance with Strong Vertical Coupling

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1. Introduction

The Mei-yu ("Baiu" in Japan and "Changma" in Korea) is a quasi-stationary system that defines the East Asian summer monsoon trough during June and July. It produces heavy rainfall along a quasi-zonal belt centered near 30°N over China and tilts slightly east-northeastward toward Korea and Japan. The heavy rainfall is mostly associated with eastward moving meso- to synoptic-scale disturbances that develop frequently along the belt. In this work, the life cycle of a disturbance that developed from the eastern edge of the Tibetan Plateau and propagate eastward to the western Pacific is studied by analyzing its potential vorticity. This disturbance developed into one of the most intense Mei-Yu systems ever observed in June when it moved out to the East China Sea. The result will demonstrate the important role played by the vertical coupling with middle and upper level midlatitude systems in the intensification of this disturbance.

The data used is the 00Z and 12Z ECMWF analysis at 2.5° x 2.5° grids for the period June 17 - 25, 1992. The potential vorticity (PV) and its anomaly from the June 17 00Z - June 23 00Z mean (PV') are computed, and the diagnostic method of inverting the Ertel's PV anomaly (EPV' or q') with the nonlinear balance equation, designed by Davis and Emanuel (1991), is used for the data analysis.

2. Results

Fig. 1 is a time series of the 850 hPa wind and geopotential height at 00Z over the region 25°N-40°N, 100°E-160°E. A disturbance (labeled *A*) first showed up near 28°N, 100°E on 00Z 6/17 and moved slowly eastward. After 6/21 it appeared to split into two centers, with the main center moving rapidly eastward together with intensification. The maximum intensity was reached between 6/23 and 6/24 near Japan, afterward it weakened in the western Pacific. This eastward propagation also corresponded to an eastward withdrawal of the subtropical ridge whose center was near the dateline. An interesting question is whether the eastward movement and intensification, which signified an eastward extension and strengthening of the East Asian monsoon trough, was due to the eastward withdrawal of the subtropical ridge.

Fig. 2 shows the EPV' along the vertical (850 hPa- 200 hPa)-longitude section at 32.5°N, for 6/18 00Z and 6/21 00Z to 6/24 00Z. At 6/18 00Z system *A* is observed near its source region on the eastern edge of the

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Tibetan Plateau. Between 6/18 12Z and 6/20 12Z (not shown) it moved down from 500 hPa along the eastern Plateau slope. The split into two 850 hPa centers can be seen at 6/21 00Z. At 6/21 00Z, another disturbance (*B*) began to emerge on the eastern Tibetan Plateau. By 6/21 12Z it moved eastward off the plateau with center stayed near 500 hPa. Twelve hours latter, at 6/22 00Z a third disturbance (*C*) appeared above 300 hPa, while *A* at the lower level and *B* at the middle level continued their eastward movement. The three systems became increasingly coupled through 6/23 00Z, when the maximum intensity was reached. Thereafter, the vertical link began to decrease, and by 6/24 00Z *A* was de-coupled from upper levels.

Fig. 3 is the 1000 hPa-100 hPa section of the height anomaly that is attributed to the effect of the nearly saturated (R.H. ≥ 70%) 850-500 hPa air with q' ≥0 through the balance. The EPV' of this air will be denoted as q'_{sat}. The correspondence between system *A* and the lower height indicates the close association of the disturbance with latent heat release. This heating effect reached maximum at 6/21 00Z and weakened afterwards. By 6/22 00Z, when system *B* moved out of the Tibetan Plateau, a diabatic heating maximum at 500 hPa near its center can be seen. The merging of the 500 hPa center and the lower level system was complete by 6/23 00Z, when the disturbance was at its maximum intensity. The diabatic effect increased for another 12 h, then decreased rapidly by 6/24 00Z.

The effect of the unsaturated air (whose EPV' is denoted by q'_{un}, representing 850-500 hPa EPV' other than q'_{sat}) generally opposed the disturbance low with positive anomalies over the disturbance as shown in the 6/21 00Z example, but it became contributive to the low-level low center by 6/24 00Z (Fig. 4). An examination of the satellite image (not shown) confirmed that this was the time the cloud area shrunk around the low-level cyclonic center. The effect of the boundary layer had strong diurnal variations, with the morning (00Z) opposing the low and the evening (12Z) contributing to it. However, when the system moved over ocean, even the morning effect was contributing (Fig. 5 shows the contribution of boundary layer potential temperature anomaly, T'b).

