A diagnostic study of the dynamics of the northern hemisphere winter of 1985–86

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SUMMARY

February 1986 was notable for a dramatic blocking pattern in the eastern N Atlantic with part of NW Europe having its coldest February in 300 years. However, the previous December and January had been mild in the same region and the hemispheric anomaly field was characterized by a pattern which could be viewed as a wave-train emanating from the tropical western Pacific. Such an extreme change within a single winter season is unusual and must be understood if reliable monthly and seasonal forecasts are ever to be produced.

The tropospheric flow field, isentropic distributions of potential vorticity, and diabatic heating calculated as a residual in the thermodynamic equation are used to illustrate the similarities and differences in the two parts of the winter and to suggest possible reasons for these differences. The mean vorticity equation is also used to investigate the importance of the change in the zonal flow, the transients and the mean divergence. It is concluded that the change in the fluxes due to the synoptic weather systems was crucial, but that a catalyst for the February block could have been provided by an unusual diabatic forcing in the S American–Caribbean region. Implications are drawn for the ingredients which must be included in dynamical models for predicting on the monthly to seasonal time-scales.

1. INTRODUCTION

The first stream of the World Climate Research Programme is concerned with the behaviour of the atmosphere on the time-scale of one month to a season, and in recent years there has been a surge of interest in and research into the possibility of producing dynamically based forecasts for this time-scale (e.g. Gilchrist 1986). Such a possibility can exist only if the length of the period and stability of the regime of flow is such that the initial conditions of the atmosphere contain crucial information or if slowly varying or predictable boundary conditions are important. By running general circulation models (GCMs) with modified initial conditions a general consensus has been reached that, beyond the month time-scale, the details of the initial state are not important. The hope for seasonal forecasting would seem to rest on the atmospheric boundary conditions: the state of the earth's surface, including soil moisture and vegetation (e.g. Mintz 1984; Rowntree and Sangster 1986), snow cover (Namias 1985b), and the state of the upper layer of the ocean including sea ice cover (e.g. Mitchell and Hills 1986) and sea surface temperatures (s.s.t.s). The last item has received particular attention over many years. Namias (1985a) has stressed the importance of the North Pacific for long-range forecasting in N America. Ratcliffe and Murray (1970) have suggested similar sensitivity in Europe to s.s.t.s in areas of the N Atlantic. Researchers using GCMs have generally had less success in confirming the reality of such relationships, though the results of Palmer and Sun (1985) supported the Ratcliffe and Murray suggestions. More unanimity has been shown for the importance of tropical s.s.t. anomalies. Since the atmosphere is driven primarily by the release of latent heat in tropical convection which tends to occur over the warmest waters (as well as over the continents), and also given that large-scale tropical s.s.t. anomalies generally have a time-scale of a season or longer, the connection is scarcely surprising. Most emphasis has been placed on the El Niño–Southern Oscillation phenomenon with observational work by many authors, including Bjerknes (1969) and Horel and Wallace (1981), and modelling studies by Rowntree (1972), Shukla and Wallace (1983), Blackmon et al. (1983), Lau (1985) and many others. However, there has also been considerable interest in the tropical Atlantic (e.g. Rowntree 1976; Hastenrath and Heller 1977; Moura and Shukla 1981).
One of the greatest advances in research capability in the subject in recent years has been the availability of relatively accurate global analyses produced on a routine basis by the advanced forecast centres. Through a joint project with the Meteorological Office, the Department of Meteorology at the University of Reading has had access to the archived data of the European Centre for Medium Range Weather Forecasts (ECMWF). This has permitted almost real time diagnosis of the behaviour of the global atmosphere over periods of half months to seasons (e.g. White 1982, 1983; Sardeshmukh 1984). The data are available four times a day as initialized or uninitialized analyses on a 1.875° grid; however, results are usually obtained only for the 12 GMT initialized analysis on a 5° grid.

