Tropical Plumes and Extreme Precipitation in Subtropical and Tropical West Africa: Part II. Large-scale Dynamic and Diabatic Processes

PETER KNIPPERTZ* and JONATHAN E. MARTIN

Department of Atmospheric & Oceanic Sciences, University of Wisconsin–Madison, Madison, Wisconsin

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*Corresponding author: Peter Knippertz, Department of Atmospheric & Oceanic Sciences, University of Wisconsin–Madison, 1225 West Dayton Street, Madison, WI 53706, USA. E-mail: pknipp@aos.wisc.edu
SUMMARY

The present observational study addresses the respective roles of large-scale tropical–extratropical interactions and diabatic processes leading to the generation of the deep upper-level troughs, characteristically involved in cases of extreme precipitation and tropical plume formation over West Africa during the cool season (see Part I of this series). The elongated, mostly positively tilted potential vorticity (PV) streamers associated with the observed troughs form as a result of an equatorward transport of high-PV air downstream of a large ridge over the North Atlantic. The rapid amplification of the ridge is achieved through a combination of negative horizontal PV advection and diabatic reduction of upper-level PV through latent heating within a cloud band that forms in connection with an explosive baroclinic development near the east coast of North America. The latter is promoted by a strong upper-level jet of low dynamic stability and the feeding of warm, moist air into the cyclo- and frontogenetic region by a low-level jet over the western North Atlantic. Diabatically enforced high-amplitude ridging near the west coast of North America supports the development near the east coast through positive momentum advection into the downstream jet and a deepening of the downstream trough. The acceleration of the participating extratropical jets is also fuelled by momentum fluxes and convective outflow from the (sub-) Tropics. Strong dynamically and diabatically generated upper-level convergence and an enhancement of trade winds are observed over the North Atlantic (and the South Atlantic in one case) during the development.


1. INTRODUCTION

In a companion paper by Knippertz and Martin (2004; KM hereafter) the synoptic evolutions of three extreme precipitation events affecting (sub-) tropical West Africa during the extended cool season (October–April) have been analyzed on the basis of observational and model data. All cases were associated with tropical plume (TP) formation, strong upper-level subtropical jet (STJ) streaks along these plumes and upper-level troughs to the west of the TPs/precipitation zones. While case I (9–11January 2002) brought extraordinary dry season precipitation to the northern West African Tropics (Senegal, Mauritania, Mali), case II (30 March–1 April 2002) produced unusually heavy rainfalls at the year-round arid subtropical southern side of the Atlas chain in western North Africa. Case II has also been investigated with respect to hydrological impacts (Fink and Knippertz 2003) and the dynamics of TP formation (Knippertz 2004). Case III (20–23October 2003) affected the entire region from Senegal to the Algerian Atlas Mountains. As shown by KM, low- to mid-level moisture export from the deep Tropics and the resulting potential instability are crucial for the extreme events. The January and March cases involved a succession of two deep troughs to the west of West Africa within 3–5 days. The first did not produce significant precipitation, but initiated the poleward moisture transport, while the second generated the ascent required to release the potential instability. Such a prior disturbance is not necessary during the October case, when the monsoon layer over Africa is shifted to the north. In all three cases, additional uplift is related to the strongly accelerating, dynamically unstable (and therefore highly divergent) STJ. The strong upper-level convergence and subsidence to the west of the upper-level troughs lead to a strengthening of the low-level trade winds (trade surge) and finally to enhanced tropical convection that supports the acceleration of the STJ.
In the present study, observational data are used to investigate the large-scale dynamics and diabatic processes in the Tropics and extratropics that lead to the penetration of deep extratropical upper-level troughs into the (sub-) Tropics during the cases described by KM. Following Hoskins et al. (1985), maps of potential vorticity (PV) on isentropic levels are used in the analysis. It will be shown that, despite the different seasons and geographical locations of the extreme precipitation, the three cases reveal similarities in the large-scale upstream flow, including rapid baroclinic developments over the western North Atlantic several days before the formation of the deep troughs near Africa. Explosively deepening surface cyclones (termed “bombs”) occur frequently in this region during wintertime and usually form downstream of a mobile trough and within or poleward of the maximum westerlies (Sanders and Gyakum 1980; Roebber 1984). Factors that support such developments include increasingly unbalanced flow, explosive jet acceleration, widespread divergence at upper-levels, the superposition of the secondary circulations in the entrance and exit regions of two upper-level jet streaks and the formation of a strong poleward directed low-level jet (LLJ; Uccellini et al. 1984; Uccellini and Kocin 1987; Uccellini 1999). Other processes that contribute to rapid deepening are strong sensible and latent heat fluxes, a decrease in vertical stability and the release of latent heat, usually producing a divergent “outflow jet” to the northeast of the storm centre and an amplification of the downstream ridge (Cammas et al. 1994; Stoelinga 1996; Uccellini 1999).

From a PV perspective this amplification can be viewed as a combination of advection of low-PV air through the jet flow ahead of the upper-level trough and diabatic reduction of PV at the tropopause through latent heating (Hoskins et al. 1985, chapter 7; Stoelinga 1996; Massacand et al. 2001; Posselt and Martin 2004). In a modelling study Massacand et al. (2001) show that the latent heating within an upstream frontogenetic band affects the equatorward transport of stratospheric PV to the east of the amplified ridge and is therefore crucial for the formation of a downstream PV streamer. Such an effect has also been observed in the context of the modification of upper-level structures by low PV originating in the outflow of tropical cyclones during their transition to extratropical systems (Jones et al. 2003). Thornicroft and Jones (2000) suggest that downstream anticyclonic wave breaking (AWB) might be promoted through the low-PV outflow. AWB is generally associated with the penetration of a positively tilted positive PV anomaly to lower latitudes and the cut-off of a negative PV anomaly and an associated blocking event in the midlatitudes (Thornicroft et al. 1993; Pelly and Hoskins 2003).

The aim of the present paper is to demonstrate the importance of upstream jets, explosive cyclogenesis, diabatic PV reduction and ridge amplification for the generation of the troughs involved in the heavy rainfall/TP events over West Africa. The analysis includes the evolution of the participating jets and PV waves over upstream North America and addresses the question of a pre-conditioning of the atmosphere through prior systems (e.g. by producing blocking situations or by enhancing baroclinicity through inducing thermal advection). The main results section of this paper contains separate analyses for each of the three cases (sections 3–5). It is preceded by an overview over the employed data (section 2) and it is followed by the main conclusions (section 6). For the sake of brevity, case I receives a fully detailed description, while the focus for the other two will be on aspects not covered in the description of case I.