The effect of the vertical coupling is evaluated in Fig. 6. Here the total nonlinear 12 h height tendency following the 850 hPa *A* center is compared with those due exclusively to air above tropopause (300-100 hPa, q'_{tp}) and saturated air (q'_{wet}, the EPV' for all 850-500 hPa air with R.H.≥ 70%), if they each acts alone. Both provide similar forcings of the height tendency and their sum nearly explain the total through 6/23 00Z, the time of maximum disturbance intensity. Afterwards the sum exceeds the total as the latent heat release and the tropopause folding were both so intense that, in the absence of other processes, additional nonlinear effects occurred. Both effects quickly diminished on 6/24 00Z, time of the vertical decoupling, and became detrimental 12 hours later.

The nonlinear height tendency for the cross-section along 32.5°N for total, q'_{tp} and q'_{wet} at three selected synoptic times are shown in Fig. 7. At 6/19 12Z, the lower troposphere total tendency downstream of System *A* was negative, corresponding to the eastward movement of *A* from the edge of the Plateau. At that time the tendency due to q'_{tp} was positive while that due to q'_{wet} was negative. This suggests that the tropopause air was opposing the actual tendency while the latent heat effect was helping. At 6/23 00Z, the starting time of the vertical coupling, the tendency of q'_{tp} was negative and with a center at 1000 hPa. Although this center did not completely coincide with the surface center of the total tendency, the tropopause air was clearly contributing to the low-level height tendency. The tendency of q'_{wet} was in good agreement with the total. By 6/23 12Z, q'_{tp} and q'_{wet} tendency centers both coincided with the total tendency center, this is the time that contribution from the tropopause air to the lowering of the 850 hPa height center reached maximum, and appears to explain that this is the time *A* reached maximum intensity.

The total nonlinearly-balanced vertical motion and those calculated from q'_{tp}, q'_{wet} and q'_{dry} (850-500 hPa air with R.H. < 70%) at five selected times are shown in Fig. 8. At 6/18 12Z the total vertical velocity was upward, so were those due to q'_{wet} and q'_{dry}, but the vertical motion due to q'_{tp} was downward, again indicating that in the initial stage the tropopause air contributed negatively to the development. By 6/22 00Z, the time of the appearance of the upper level system *C*, the tropopause effect was clearly upward and nearly represent the total upward motion. At this time both q'_{wet} and q'_{dry} gave very little vertical motion. Twelve hours latter, the tropopause air remained important for the rising motion, while the moist air appeared to contribute again, apparently because *A* has moved over the East China Sea. These two effects appeared to give rise to two rising motion centers. The drier air remained nearly dormant.

By 6/23 00Z, the moist air exerts increased effect and gave rise to an upward motion center in the same position as the total center. This was the beginning of vertical coupling, and the tropopause effect appeared to interact with the moist air effect, with mutually adjustments that brought the upward motion centers closer. The upward motion reaches maximum at 6/23 12Z when effects of all air, including that due to the drier air, gave similar patterns with centers in all panels coincide.

In summary, in addition to the diabatic heating, the vertical coupling through the troposphere and lower stratosphere was an important factor to the development of this Mei-yu disturbance. Since the upper level disturbance originated from higher latitudes, the development was more than just a result of the subtropical ridge withdrawal.

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Reference

Davis, C. A. and K. A. Emanuel, 1991: Potential Vorticity Diagnostics of Cyclogenesis. *Mon. Wea. Rev.*, **119**, 1929-1953.