The season December 1985 to February 1986 contained numerous interesting features. One of these was the sudden transition in most of Europe from a very mild December and January to an extremely cold February. In England this transition was unparalleled in the last 249 years (Ratcliffe 1986). To illustrate the contrasting patterns of the winter, 700 mb temperature differences from a six-year average for December–February are shown in Fig. 1 for December plus January and for February. The most prominent pattern in the early winter is that of alternating positive and negative anomalies in a wave-train from the subtropical western N Pacific across N America to the subtropical western N Atlantic. In February this pattern is replaced by a zonally elongated dipole in the N Pacific. However, the dominant signature is now in the N Atlantic/European region with anomalies of more than 4 K.

The change in weather regime that occurred over Europe between the end of January and the beginning of February proved difficult to capture in medium-range weather forecasts. For example the skill of ECMWF forecasts was below average for this period until the new regime was essentially present in the initial conditions. The mechanisms for the stability of each regime and the transition between them are crucial for our understanding of the forecast problem on time-scales from a few days up to a season. If such transitions are random events this will provide an upper bound to the skill that can ever be attained.

Figure 1. Anomalies in northern hemisphere 700 mb temperature from a six-year December–February average for December + January (left) and February (right). Contours are drawn at intervals of 2 K. Areas with values greater than 2 K are shaded while those with values less than −2 K are stippled. The figures are bounded by the 20°N latitude circle. The 80° latitude circle and lines of longitude every 60° are indicated.
In this paper, large-scale diagnostics for the northern hemisphere winter 1985–86 are presented with a view to isolating dynamical mechanisms that may be of particular importance for this season. Of particular interest is the change from December and January to February in the European region. Basic diagnostic fields are presented in section 2. In section 3 a simplified vorticity equation is used to diagnose the importance of different processes. Some concluding remarks are made in section 4.

2. Atmospheric Diagnostics

(a) The wind field

The behaviour of the northern hemisphere 250 mb wind field is summarized in Figs. 2 and 3. Shown in Fig. 2 are streamfunction contours and isotachs for the six half-month periods which should be compared with a 1979–84 December–February average given in Fig. 3(a). The Pacific jet fluctuates somewhat in intensity and length, but more variability is seen in the amplitude and location of the anticyclone on its equatorial flank. The ridge over the west coast of N America in December I, intensifies in December II, is centred more over the Rockies in January and is highly variable in February, a point which will be returned to below. The Atlantic jet varies from a little more than 35 m s⁻¹ in January II and February II to more than 55 m s⁻¹ in December II. It extends towards western Europe in December and January consistent with the warmth there, but both halves of February exhibit a blocking signature with split flow in the eastern Atlantic. The centre of the split moves from near 0° in the first half to 30°W in the second half. In the second half along the 20°W longitude line there are three separate westerly wind maxima of more than 25 m s⁻¹.

The flow in the region of the Rocky Mountains varies from a weak north-westerly in the first half of February to a strong westerly in the second half. This variation is so extreme when viewed in the context of the previous four half months that an interaction of the flow with the Rockies appears an unlikely mechanism for the blocked flow downstream. Thus in this case the orographically induced multiple equilibrium theory of Charney and DeVore (1979) and Charney and Strauss (1980) and others would seem to be not relevant.

Figure 3(b) shows the 250 mb streamfunction contours and isotachs for the period December–February 1985–86. Comparing with the six-year average in Fig. 3(a) shows that the departure of the seasonal mean from the 6-year average is small and essentially irrelevant as a measure of the winter's weather. The remaining four panels show the 250 mb streamfunction anomalies from the 6-year average for December + January, February and the first and second halves of February respectively. The two-month period shows an identical extratropical wave-pattern to that of Fig. 1(a), with 700 mb warmth corresponding to 250 mb anticyclonic anomaly, consistent with the simple equivalent barotropic structure of such anomalies stressed by Wallace and Gutzler (1981). The continuation of this pattern into the tropics at upper levels is also evident. Again the changed anomalies in February, in particular the blocking anticyclone poleward of a cyclone, are equivalent barotropic. The pictures from the two halves of February accentuate the changes in the N Pacific and western N American regions but indicate only a shift of the block in the N Atlantic/European sector.