2. DATA

For the representation of all large-scale atmospheric variables, twice daily (00 and 12 UTC) European Centre for Medium-Range Weather Forecasts (ECMWF) Tropical Ocean and Global
Analysis (TOGA) data were used with 2.5°×2.5° grid spacing on standard pressure levels (Trenberth 1992). The General Meteorological Package (GEMPAK) was employed for the interpolation to isentropic levels and the computation/display of various meteorological parameters. Total and ageostrophic winds were calculated for isentropic surfaces between 325 and 350 K at 5-K increments in order to analyze both the polar and the subtropical jet (c.f. Uccellini et al. 1984). The partitioning of the upper-level wind field into its rotational and divergent components was accomplished by solving the Poisson equations for the stream function and velocity potential on the entire globe using the software package SPHEREPACK (Adams and Swarztrauber 1997). Water vapour (WV) and infrared (IR) images from the geostationary GOES and Meteosat satellites have been obtained through the Space Science and Engineering Center of the University of Wisconsin. Surface weather information for the USA was taken from daily weather maps issued by the National Weather Service and from the Climatological Data published monthly by the National Climatic Data Center. In addition, daily weather maps for the northern hemisphere extratropics and the North Atlantic/European region from the German Weather Service were used. Over the ocean, one-degree daily resolution precipitation data from multi-satellite observations were considered for qualitative estimates (Huffman et al. 2001).

3. SYNOPTIC EVOLUTION OF CASE I (JANUARY 2002)

The extreme precipitation over West Africa during case I is connected to the successive equatorward penetration of two deep upper-level troughs within a few days (KM). The following analysis addresses the large-scale evolution that produced each of the two circulation anomalies separately and concludes with a summary and discussion of the entire case.

(a) Trough 1 (2–6 January 2002)

Figure 1 shows GOES IR images with superimposed streamlines and isotachs at the 335-K isentropic level for three analysis times prior to the development of Trough 1. This level is located near the tropopause at midlatitudes (250–200 mb) and reaches down to below 400 mb in the Tropics. Given the tendency of large-scale atmospheric flow at upper-levels to conserve its potential temperature Θ, it seems more adequate to consider jet evolutions on isentropic than on pressure surfaces (see Uccellini et al. 1984), in particular in situations of tropical–extratropical interactions. At 12 UTC 2 January the large-scale flow pattern in the extratropics is dominated by a ridge over the Rocky Mountains, a SW–NE oriented trough with an axis reaching from Lake Huron to New Mexico and a strong upper-level jet of more than 70 m s⁻¹ over the western North Atlantic (Fig. 1(a)). Clouds and precipitation (up to 50 mm) over the Pacific Northwest suggest a contribution to the amplification of the western ridge through diabatic destruction of upper-tropospheric PV. At upper-levels the trough extends farther south-westward into the equatorial Pacific, resulting in large poleward momentum fluxes into the extratropical jet. Convection is observed in the right entrance region of the jet over Florida (Fig. 1(a)). When the upper-level trough over North America approaches the US East Coast at 00 UTC 4 January a cloud band forms along the anticyclonic flank of the jet that extends into the tropical East Pacific Ocean (Fig. 1(b)). Ageostrophic winds in the entrance region over the Gulf of Mexico reach 32 m s⁻¹ and over the following 12 hours wind speeds in the jet core increase from 79 to 92 m s⁻¹. By 00 UTC 5 January the extratropical portion of the cloud band has thickened and wind speeds reach a rarely observed maximum of 95 m s⁻¹ (98 m s⁻¹ at 330 K) to the southeast of
Newfoundland (Fig. 1(c)). A high-amplitude ridge has formed over the North Atlantic downstream of the jet and Trough 1 begins to penetrate equatorward to the west of West Africa. The explosive intensification of the jet and the cloud band is associated with strong divergence along the jet axis at the 345 K isentropic level, where the wind speed maximum still reaches 87 m s⁻¹ (Fig. 2). A large area of strong north-westerly divergent winds covers the entire ridge over the Atlantic Ocean and forms an elongated NE–SW oriented band of convergence in the highly curved exit region of the jet. Widespread strong ageostrophic winds are observed in the northern portion of the convergent region between 325 and 345 K, reaching a very unusual value of 53 m s⁻¹ (69% of the total wind) at 330 K to the north of the Azores (not shown). Farther south the convergence is enhanced through the divergent outflow from convection over the tropical Atlantic crossing the entrance region of the STJ over Africa. In contrast to the extratropical portion, these divergent winds are more confined to the 345 K level.

The phase of rapid jet intensification is accompanied by explosive cyclogenesis, strong frontogenesis and the massive destruction of upper-level PV through latent heat release within the cloud band shown in Figs. 1(b) and (c). For 12 UTC 4 January, Fig. 3(a) shows the PV distribution at the jet level (335 K), displaying a strong PV gradient along the region of highest wind speeds and a PV ridge farther east over the North Atlantic. At lower levels the jet coincides with a pronounced baroclinic zone (there is a 35 K Θ contrast at 925-mb over the ~2000 km between maritime Canada and the subtropical Atlantic). Two cyclones (black dots in Fig. 3) form as a result of an interaction of this baroclinicity with (1) the PV trough over the East Coast and (2) a cyclonic centre over Labrador (Fig. 3(a)). The grey shading in Fig. 3(a) indicates the diabatic PV tendency in the 200–150 mb layer, which includes or is located just above the 335-K level in the extratropics. The calculation of this parameter follows the method described in Cammas et al. (1994), and Posselt and Martin (2004) and is based on the parameterization of latent heating rates employed by Emanuel et al. (1987). Figure 3(a) shows very intense reduction of upper-level PV within the cloud band over the western North Atlantic. For 00 UTC 5 January, Fig. 3(b) depicts wind vectors, Θ and frontogenesis (calculated in the same way as in KM) at 925 mb. In only 12 hours the baroclinic zone has quickly moved north-eastward towards Greenland and has obtained a cyclonic curvature in association with the deepening of the northern low-pressure system by 23 mb. The southern low has followed the movement of the baroclinic zone without significant intensification. Very pronounced frontogenesis is observed underneath the northern portion of the cloud band between a south-westerly to southerly LLJ of relatively warm (and moist) air over the North Atlantic and a westerly to north-westerly flow of much colder (and drier) air off the North American continent. Underneath the entrance region of the upper-level jet (see Fig. 1(c)) strong cold advection is found in the vicinity of Bermuda. During the next 12 hours the band of clouds, frontogenesis, upper-level divergence and PV destruction continues to move north-eastward (Fig. 3(c)). The southern low-pressure system remains close to the baroclinic zone and intensifies only marginally. The northern low continues to deepen by 22 mb to 944 mb and moves towards the southern tip of Greenland. The combination of ongoing diabatic PV destruction and advection of low PV air into the ridge by the strong south-westerly winds rapidly amplifies the PV ridge (compare Figs. 3(a) and (c)) and creates a large negative PV anomaly with values below 0.5 PVU (1 PVU = 10⁻⁶ m² s⁻¹ K kg⁻¹) as far north as 50°N. The abrupt ridge amplification forces equatorward advection of high PV air downstream of the ridge axis, leading to the formation of Trough 1 to the west of Africa (see also Fig. 1(c)).
(b) Trough 2 (5–9 January 2002)