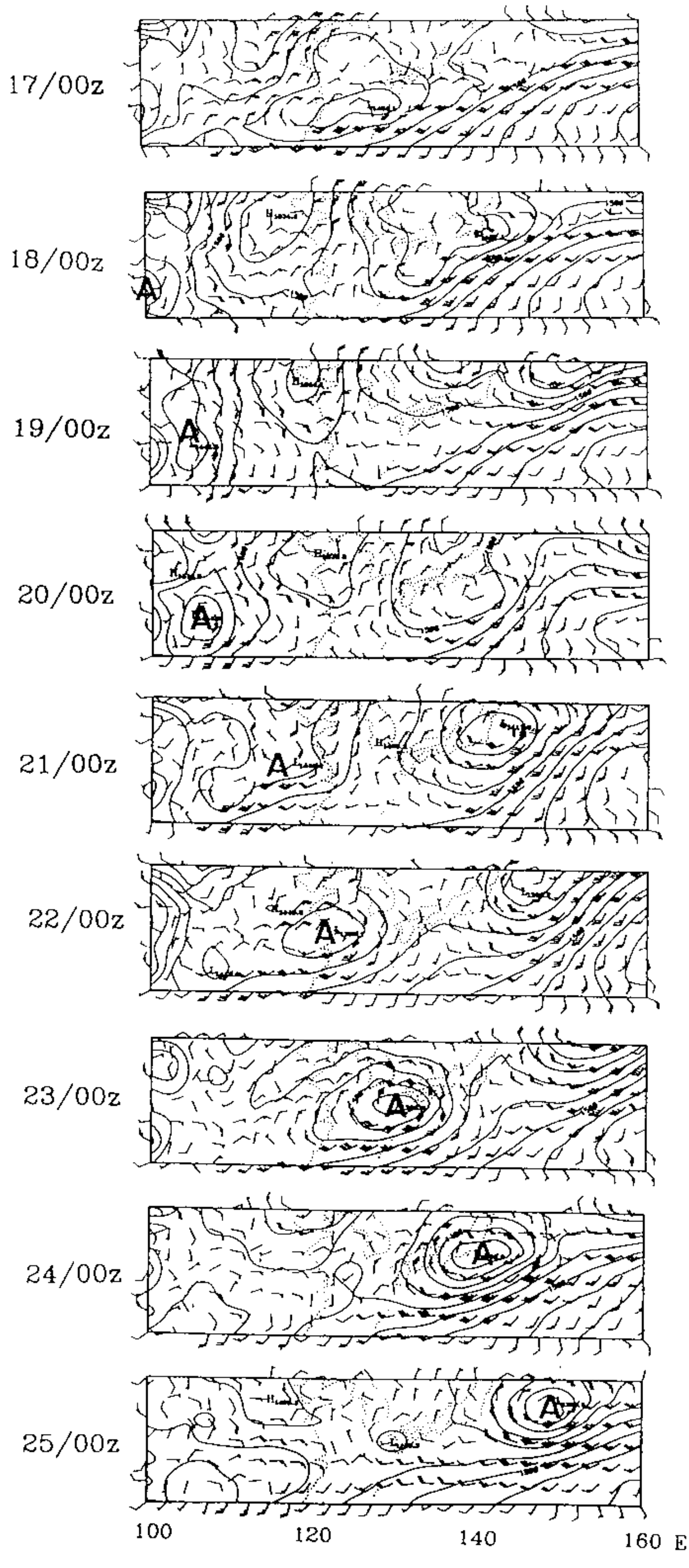

Fig. 1 850 hPa wind and height between 25°N-40°N for 00Z 17-25 June.

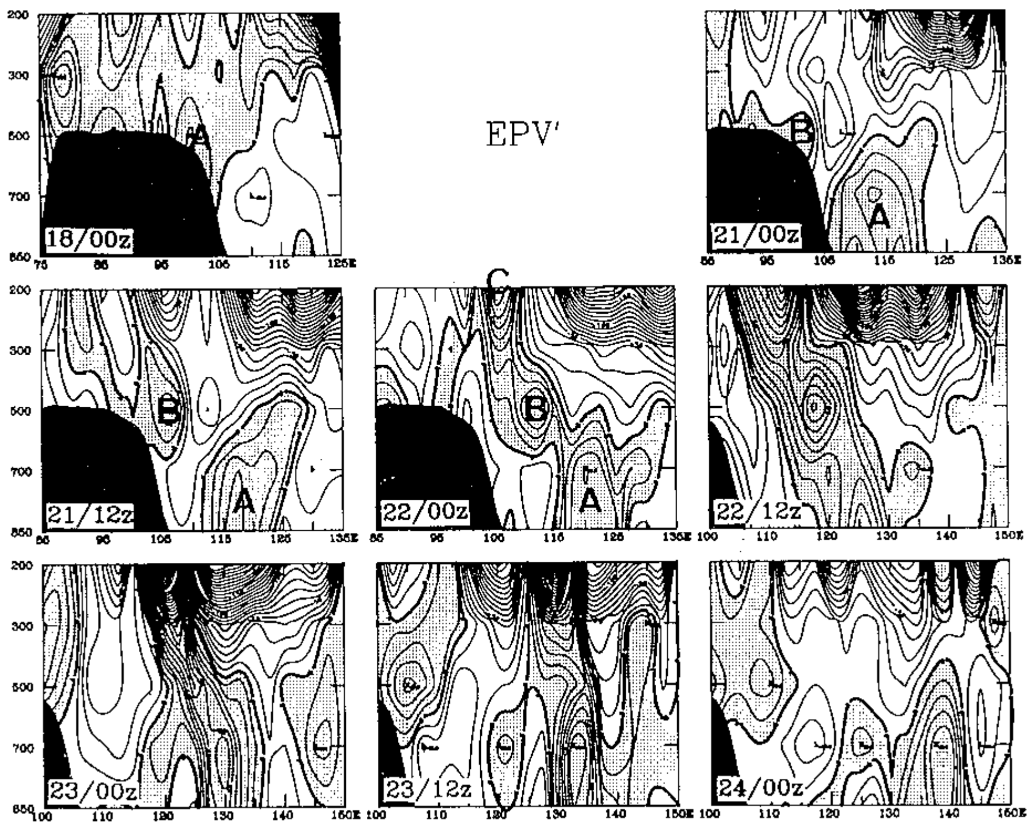

Fig. 2 Vertical (850 - 200 hPa) section of EPV' (interval 0.1 PVU) at 32.5°N for 6/18 00Z and 6/21 00Z to 6/24 00Z. The longitudinal domain shifts with the disturbance center.