The anomalies shown in Figs. 3(c) and (d) show a very large measure of agreement with those given in the Climate Analysis Center Bulletins (1985, 1986a, b) for the anomaly of monthly average 200 mb streamfunction from a monthly climatology for a sixteen-year base period. As shown in their figures, the December + January picture is the average
Figure 2. The northern hemisphere 250 mb wind field for the six half-month periods from 1 December 1985 to 28 February 1986. (a) to (f) respectively. Shown by heavy contours are isotachs every 10 m s\(^{-1}\) starting at 25 m s\(^{-1}\), and by light contours the streamfunction with contour interval 5 \times 10^6 m^2 s\(^{-1}\). The bounding circle is the equator.
Figure 3. The December–February northern hemisphere wind field for (a) six-year average and (b) 1985–86. The conventions are as in Fig. 2. Shown in (c)–(f) are the 1985–86 250 mb streamfunction anomalies from the six-year average (shown in (a)): (c) December + January; (d) February; (e) first half of February; (f) second half of February. The contour interval is $5 \times 10^9$ m$^2$s$^{-1}$ and negative contours are dashed.
of a pattern that shifted $1/8$ to $1/4$ wavelength downstream from December to January. The movement of the ridge from the west coast of N America to the Rocky Mountain region noted above as occurring in Fig. 2 was part of this signature. The latter month has a cyclonic anomaly in the equatorial Pacific west of the date line and this produces the residual signature seen there in Fig. 3(c). February has an anticyclonic anomaly in the equatorial Pacific similar to that of December, but the rest of the pattern is different.

The zonally averaged westerly winds for December + January and for February are given in Figs. 4(a) and (b) respectively. The picture for December + January is little different from the six-year average. On the other hand, February shows a considerably stronger subtropical westerly jet, a very weak westerly or even easterly flow near 60$^\circ$N and comparatively strong westerlies near 75$^\circ$N. The blocking signature is clearly present in the zonal average.

(b) Isentropic distributions of potential vorticity (PV)

Hoskins et al. (1985) have discussed the PV invertibility principle which in the global domain says that given the total mass between isentropic surfaces then the isentropic distribution of PV (IPV) plus a low-level distribution of temperature is sufficient to give all the information about a balanced flow. Further, in the absence of diabatic processes, PV is advected on isentropic surfaces. These concepts were used to illustrate the dynamics of several synoptic systems including one particular blocking high development. Further discussion has been given by Shutts (1986). The essential ingredient of a blocking high is a large poleward excursion of low PV subtropical air ahead of a slowly moving, meridionally elongated trough. This anomalously low PV air tends to develop its own anticyclonic circulation which then cuts it off from its source region. Given the conservation properties of PV, the only way for this anticyclone to decay is by the low PV air moving back to the subtropics or by diabatic processes. In this case the latter would probably only be radiative and would have a time-scale of a week or more. Hoskins et al. (1983), Shutts (1983), Foreman (1985) and Haines and Marshall (1987) have shown that this large poleward excursion of low PV air is most likely to occur at the end of a storm-track where there is already a tendency for the flow to split. Thus, on average, decaying synoptic systems act to reinforce a blocking anticyclone.

Figure 5(a) shows the PV and flow vectors on the 315 K surface for the December + January flow time-averaged on $p$ surfaces, and Fig. 5(b) is the corresponding picture for
the first half of February. Strictly these figures should be obtained using time-averages on isentropic surfaces, but comparison of a 320 K full February picture with one obtained in the correct manner by N. Butchart (personal communication) shows only small differences. The December + January IPV contours, like the height field on a pressure surface, show rather little diffuence in the eastern Atlantic, consistent with the extension of the storm-track and jet into Europe. However, even in this time-average, Fig. 5(b) shows cut-off low PV in the blocking anticyclone region and high PV corresponding to polar stratospheric air in troughs to the south-east and south-west. The picture for the second half of February (not shown) is again markedly different over the Pacific and western N America, but the Atlantic shows only a westward shift of the low PV cut-off to southern Greenland and a similar shift of the 'troughs' on either side.