At 00 UTC 5 January, several days before the development of Trough 2 over the North Atlantic, a high-amplitude ridge-trough pattern is observed over the western half of North America (Fig. 1(c)). Three-day precipitation totals of more than 250 mm in the Pacific Northwest between 6 and 8 January suggest a diabatic contribution to the amplification of the wave and the acceleration of the upper-level jet upstream of the ridge (e.g. 73 m s⁻¹ at 330 K off the Oregon coast at 12 UTC 7 January) through latent heating. Convection is triggered ahead of the trough and a weak surface low forms over the US Gulf Coast on 5 January (Fig. 1(c)). It moves across the south-eastern USA during the following days and is located near maritime Canada at 12 UTC 7 January (see its comma cloud in Fig. 4(a)). In the meantime the increasingly SW–NE oriented trough axis propagates eastward and 335-K wind speeds increase by 15 m s⁻¹ in 24 hours to 74 m s⁻¹ (Fig. 4(a)). At that time no extension of the trough over North America into the tropical Pacific is observed. The upper-level jet continues to accelerate and reaches a maximum of 83 m s⁻¹ by 12 UTC 8 January. By 00 UTC 9 January the cloud mass over the North Atlantic has thickened, and an elongated cloud band reaches south-westward into the Tropics (Fig. 4(b)). It is likely that outflow from this band contributed to the maintenance of the jet. Farther to the east a high-amplitude ridge has formed and Trough 2 starts penetrating into the Tropics to the west of Africa. Figure 5 depicts 345-K divergence at 12 UTC 8 January, by which time the East Coast jet had reached its maximum speed. It is evident that the region of the comma cloud coincides with pronounced upper-level divergence. The divergent winds are directed south-eastward towards the eastern flank of the ridge. Large contributions to the convergence over the subtropical Atlantic are made by outflow from tropical convection and by the divergent winds originating in the region of Trough 2 and the STJ streak over Africa.

Figure 6 shows upper-level PV and PV tendencies (Figs. 6(a) and (c)), together with 925-mb Θ distributions and frontogenesis (Fig. 6(b)), for the period of jet and cloud band intensification. Note that a different layer is used for the PV tendency calculations than in Fig. 3 in order to show the level of strongest signals. As before, the positions of surface cyclones are indicated by black dots. At 12 UTC 7 January the area of the jet is clearly marked by strong 335-K PV gradients and a PV ridge is located farther to the east over the North Atlantic (Fig. 6(a)). Rather localized PV destruction is observed in association with the cyclone located near New England, which has moved there from the Gulf Coast over the previous days. Twelve hours later, cold advection is found along the southern and eastern sides of the upper-level trough and strong frontogenesis is observed close to maritime Canada in association with a southerly LLJ over the Atlantic Ocean (Fig. 6(b)). The interaction of the baroclinic zone near the east coast of North America with the cyclonic centre over eastern Canada (Figs. 4(a) and 6(a)) triggers the formation of a second low to the northeast of Newfoundland (Fig. 6(b)). During the following 24 hours the southern low intensifies very little and remains near the baroclinic zone, while the other system moves to the southern tip of Greenland and deepens by 29 mb to 958 mb (Fig. 6(c)). By that time the continuous diabatic PV destruction associated with the two systems has amplified the PV ridge, which supports the generation of Trough 2 to the west of Africa.

(c) Summary and discussion

Figure 7(a) shows a schematic overview over the large-scale circulation associated with the formation of Trough 1. The evolution began with a high-amplitude wave over North America and a strong upper-level jet over the western North Atlantic (depicted as a dark grey arrow in
Fig. 7(a)). The intensification of this jet was favoured by momentum fluxes from the Tropics and by convective outflow along the anticyclonic flank of the jet during the previous days. When the trough over North America approached the jet (see the 4-PVU contour depicted as a thick black line), explosive acceleration, widespread upper-level divergence and strong ageostrophic motions were observed. This period was accompanied by low-level fronto- and cyclogenesis (black dots) and the formation of an elongated cloud band (grey shaded area). The cyclone under the left exit region of the jet deepened explosively by 45 mb in one day (Fig. 7(a)). This system fulfils the criteria for the “bomb” classification by Sanders and Gyakum (1980) and ranks above the 99%-precentile of the climatology by Roebber (1984). Following the analysis of Uccellini et al. 1984 the upper-level features suggest a phase of dynamic (or inertial) instability that is caused by the decrease in wavelength of the trough-ridge system over the western North Atlantic, making it increasingly difficult to balance the flow around the ridge crest. The associated strong ageostrophic motions drove widespread upper-level divergence and a LLJ of warm and moist air that supported the rapid surface development. The strong latent heat release within the frontogenetic cloud band associated with the explosively deepening cyclone led to massive divergent outflow (thin black arrows) and diabatic destruction of upper-level PV (hatching) that supported the instability of the jet and the rapid amplification of the ridge over the North Atlantic. No direct precipitation observations are available over this region, but grid point values of up to 100 mm simulated by the model employed by KM and satellite estimates of the same magnitude corroborate the existence of massive latent heat release. The resulting negative PV anomaly forced a downstream equatorward transport of high PV and thereby initiated the penetration of Trough 1 to lower latitudes near West Africa (Fig. 7(a)). To better understand the impacts of the explosively amplifying ridge on the upper-level circulation, it is useful to adopt a Lagrangian perspective. Forced by the diabatically and dynamically driven divergent upper-level flow within the cloud band, air parcels accelerate across the jet axis towards the stratosphere. Assuming PV conservation, the strong gradient in vertical stability leads to compression, generation of anticyclonic relative vorticity, sharp turning back into the troposphere and deceleration in the eastern part of the quickly north-eastward spreading PV ridge. Farther to the east, stratospheric parcels are forced to flow southward down the sloping isentropic surfaces into the less stable troposphere, resulting in column stretching, convergence and sharp cyclonic turning around the positive PV anomaly near Africa. The stratospheric intrusion is evident as a conspicuous dry line behind the trough in the WV image for 0430 UTC 5 January (Fig. 8(a)).