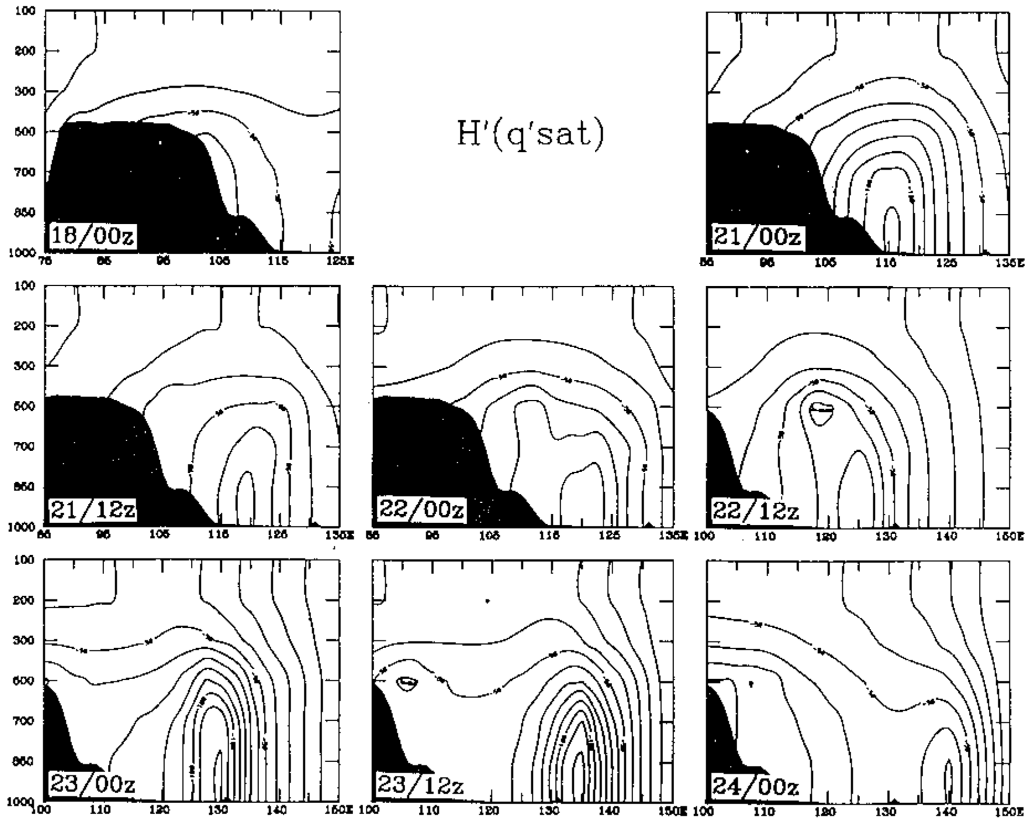

Fig. 3 Same as Fig. 2 except for the contribution of q'_{sat} to height anomaly (interval 10 m). The vertical coordinate is 1000 - 100 hPa.

H'(q'un)

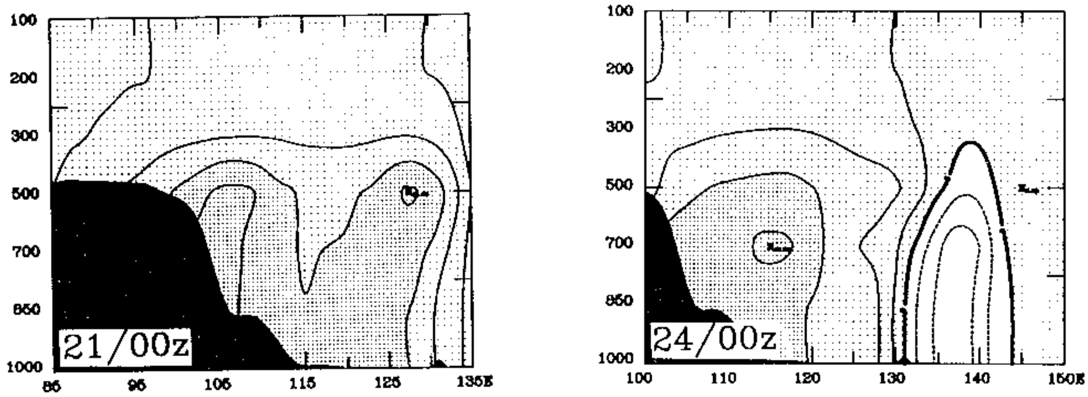

Fig. 4 As Fig. 3 except for the contribution of q'_{un} to height anomaly for 6/21 00Z and 6/24 00Z.