It is tempting to interpret these mean pictures in terms of advection by the mean flow and mean diabatic forcing. However, inspection of instantaneous IPV maps for each day of February emphasizes the transient nature of the flow. There are five separate events in the first half of February in which a cut-off of low potential vorticity ahead of a decaying mobile trough renews the blocking system. Instantaneous 315 K IPV maps for the first four of these events are shown in Fig. 6. On 3 February, (a), the first major incursion of low PV air from the subtropical Atlantic is just about to cut-off. Near 80°E is the remains of a low PV incursion that had occurred over western USSR in the last few days of January. By 5 February, (b), the new PV cut-off has broken into blobs from 90°E to 0°E and another PV cut-off is about to occur. This appears to amalgamate with the blob near 0°E and by 8 February, (c), the remnants are over the British Isles and along north-eastern Europe. The next low PV cut-off is also imminent and by 10 February, (d), this forms a ribbon from 90°W to 0°. In the next few days this ribbon appears to thin and dissipate while the PV cut-off occurring near 10°W on 10 February moves east and south-east. Another event (not shown) occurs on 12 February, and there is a minor episode on 14 February. The low PV from them appears to split with one part moving south-east and the other moving westwards over the north of Greenland.
Figure 6. 315 K IPV maps for 3, 5, 8, and 10 February at 12 GMT, (a)–(d) respectively. Contours of potential vorticity are drawn every 0.5 units from 0.5–1.5 units. The region 2–3 units is blackened representing the likely tropopause position. The 'stratospheric' region greater than 3 units is stippled with contours every 2 units. The scale for the wind vector is such that the arrow underneath the figures would represent 100 m s\(^{-1}\). The region shown is bounded by the 20°N latitude circle, 90°W and 90°E.

The next synoptic development occurs much further west, close to the east coast of N America. It is a major one with a surface pressure minimum of less than 960 mb at 00 GMT on 17 February, having deepened about 20 mb in the previous 12 h. The poleward flux of low PV ahead of this was into the region of southern Greenland. Three more events during the second half of February reinforced the block in this region. Colucci (1985) has studied an explosive cyclogenesis event that also appeared to lead to a shift and intensification of a blocking regime.

(c) Diabatic heating

The Climatic Analysis Center Bulletin (1986b) gives a longitude-time plot of outgoing long-wave radiation (o.l.r.) anomalies averaged from 10°N–10°S for the year up to the end of February 1986. This gives a measure of anomalous high cloud and thus an indication of anomalous diabatic heating in the equatorial region. The figure suggests, and a similar velocity potential figure supports, the notion that a convective heating maximum occurred in the Indonesian region during early December. A stronger event moved across the Indian and W Pacific Oceans in January. Another weaker maximum near Indonesia is indicated in the second half of February. There is therefore evidence of significant equatorial activity on the 30–60-day time-scale.

However, it is still of interest to consider averages over periods of one or two months. Monthly figures for o.l.r. and its anomaly as functions of position are given in the CAC Bulletins, but these can be at best only crude indicators of tropical diabatic heating. As will be discussed elsewhere, useful estimates of this heating may be obtained from residual calculations using ECMWF operational analyses. In particular it is of interest here to see if such calculations yield any clues as to the change in regime that occurred during the winter.
The time-averaged thermodynamic energy equation can be written

$$c_p \left[ \nabla \cdot \nabla T + \left( \frac{p}{p_o} \right)^\kappa \frac{\partial T}{\partial p} + \nabla \cdot \mathbf{v} \cdot \mathbf{T} + \left( \frac{p}{p_o} \right)^\kappa \frac{\partial \omega}{\partial p} \omega \right] = \overline{H}$$  \hspace{1cm} (1)