The large-scale evolution leading to the generation of Trough 2 also involved a high-amplitude wave over North America and a strong upper-level jet farther to the east (Fig. 7(b)). The conspicuous ridge and its associated jet were supported by strong latent heating within the upstream cloud band. Again the approach of the trough to the jet over the western North Atlantic resulted in a phase of jet acceleration, formation of a LLJ, frontogenesis, “bomb”-like cyclone deepening, cloud band formation, divergent outflow and diabatic PV destruction (in agreement with satellite precipitation estimates). Finally a PV ridge formed over the North Atlantic followed by a PV trough close to West Africa (Fig. 7(b)). In comparison to Trough 1 there is less evidence of an acceleration of the jet over the western North Atlantic through convective outflow and momentum transports from the (sub-) Tropics, suggesting an important role for advection from the strong upstream jet, and the downstream ridging is somewhat weaker (Fig. 7). Consistent with the less dramatic development over the North Atlantic, signs of a stratospheric intrusion are also less distinct (Fig. 8(b)). There is some evidence that the evolution depicted in Fig. 7(a) had an impact on the subsequent development. Figures 6(a) and (c) show that the AWB
that took place after the generation of Trough 1 produced a large negative PV anomaly that slowly moved from northern to central Europe between 7 and 9 January (see Fig. 7(b)). It is deemed likely that it acted as a block to the westerly flow and helped to deflect Trough 2 towards the Tropics.

For both Troughs 1 and 2, the combination of the ageostrophic motions in the curved jet exit region with strong outflow from both the extratropical cloud band and from tropical convection led to massive upper-level convergence over the subtropical Atlantic (thin black arrows in Fig. 7). The weak PV gradients in the eastern portion of the ridge allowed tropical outflow to penetrate far into the subtropics. The widespread sinking motions underneath the upper-level convergence led to drying (see the dark regions near 20°N to the south-west of the troughs in Fig. 8). The lower-level divergence underneath the subsidence initiated an enhancement of the trade winds over the eastern North Atlantic (thick grey arrows in Fig. 7). The first trade surge was instrumental in the export of tropical moisture (KM) and triggered convection over the equatorial Atlantic whose outflow accelerated the downstream STJ several days later (Fig. 7(b)). The enhanced outflow also increased convergence in the subtropics (compare Figs. 2 and 5), which in turn intensified subsidence, drying (compare Figs. 8(a) and (b)) and trade winds. Finally the resulting locally enhanced Hadley overturning led to a strong STJ acceleration, the formation of a TP and an extension of rainfalls from the deep Tropics into West Africa (KM).

4. SYNOPTIC EVOLUTION OF CASE II (MARCH/APRIL 2002)

As in case I, a succession of two upper-level troughs is responsible for the generation of extreme precipitation over Africa during case II (see Introduction). These are analyzed separately in the first two subsections, followed by a short summary and discussion of the entire evolution.

(a) Trough 1 (16–29 March 2002)

On 16 March (two weeks before the heavy precipitation event in north-western Africa) an extraordinarily large TP forms over the tropical East Pacific and is associated with strong ageostrophic motions and jet acceleration over North America during the following days (not shown). When the trough of a high-amplitude wave over the North Pacific/western North America approaches the slowly eastward moving jet from the west, wind speeds at 335 K rapidly accelerate to almost 90 m s\(^{-1}\) at 12 UTC 19 March, accompanied by ageostrophic winds near the jet maximum of up to 37 m s\(^{-1}\) (Fig. 9(a)). Along the anticyclonic flank of the jet a cloud band stretches from northern Mexico to the US East Coast (Fig. 9(a)) that produces heavy precipitation (up to 220 mm) over Texas, Arkansas and Oklahoma on 19 and 20 March. At 00 UTC 21 March the western part of the cloud band is located over the south-eastern USA, while the eastern part has moved out to the North Atlantic and started to curve cyclonically into maritime Canada (Fig. 9(b)). The former is characterized by strong upper-level divergent outflow that maintains the jet maximum of just above 70 m s\(^{-1}\) to its northeast. In the exit region of the jet over the North Atlantic, where the fast upper-level flow is steered around a large cyclonic centre (Fig. 9(a)), a comma cloud is evident at 00 UTC 21 March near 35°W (Fig. 9(b)).

At the same time the upper-level jet is marked by a distinct PV gradient at 335 K over the eastern USA, and a weak cyclone, which has formed along the associated baroclinic zone 12 hours earlier, is found off the US East Coast (Fig. 10(a)). The two cloud features over North America seen in Fig. 9(b) generate rather localized regions of moderate diabatic PV destruction (Fig. 10(a)). The comma cloud over the North Atlantic (see Fig. 9(b)) belongs to a second
cyclone, which formed ahead of the PV trough on 20 March and deepened by 14 mb during the 12 hours preceding Fig. 10(a). Strong localized diabatic PV destruction is associated with this feature. One day later, another surface low (core pressure 1000 hPa) has formed along the baroclinic zone over the eastern USA (Fig. 10(b)). The strong north-westerly cold advection underneath the upper-level jet promotes lower-tropospheric frontogenesis, but the air is too dry to produce significant clouds and precipitation. The low farther to the east has moved out to the Atlantic Ocean, intensified moderately and shows weak frontogenesis on its eastern side in association with a south-westerly flow of warm and moist air (Fig. 10(b)). The cyclone over the central North Atlantic has deepened by another 20 mb to 964 mb and is characterized by a large cyclonic circulation, a thick spiralling cloud band and moderate frontogenesis on its northern side. By 00 UTC 23 March the cyclone over the western Atlantic has intensified only little, but the system to the northwest has deepened by 21 mb and moved into eastern Canada in connection with the deepening upper-level PV trough (Fig. 10(c)). Yet the associated cloud evolution and diabatic PV tendencies are relatively weak, probably due to the dry environment. Consistent with that, the amplification of the downstream ridge is at best moderate. In contrast, the strong development of the Atlantic cyclone, which has started to fill by 23 March, produced a large negative PV anomaly over Europe and an upstream positive PV anomaly over the subtropical Atlantic (Figs. 10(c)). The latter supports the penetration of Trough I to lower latitudes, which eventually cuts off from the stratospheric reservoir of high PV to the north (Figs. 12(a)) and generates a closed circulation at lower-levels (Figs. 12(b)). By 29 March the PV cut-off moves into the western Mediterranean Sea and decays (Fig. 12(c)).