H'(T'b)

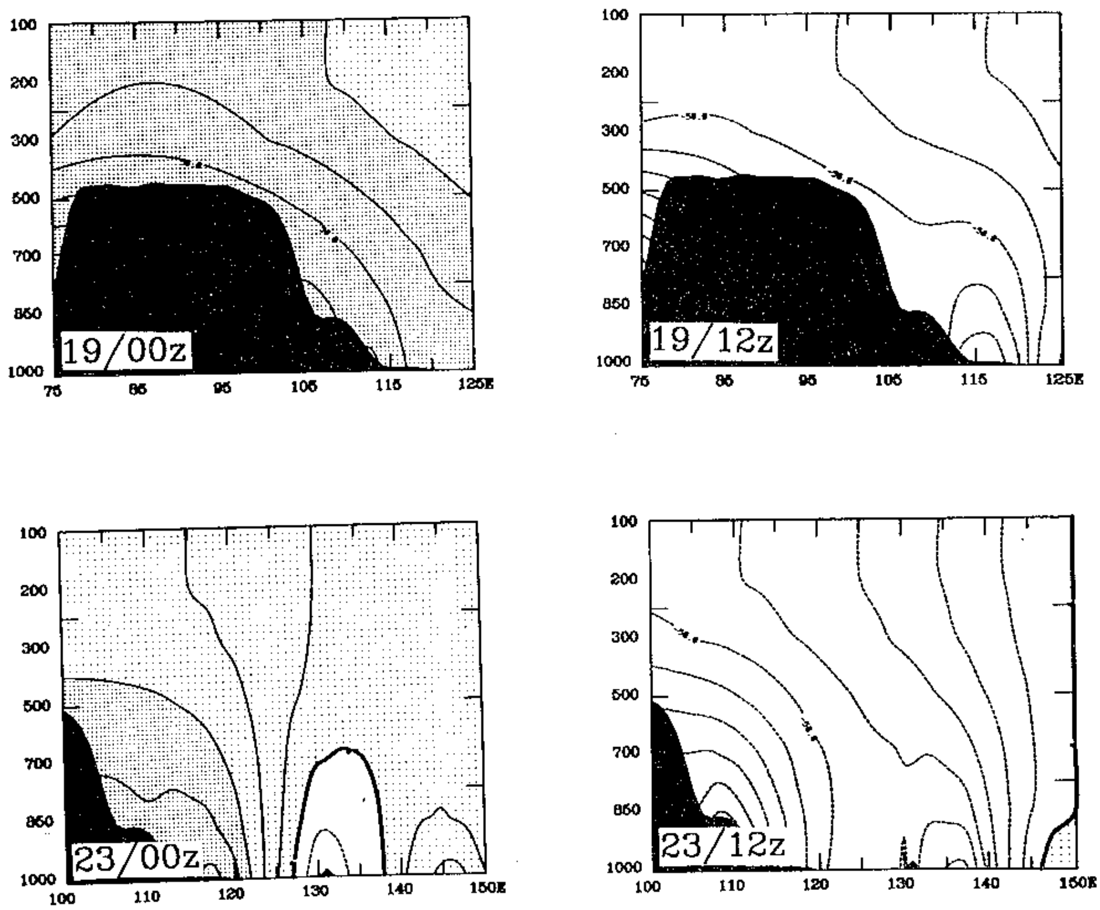

Fig. 5 As Fig. 3 except for the contribution of boundary layer potential temperature anomaly to height anomaly for 6/19 and 6/23, 00Z and 12Z.