where the bar and prime refer to a time-average and the deviation from it, respectively, $p_o = 1000 \text{ mb}$, $\kappa = R/c_p$, $\overline{H}$ is the time-averaged diabatic heating, and a very small term associated with the change in $T$ over the season has been neglected. The equation will be discussed in terms of its vertical average up to 50 mb. The first and second terms on the left-hand side are referred to as the steady horizontal and vertical terms and the third and fourth terms together as the transient term. Note that a positive sign in these terms implies a tendency to cool the mean flow. This convention is used so that the sum of the steady and transient terms is equal to the diabatic heating.

In December + January, the dominant signature in the transients (Fig. 7(b)) is that of the storm-tracks in the southern hemisphere, N Pacific and N Atlantic. For example, the transients act to warm the mean flow over eastern N America at a rate which would need to be offset by diabatic cooling of more than 300 W m$^{-2}$. Associated with the more than normally extended storm-track, the region of cooling by the transients of more than 150 W m$^{-2}$ extends from the western N Atlantic to Scotland. The total steady terms (Fig. 7(c)) correspond to cooling associated with large-scale ascent in tropical convective regions such as Indonesia. Over eastern N America and the equatorward side of the Atlantic storm-track the steady and transient terms have a large measure of cancellation.

The mean, depth-averaged diabatic heating, which is the sum of the terms shown in Figs. 7(b) and (c), is presented in Fig. 7(a). There is a maximum in the Indonesian region of nearly 250 W m$^{-2}$ with large values extending across the Indian and Pacific Oceans and along the South Pacific Convergence Zone. There are also maxima over S America and southern Africa. There is significant cooling in the subtropics off the western coasts of the continents and over the northern hemisphere continents. In the regions of the N Atlantic and Pacific storm-tracks there are elongated heating maxima of more than 150 W m$^{-2}$, presumably associated with latent heat release and low-level sensible heating.

The corresponding pictures for February are shown in Fig. 8. One of the main changes in the transients term (Fig. 8(b), cf. Fig. 7(b)) is the decrease in extent and intensity of the signature of the N Atlantic storm-track. This change is in accord with comments made above about the relative extents of the storm-track, but also implies that the very significant poleward PV fluxes by the transients discussed in section 2(b) are not accompanied by large heat fluxes, which is consistent with the equivalent barotropic nature of the decaying synoptic systems that perform this PV flux. The steady term (Fig. 8(c)) is broadly similar to that for the previous two months (Fig. 7(c)).

The tropical diabatic heating field of February (Fig. 8(a)) differs somewhat from that for December + January (Fig. 7(a)) but the significance of such differences is not yet known. The monthly o.l.r. pictures indicated below average o.l.r. in the equatorial region from 90°E–180°E, strengthening from December to January. In February the negative o.l.r. in this region was predominantly north of the equator. The study of Palmer and Owen (1986) suggests that increased precipitation in the western tropical Pacific leads to a Pacific–N American anomaly pattern very similar to that of December. However, the shift of this pattern in January and its total change in February are not at present easily related to changes in precipitation patterns in the tropical Pacific. It is possible that the cooling over SE Asia, which is larger in December + January (Fig. 7(a)) than in February (Fig. 8(a)), is also of importance.