(b) Trough 2 (25–29 March 2002)

On 25 March a zonally oriented polar jet streak and a low-level baroclinic zone stretch across the central and eastern USA, while a STJ streak, associated with convection over the eastern tropical Pacific, moves across the southern USA (not shown). When a weak upper-level trough slowly approaches from the west, a conspicuous cloud band forms over the eastern USA and moves to the western North Atlantic by 12 UTC 27 March (Fig. 11). Wind speeds at 335 K along the cloud band barely exceed 50 m s\(^{-1}\). At the same time, the positively tilted trough of a high-amplitude wave over the North Pacific triggered convection to the southwest of Baja California (Fig. 11). Northward momentum fluxes and convective outflow generate another strong STJ streak over the southern USA. Figure 12 shows upper-level PV distributions/ tendencies and 925-mb frontogenesis/\(\Theta\) for the period, when the cloud band seen in Fig. 11 moved out onto the North Atlantic. In accordance with the moderate upper-level jet, PV gradients are relatively weak over eastern North America at 00 UTC on 27 March (Fig. 12(a)). The cloud band, however, produces widespread and strong upper-level PV destruction that has already led to the formation of a weak downstream ridge. To the west of the cloud band a shallow cyclone is observed that has formed over Texas on 25 March in connection with the first STJ streak mentioned above. Lacking a strong upper-level PV anomaly to lock onto, the low has not deepened at all since its genesis. As shown by Fig. 12(b), the cloud band is coincident with strong low-level frontogenesis in connection with the distinct southerly flow on the western side of a huge anticyclone over the North Atlantic. At 00 UTC 29 March a deep PV trough is observed to the east of the Antilles (Fig. 12(c)), downstream of the second STJ mentioned above (Fig. 11). Convection ahead of this trough and the frontogenetic zone farther north combine to form a band of very strong diabatic PV destruction, reaching from near Puerto Rico to the North Atlantic at about 55°N. The resulting massive upper-level PV ridge over the North Atlantic is associated
with the strong low-level anticyclone that supports the frontogenesis as shown in Fig. 12(b). Downstream of the rapidly amplifying ridge Trough 2 penetrates into the subtropics.

(c) Summary and discussion

The development that led to the generation of Trough 1 involved a large-scale wave over North America and a strong upper-level jet farther east over the western North Atlantic (Fig. 13(a)). Two surface cyclones formed over the eastern USA in the entrance region of the jet, propagated north-eastward and intensified (the one moving into Canada characterized by “bomb”-like deepening rates; Fig. 13(a)). Both systems are associated with distinct cloud bands, low-level frontogenesis and upper-level divergence. This is in general agreement with the developments during case I (see section 3c). Due to the dry air involved in the cyclogenesis over Canada, however, the associated diabatic destruction of upper-level PV is only moderate and consequently the formation of a PV ridge over the North Atlantic is less distinct (Fig. 13(a)). A fundamental difference between case II and case I is the substantial disturbance of the large-scale circulation caused by an extraordinarily strong TP event over the tropical eastern Pacific several days before the situation shown in Fig. 13(a). Not only is the strong jet associated with this TP involved in the cyclogenesis over North America (see above), it also supports the formation of a third low in its left exit region over the subtropical Atlantic (Fig. 13(a)). With a 24-hour deepening rate of 25 mb this system fulfils the definition of a “bomb”. According to Sanders and Gyakum (1980) and Roebber (1984) both the time of the year and the location are unusual for such a development that reveals some similarities to a Pacific case described by Martin and Otkin (2004). The diabatic PV destruction associated with this system, together with the advection of low PV on its eastern side, built up a large negative PV anomaly over the eastern North Atlantic (Fig. 13(a)). Advection of high PV on its western side supports the penetration of Trough 1 towards lower latitudes, which finally forms a cut-off with a persistent cyclonic circulation near north-western Africa. The far-reaching impact of the TP over the Pacific on the large-scale circulation is also evident in the cut-off of high PV over the Caribbean Sea (Fig. 10(a)) that has been produced by the breaking of the subtropical ridge that accompanied the TP in the fashion described by Postel and Hitchman (1999).

The generation of Trough 2 began with the formation of a shallow north-eastward moving surface cyclone, resulting from an interaction of a baroclinic zone/upper-level jet across the eastern USA with a weak extratropical trough. The strong frontogenesis associated with this system is likely to have profited from the secondary circulation in the exit region of a STJ streak over the southern USA (c.f. the wintertime cases described by Uccellini and Kocin 1987). The generation of the baroclinic zone was supported by widespread cold advection to the west of the cyclonically curving deep low over eastern Canada (referred to as LC2 by Thorncroft et al. 1993; see Fig. 13(a)) during the previous days (not shown). At late stages the frontogenetic cloud band moved to the western North Atlantic and was extended into the subtropics by convection ahead of a deep PV trough to the east of the Antilles (Fig. 13(b)). Knippertz (2004) shows that the formation of this trough is related to the divergent outflow from an upstream cloud band into the STJ over the southern USA (Fig. 13(b)) and the convergence induced by tropical outflow farther south. The prolonged strong diabatic PV destruction within the elongated cloud band resulted in the generation of a large PV ridge over the North Atlantic, a strong low-level anticyclone feeding moist air into the cloud band on its western side and finally a penetration of the positively tilted Trough 2 to lower latitudes near northwest Africa (Fig. 13(b)). Fundamental differences in this evolution compared to that of case I are the enhancement of baroclinicity due to the cyclone over
eastern Canada, the interactions with the STJ streaks over the southern USA and the weak cyclogenetic, but strongly frontogenetic development over North America.

In analogy to case I, divergent outflow from the extratropical cloud bands and from tropical convection caused upper-level convergence in the subtropics associated with enhanced subsidence and low-level trade winds (Fig. 13). For Trough I the ensuing response of tropical convection led to the formation of a weak STJ streak and an associated thin cloud band over Africa (not shown). Again the feedback between the trade winds and the tropical convection resulted in the generation of a STJ, an accompanying TP and heavy rainfalls over north-western Africa as described by KM. A detailed analysis of the processes in the subtropics leading to the TP formation during this case can be found in Knippertz (2004).