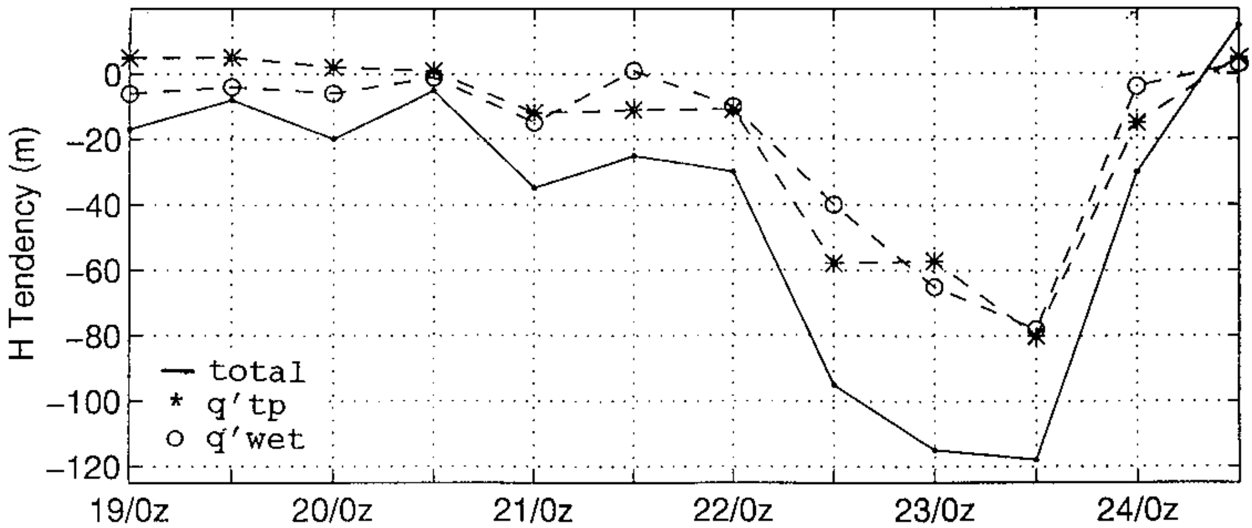

Fig. 6 Nonlinearly balanced 12h height tendency at the 850 hPa low center associated with total (solid), q'_{tp} (dash-*) and q'_{wet} (dash-circle).