Another feature of interest is the northern subtropical region from the eastern Pacific
Figure 7. (a) The vertically averaged diabatic heating for December + January calculated as a residual in the thermodynamic equation (1). The contributions of the transient and steady terms are shown in (b) and (c) respectively. The fields have been smoothed using a 'V' filter having a weighting of 0.1 at n = 24 as described in Sardeshmukh and Hoskins (1984). Contours for positive values are continuous and the regions with values greater than 50 W m\(^{-2}\) have dark stippling. Negative contours are shaded and the regions with values less than -50 W m\(^{-2}\) are lightly stippled.
Figure 8. Vertically averaged diabatic heating, (a), and the transient and steady contributions, (b) and (c) respectively, for February. Conventions as in Fig. 7.
through to the eastern Atlantic. In December + January (Figs. 7(a) and (c)) there is the usual diabatic cooling and compensating sinking motion in the eastern ocean basins. In February (Figs. 8(a) and (c)) the whole band exhibited this signature, with diabatic cooling of more than 100 W m\(^{-2}\) in the Caribbean, and differences between the two periods in this region of the same order. The o.l.r. anomalies provide some confirmation of this change. In December and January anomalies in the region were not very significant, but in February there were positive anomalies at 20°N from the coast of central America to the central N Atlantic. The implication is that in February there was reduced cloudiness and increased radiative cooling in the region. Rainfall figures also show a general change in the Caribbean from average or above average values in December and January to below 50% of normal in February. For example Merida in the Yucatan Peninsula (21°N 90°W) only had 1 mm of rain in February, which is 3% of its normal for the month.

In middle latitudes the dominant change from Fig. 7(a) to Fig. 8(a) is that the heating in the N Atlantic is confined to the west, clearly implying a different forcing of the mean atmospheric flow in February. However, this change should probably be viewed as a signature of the truncated storm-track rather than a cause of the differing weather regimes.

3. Simple model diagnosis

It is often convenient to split the horizontal flow into its rotational and divergent components, \(v_\psi\) and \(v_x\) respectively. It is apparent from Figs. 7 and 8 that in the tropics latent heating in regions of convection is largely balanced by adiabatic cooling associated with ascent. Thus such a region is one of upper-level divergence and low-level convergence, corresponding to \(\nabla \cdot v_x\) positive and negative respectively. Splitting the time-mean flow into its rotational and divergent components, the complete vertical component of the time-mean vorticity equation may be written

\[
\bar{v}_\psi \cdot \nabla \bar{\zeta} = -\nabla \cdot (\bar{v}_x \bar{\zeta}) - \nabla \cdot \bar{v} \cdot \bar{\zeta} + F
\]  

(2)

where the overbar represents a time mean, the prime a deviation from this mean, \(\zeta\) is the absolute vorticity and \(F\) is the sum of the vertical advection, twisting and frictional terms. Here we shall model \(F\) by a simple linear damping plus a small biharmonic diffusion acting on a time-scale of \(\frac{1}{4}\) day at the length-scale of the spherical harmonic of degree \(n = 42\). The scale dependence of the latter is so strong that at \(n = 10\) the timescale is about 67 days.

Separating the mean rotational flow into its zonal-averaged portion, denoted by square brackets, and its zonally asymmetric portion, denoted by asterisks, Eq. (2) may be written

\[
[u] \frac{\partial \bar{\zeta}^*}{\partial x} + \bar{v}_\psi \cdot \nabla ([\bar{\zeta}] + \bar{\zeta}^*) = -\nabla \cdot \bar{v}_x ([\bar{\zeta}] + \bar{\zeta}^*) - \nabla \cdot \bar{v} \cdot \bar{\zeta}^* + F.
\]  

(3)

If the zonally averaged time-mean rotational flow, the time-mean divergent velocity \((\bar{v}_x)\), and the transient term are all specified and \(F\) is modelled as discussed above, then Eq. (3) may be viewed as an equation for the zonally asymmetric time-mean rotational motion \((\bar{v}_\psi^*\) and \(\bar{\zeta}^* = f + \mathbf{k} \cdot \nabla \times \bar{v}_\psi^*\). To solve this equation a term \(\frac{\partial \zeta^*}{\partial t}\) is added to the left-hand side and the equation is integrated forward in time until a steady state is achieved. For the cases described below the linear damping part of \(F\) acts on a time-scale of five days. Experiments using a time-scale of 14.7 days did not lead to a steady solution,
suggesting that the solutions of Eq. (3) with this damping are probably unstable. This is consistent with the results of Simmons et al. (1983). However the time-mean of the solutions with this smaller damping appeared to be close to the steady states achieved with the larger damping. The numerical code used to solve Eq. (3) was a modification of a simple barotropic vorticity equation model employing spherical harmonic expansions truncated at $n = 42$.