5. SYNOPSIS EVOLUTION OF CASE III (OCTOBER 2003)

In contrast to the other two cases, only one significant upper-level trough is involved in the generation of the heavy precipitation over West Africa during case III. The evolution of this feature is described in part (a) of this section. As mentioned by KM, an important contribution to the rainfall generation is made by an intensification of the trade wind flow from the southern hemisphere into West Africa and the adjacent tropical Atlantic. Therefore an analysis of the large-scale evolution over the South Atlantic and its connection to the trade wind intensification is presented in part (b). As for the other cases, part (c) contains a summary and discussion.

(a) Northern hemisphere evolution (16–21 October 2003)

Several days before the generation of the deep trough near West Africa, prolonged diabatic PV destruction within an elongated cloud band near the west coast of North America supports the formation of a high-amplitude ridge, a strong upper-level jet around its crest and a subsequent AWB (not shown). The extreme latent heat release is illustrated by three-day precipitation totals of up to 318 mm in British Columbia and the Pacific Northwest. By 12 UTC 18 October the ridge and the associated elongated jet of up to 70 m s\(^{-1}\) span across the entire North American continent (Fig. 14(a)). The sharp downstream trough induces convection over the Caribbean Sea and to the east of Florida, as well as a thick cloud feature over the western North Atlantic. A 12-hour increase in 335-K wind speed of 13 m s\(^{-1}\) to 72 m s\(^{-1}\) by 12 UTC 18 October is observed to the east of the trough (Fig. 14(a)), associated with ageostrophic winds of up to 26 m s\(^{-1}\) and strong upper-level divergence. During the next two days the cloud band further elongates and propagates eastward, accompanied by an increasingly SW–NE oriented compact jet streak that reaches 75 m s\(^{-1}\) at 12 UTC 20 October (Fig. 14(b)). A large ridge forms over the North Atlantic, downstream of the cloud band and jet, and a deep positively tilted trough penetrates to low latitudes near Africa (Fig. 14(b)). Ageostrophic winds reach 38 m s\(^{-1}\) in the exit region of the jet at 00 UTC 21 October. In Fig. 14 the cloud feature of tropical storm Nicholas is evident near 18°N, 47°W, which moves very slowly north-westward between 16 and 21 October.

Figure 15 shows upper-level PV distributions/tendencies and 925-mb Θ/frontogenesis for the period, when the cloud band moves out to the North Atlantic. At 12 UTC 19 October relatively strong PV gradients are found in connection with the jets along the ridge crest over central Canada and close to the east coast (Fig. 15(a)). A weak surface cyclone is observed at the eastern side of the PV trough that formed as a frontal disturbance near New England 24 hours earlier. The associated cloud band (see Fig. 14) generates strong diabatic destruction of upper-level PV, which has already contributed to the production a large downstream PV ridge (Fig. 15(a)).
00 UTC 20 October the low has moved underneath the left exit region of the jet and has deepened by 9 mb. A LLJ of moist and warm air over the Atlantic is associated with moderate frontogenesis to the north and east of the low (Fig. 15(b)). One day later, the cyclone’s core pressure has dropped by another 12 mb while its centre is moving along the Labrador coast (Fig. 15(c)). The cloud band over the North Atlantic has detached itself from the cyclone centre and stretches from near the Antilles to the crest of the amplified PV ridge near the southern tip of Greenland, where very strong diabatic PV destruction is observed (Fig. 15(c)). PV values of below 0.5 PVU are found north of 60°N, while a positive PV anomaly is located near the African Tropics (Fig. 15(c)). It is likely that low-PV outflow from Nicholas (see its low-level circulation in Fig. 15(b)) has contributed to the generation of the PV ridge over the North Atlantic.

(b) Southern hemisphere evolution (16–19 October 2003)

At 00 UTC 16 October a large anticyclone with a jet maximum of almost 80 m s\(^{-1}\) to the west of the ridge crest is observed over South America (Fig. 16). Downstream of the anticyclone a trough stretches into the Tropics, followed by a thick cloud band and another jet streak of more than 70 m s\(^{-1}\) (Fig. 16). It appears that the cyclonic centre over tropical South America and the scattered convection to its north and east feed momentum into the eastern jet streak. Figure 17(a) shows that by 00 UTC 17 October a shallow surface cyclone has formed underneath the sharp upper-level PV anomaly associated with the trough over the South Atlantic. At this time upper-level PV destruction within the cloud band is relatively weak. One day later, the low has intensified only little (Fig. 17(b)). Low-level baroclinicity is relatively weak in the cyclogenetic region over the ocean consistent with the feeble frontogenesis. This suggests that convective processes, triggered by the reduction of vertical stability and vorticity advection ahead of the PV anomaly, are responsible for the cloud band generation, as typical for maritime subtropical cyclones (Ramage 1962). The distinct north-easterly flow along the western flank of the large subtropical anticyclone over the eastern South Atlantic is likely to provide air with sufficiently high equivalent-potential temperatures to sustain the convection (Fig. 17(b)). By 12 UTC 19 October the surface cyclone has continued to move south-eastward under the upper-level PV trough and has deepened to 998 mb (Fig. 17(c)). The associated cloud band reveals substantial diabatic PV destruction now and the downstream PV ridge has markedly amplified, which is likely to strengthen the anticyclone shown in Fig. 17(b). In the meantime, a large cloud band has formed ahead of the downstream trough over southern Africa with strong convection over its continental portion (see the associated diabatic PV tendencies in Fig. 17(c)).

(c) Summary and discussion

Figure 18 shows the large-scale circulation over the North and South Atlantic about one day before the heavy rainfalls over West Africa started. The development in the northern hemisphere involved a strong jet over the western North Atlantic ahead of a pronounced upper-level PV trough (Fig. 18). A cyclone formed along the baroclinic zone ahead of this trough and quickly deepened accompanied by a LLJ, frontogenesis, cloud band formation, diabatic PV destruction (consistent with one-day precipitation totals in eastern Canada of nearly 50 mm) and an ensuing amplification of the downstream PV ridge and trough over the North Atlantic. The upper-level forcing through the southern part of the trough triggered convection and thereby extended the cloud band into the subtropics (Fig. 18). As in case II, the baroclinicity had been previously enhanced through strong south-eastward cold advection on the western side of a large cyclonic centre/PV trough over eastern Canada (see Figs. 14(a) and 15(a)), which originated from a prior
explosive LC2 development near New England (not shown). Two factors seem to have contributed to the acceleration of the upper-level jet: Momentum advection from the upstream jet around the crest of a high-amplitude ridge over western North America (Fig. 18) and divergent outflow from the cloud band. In the eastern portion of the ridge over the North Atlantic upper-level convergence led to a moderate enhancement of trade winds and tropical convection after the time presented in Fig. 18. Most of the described features reveal similarities to the other two cases. Due to the lack of an antecedent system as in the prior two cases, tropical contributions to the convergence are mainly related to the outflow from tropical storm Nicholas.