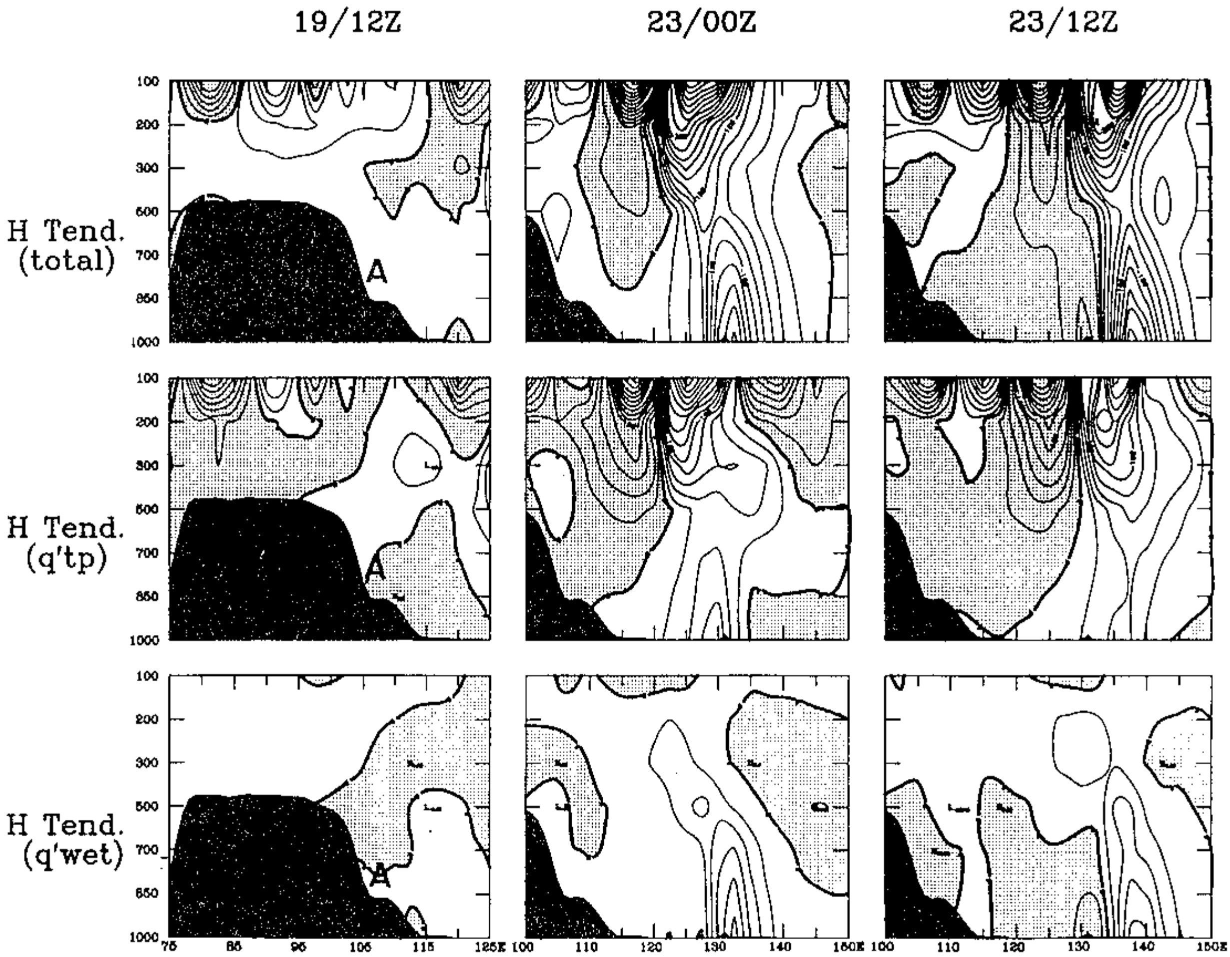

Fig. 7 As Fig. 3 except for the height tendency (interval 20 m/12 h) associated with total (top), q'_{tp} (middle) and q'_{wet} (bottom), for 6/19 00Z, 6/23 00Z and 6/23 12Z. The position of disturbance A is marked on the 6/19 00Z charts.