In the southern hemisphere a high-amplitude PV wave with strong jets along the crests of the two ridges, and distinct downstream troughs and associated convective cloud bands was observed over the South Atlantic (Fig. 18). A subtropical cyclone formed underneath the western trough and intensified moderately, while the associated convective cloud band revealed very pronounced PV destruction and thereby contributed to the amplification of the downstream ridge and the associated low-level anticyclone. Dynamic forcing and divergent outflow from this cloud band and the one further to the east over southern Africa resulted in pronounced upper-level convergence in the eastern portion of the ridge over the south-west African coast and an intensification of the trade winds over the eastern South Atlantic. Supported by a Hadley-type circulation in equatorial regions (see the upper-level divergent winds and low-level trade winds near southern West Africa in Fig. 18) this led to a stronger inflow into the West African monsoon and the maritime ITCZ over the northern hemisphere tropical Atlantic (Fig. 18). When the trough from the northern hemisphere finally approached West Africa, an explosive intensification of convection and the STJ was observed (see KM).

6. CONCLUSIONS

This study has examined the large-scale evolutions that led to the equatorward penetration of extratropical upper-level troughs in three cases of heavy rainfalls and TP formation over West Africa described in a companion paper by KM. Since the evolution of two cases (I and II) involved initiation of moisture transport from the Tropics through a previous (dry) trough, these were included in the investigation. For case III the analysis additionally covered the development that led to an intensification of the trade and monsoon flow from the southern hemisphere into the precipitation region. ECMWF analyses, satellite images and surface observations were used for the investigation. The most important findings are:

- The upper-level troughs examined took the form of positively tilted, elongated PV streamers that formed through an equatorward transport of high-PV downstream of a high-amplitude ridge. Dry lines in WV imagery to the west of the troughs indicate stratospheric intrusions. In some cases AWB produced a cut-off of low PV over Europe/the eastern North Atlantic and/or a cut-off of high PV air near Africa. These characteristics are consistent with the LC1 life cycle of baroclinic waves described by Thornicroft et al. (1993). During case II the equatorward advection of high PV was supported by an explosive subtropical cyclogenesis in the exit region of a strong upper-level jet over North America.

- The rapid amplification of the ridges (the AWB) was achieved through south-westerly advection of low PV and diabatic reduction of upper-level PV due to latent heating within an upstream cloud band in agreement with work by Stoelinga (1996) and Massacand et al. (2001). The strong latent heating was connected to explosive cyclogenesis and/or strong frontogenesis near the east coast of North America, where rapid deepening is promoted by the strong
land–sea temperature contrast during the winter half year (Sanders and Gyakum 1980; Roebber 1984), and in some cases to convection in the subtropics.

Factors contributing to the rapid baroclinic developments were (1) pronounced low-level baroclinicity, (2) an upper-level trough/positive PV anomaly approaching from the west and (3) a strong upper-level jet (c.f. Uccellini 1999). In some cases signs of low dynamic stability (explosive jet acceleration, strong ageostrophic motions, widespread upper-level divergence) were observed. This was related to diabatically driven divergent outflow near the tropopause and to unbalanced flow passing through a trough-ridge pattern with decreasing wavelength (Uccellini et al. 1984; Cammas et al. 1994). As part of the three-dimensional secondary circulation associated with the accelerating curved jet, a southerly LLJ over the western North Atlantic fed warm, moist air into the cyclo- and frontogenetic region, probably associated with strong sensible and latent heat fluxes (Uccellini et al. 1984). During case II frontogenesis was enhanced by the superposition of the left exit region of the STJ and the right entrance region of the polar jet (c.f. Uccellini and Kocin 1987).

The development over North America was supported by diabatically enforced high-amplitude ridging (AWB) near the west coast that was associated with positive momentum advection into the downstream jet and the formation/amplification of the downstream trough.

Positively tilted subtropical troughs and convection/TP formation along their eastern flanks indicate an acceleration of the extratropical jets through poleward momentum fluxes and divergent outflow near the tropopause (c.f. Hurrell and Vincent 1990).

The developments in the extratropics were supported by previous modifications to the large-scale circulation in two ways: (1) Cyclonically breaking baroclinic developments (referred to as LC2 by Thorncroft et al. 1993) acted to enhance low-level cold advection over North America. (2) The isolation of a low-PV region over Europe/eastern North Atlantic through AWB led to a blocking situation that deflected the following trough to the south.

To the west of the troughs penetrating to low latitudes near Africa strong upper-level convergence was generated through the highly ageostrophic winds in the exit region of the strong jet and outflow from both the upstream extratropical cloud band and tropical convection. The resulting sinking motion and low-level divergence enhanced the trade winds and triggered tropical convection, which in turn generated more convective outflow, leading to a positive feedback on the local Hadley overturning (in particular in the case of repeated extratropical forcing by two troughs), an acceleration of the downstream STJ and the formation of a TP (see Knippertz 2004). The trade surge circulation is also instrumental for the moisture export from the Tropics (KM). During case III baroclinic/convective activity related to a high-amplitude PV wave and strong jets over the South Atlantic fuelled strong trade flow into West Africa.

With respect to seasonal variations the present study reveals largest low-level baroclinicity, upper-level jet speeds and cyclone deepening rates during the January case, while the transition season cases show stronger influences by convective processes. The described mechanism is geographically fixed due to the distributions of orography, land and sea, and sea surface temperatures, leading e.g. to the general tendency for ridging close to the west coast of North America and the cool season quasi-permanent baroclinic zone along the east coast.

The present results reveal some interesting agreement with statistical investigations relating tropical convection/TP formation to the penetration of extratropical Rossby waves into the westerly duct regions in the tropical eastern Pacific and Atlantic (Webster and Holton 1982; Hartmann and Liebmann 1984; Kiladis and Weickmann 1992, 1997; Iskenderian 1995; Kiladis 1998). Despite the smoothing introduced through filtering, averaging or compositing these
studies equally find an influence of strong upstream jets and ridges, trade surges and upper-level convergence between outflow from tropical convection and an extratropical cloud band. It appears, however, debatable, whether the waves can propagate into the Tropics due to the existence of low-latitude westerlies (according to a linear wave dispersion theory) or whether the diabatic and highly nonlinear interaction processes described here force the equatorward penetration of troughs, whose circulations produce the low-latitude westerlies.

Future work should investigate more cases of deep equatorward penetrations of extratropical troughs over the Atlantic and Pacific Oceans in order to better understand and to compare the processes affecting the two main westerly duct regions. An interesting extension to purely observational studies would be to test the net effect of latent heating on the large-scale evolution by suppressing it in a model simulation as in Stoelinga (1996) and Massacand et al. (2001). In particular the evolutions of cases I and II suggest that one substantial disturbance of the circulation (e.g. through a massive TP/STJ or a highly unstable jet/explosive cyclogenesis) can initiate a chain of extraordinary developments (heavy precipitation, strong jets, explosive cyclogenesis etc.) that outlasts single synoptic events. This seems consistent with the recent observation of a Rossby wave-based coherency to periods of prevalent natural catastrophes like flooding and storms (M. A. Shapiro 2003, personal communication). It would be worth investigating whether such periods are characterized by a general tendency for enhanced tropical–extratropical interactions, strong meridional flow, large amplitude waves, elongated SW–NE cloud bands and STJ streaks as during the present cases. The existence of active periods that last longer than individual systems was recently demonstrated by Otkin and Martin (2004) with respect to subtropical cyclogenesis. In addition, the analyzed cases give evidence that the generation of deep troughs over the North Atlantic is influenced by TPs, meridional momentum fluxes and upper-level jets over the North Pacific and North America, which are sensitive to the El Niño/Southern Oscillation (ENSO; McGuirk et al. 1987; Matthews and Kiladis 1999; Shapiro et al. 2001). This result could be used to further examine the observed linkages of ENSO to West African precipitation (e.g. Nicholson and Kim 1997; Knippertz et al. 2003).

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FIGURE CAPTIONS

Figure 1. GOES IR satellite images with superimposed 335-K isotachs in white (units are in m s\(^{-1}\)) and streamlines in black for (a) 12 UTC 2 January, (b) 00 UTC 4 January and (c) 00 UTC 5 January 2002 (case I).

Figure 2. Isotachs (contours are 30, 50, 60, 70 and 80 m s\(^{-1}\)), divergent wind (vectors according to scale in upper left corner; only winds greater than 5 m s\(^{-1}\) are plotted) and divergence (shading/hatching according to legend; units are in \(10^{-6}\) s\(^{-1}\)) at 345 K for 00 UTC 5 January 2002 (case I).

Figure 3. (a) PV (contours are 0.5, 1, 2, 3, 4, 6 and 8 PVU with the 1-, 4- and 8-PVU contours printed in bold) and total wind (vectors according to scale in upper right corner; only winds greater than 15 m s\(^{-1}\) are plotted) at 335 K, and diabatic PV tendency in the 150–200 mb layer (shading according to legend; units are in PVU day\(^{-1}\)) for 12 UTC 4 January 2002 (case I). (b) 925 mb potential temperature (isopleths every 5 K with the 260-, 280- and 300-K contours printed in bold), total wind (vectors according to scale in upper right corner; only winds greater than 5 m s\(^{-1}\) are plotted) and frontogenesis (shading according to legend; units are in K (100 km\(^{-1}\)) day\(^{-1}\)) for 00 UTC 5 January. (c) As (a) but for 12 UTC 5 January 2002. The position and core pressure (in mb) of selected low-pressure systems are indicated by black dots.

Figure 4. As Fig. 1 but for (a) 12 UTC 7 January and (b) 00 UTC 9 January (case I).

Figure 5. As Fig. 2 but for 12 UTC 8 January 2002 (case I).

Figure 6. As Fig. 3 but for (a) 12 UTC 7 January (b) 00 UTC 8 January and (b) 00 UTC 9 January 2002 (case I). The diabatic PV tendencies in (a) and (c) are calculated for the 200–250 mb layer.

Figure 7. Schematic overview over case I, showing (a) 00 UTC 5 January and (b) 00 UTC 9 January 2002. Displayed are the major cloud features (grey shading), the jet maxima (dark grey arrows) and 1- and 4-PVU contours at 335 K (thick black lines), divergent-wind streamlines at 345 K (thin black arrows), upper-level PV destruction (hatching), low-level trade surges (thick light grey arrows) and the 12-hourly position and core pressures (in mb) of selected low pressure systems. Bold Ls mark the positions that correspond to the time of the respective figure.

Figure 8. Meteosat WV satellite image for (a) 0430 UTC 5 January and (b) 0030 UTC 9 January 2002 (case I).

Figure 9. As Fig. 1 but for (a) 12 UTC 19 March and (b) 00 UTC 21 March 2002 (case II).

Figure 10. As Fig. 3 but for (a) 00 UTC 21 March, (b) 00 UTC 22 March and (c) 00 UTC 23 March 2002 (case II). The diabatic PV tendencies in (a) and (c) are calculated for the 200–250 mb layer.

Figure 11. As Fig. 1 but for 12 UTC 27 March 2002 (case II).
Figure 12. As Fig. 3 but for (a) 00 UTC 27 March, (b) 00 UTC 28 March and (c) 00 UTC 29 March 2002 (case II).

Figure 13. As Fig. 7 but for (a) 00 UTC 23 March and (b) 29 March 2002 (case II).

Figure 14. As Fig. 1 but for (a) 12 UTC 18 October and (b) 12 UTC 20 October 2003 (case III).

Figure 15. As Fig. 3 but for (a) 12 UTC 19 October, (b) 00 UTC 20 October and (c) 00 UTC 21 October 2003 (case III).

Figure 16. As Fig. 1 but for 00 UTC 16 October 2003 (case III).

Figure 17. As Fig. 3 but for (a) 00 UTC 17 October, (b) 00 UTC 18 October and (c) 12 UTC 19 October 2003 (case III). The diabatic PV tendencies in (a) and (c) are calculated for the 200–250 mb layer.

Figure 18. As Fig. 7 but for 12 UTC 19 October 2003 (case III) displaying both northern and southern hemisphere evolutions.
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Figure 16. As Fig. 1 but for 00 UTC 16 October 2003 (case III).
Figure 17. As Fig. 3 but for (a) 00 UTC 17 October, (b) 00 UTC 18 October and (c) 12 UTC 19 October 2003 (case III). The diabatic PV tendencies in (a) and (c) are calculated for the 200–250 mb layer.
Figure 18. As Fig. 7 but for 12 UTC 19 October 2003 (case III) displaying both northern and southern hemisphere evolutions.