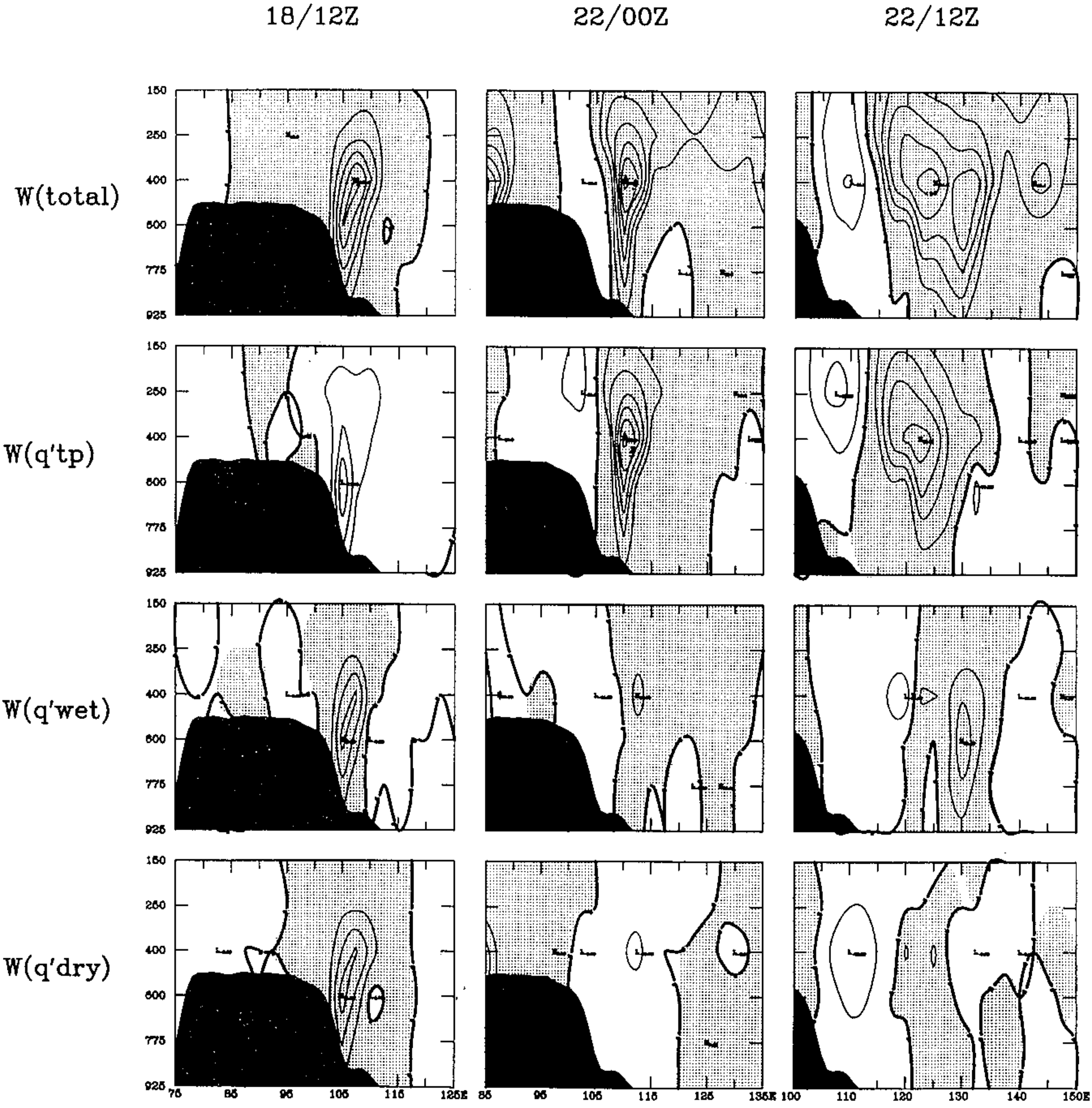

Fig. 8 Vertical (925 - 150 hPa) section of vertical velocity (interval 2 cm/s) associated with total (top), q'_{tp} (upper middle), q'_{wet} (lower middle) and q'_{dry} (bottom), for 6/18 12Z, 6/22 00Z, 6/22 12Z, 6/23 00Z and 6/23 12Z.

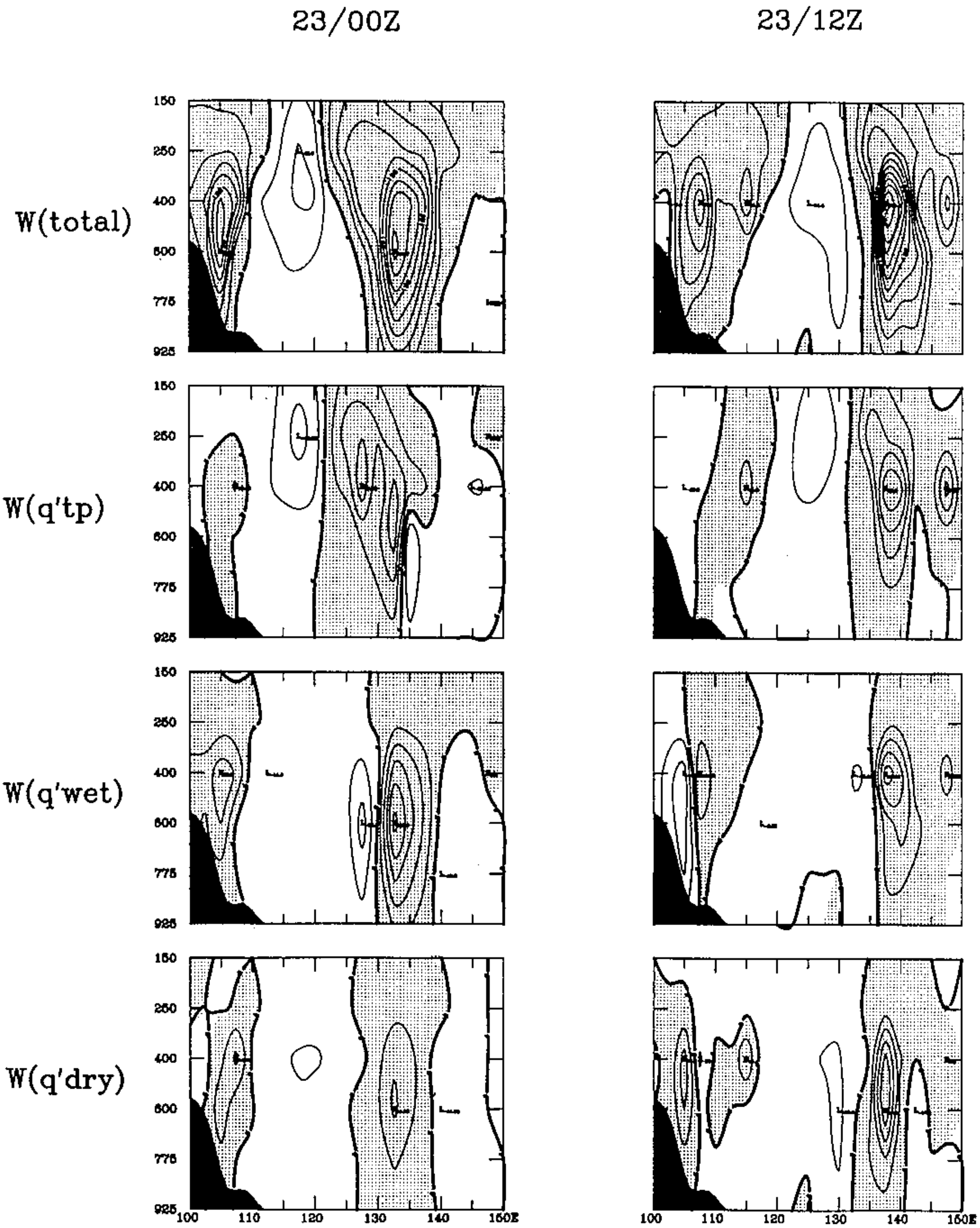

Fig. 8 (continued)