The observed zonally asymmetric streamfunctions at 250 mb for December + January and February are shown in Figs. 9(a) and (b) respectively. The corresponding results from the vorticity equation model are shown in Figs. 9(c) and (d). The agreement for December + January ((a) and (c)) is extremely good. The agreement between (b) and (d) is not as good in detail though the major structures are reproduced. Differences from climatology are always relatively subtle when viewed in such pictures, and although the model being used is not an anomaly model the results will be shown in terms of anomalies from a winter climatology. Such pictures for the model are shown in Figs. 9(e) and (f). They should be compared with the observed anomalies which were given in Figs. 3(c) and (d). It is seen that the model reproduces all the anomaly centres in December + January. However, the anomaly field for February is poor except over the sector from 60°W to 30°E, which fortunately includes the blocked region emphasized previously.

There are really only two possible causes for error in the vorticity model: either the data are not totally consistent or the vertical advection and twisting terms are important. It is a very stringent test of the data, and it seems likely that when errors exist they are probably due to errors in the analysed divergence. In fact, in Sardeshmukh and Hoskins (1987) the model has been used to provide an improved estimate of the divergent flow.

As discussed above, there are three input fields for this diagnostic model: the zonal flow, the mean divergent wind and the convergence of the transient vorticity flux. Each of these input fields will differ from its climatological value during both December + January and February and solutions of Eq. (3) would allow us to determine which anomalous input has most association with anomalies in the mean rotational flow in each period. At present the climatological values of these inputs are not sufficiently accurate for this approach to be possible. However, we can consider the relative importance of the change in the inputs from the first period to the second in producing the change in the simulated anomalous streamfunction. Figure 10 shows the anomaly results obtained by using the December + January inputs but changing the fields of either, (a), the zonal flow or, (b), the transient term to their February values. These should be compared with both the observed anomalies (Figs. 3(c) and (d)) and the full simulations (Figs. 9(e) and (f)). Changing the zonal flow does not produce a dramatic effect but clearly degrades both the amplitude and position of the anomalies when viewed as a simulation of December + January. For example, the Pacific anomalies equatorward of 40°N are moved at least 30° eastwards, with the positive anomaly merging with that over the west coast of N America. However, the result is not similar to the full February simulation. Using the February transients (Fig. 10(b)) produces a larger degradation and probably leaves little useful signal as a simulation for December + January. Interestingly, this simulation shows a blocking anomaly signature near 20°W, similar to that observed and simulated for February.

In the absence of good climatological data we now confine ourselves to the change that occurred in the Atlantic sector. That this is simulated well is confirmed in Figs. 11(a) and (b) which show $\psi_{\text{Feb}} - \psi_{\text{Dec + Jan}}$ from the observations and the model. The importance of the transient fluxes of vorticity in this change is shown by the same field derived from two simulations with this term omitted from Eq. (3) (Fig. 11(c)). The blocking signature in the eastern Atlantic is totally missing.
Figure 9. The observed zonally asymmetric streamfunction at 250 mb for December + January (a) and February (b). The predictions of these fields by the vorticity equation model (Eq. (3)) are shown in (c) and (d) respectively. The differences of these model solutions from the December–February six-year average (shown in Fig. 3(a)) are given in (e) and (f). These should be compared with the corresponding differences for the observed fields given in Figs. 3(c) and (d). The contour interval is $5 \times 10^5 \text{m}^2\text{s}^{-1}$ throughout.
Figure 10. The results from the vorticity equation model for December + January but using (a) the February zonal flow and (b) the February transient vorticity flux convergence. The fields shown are the difference of the 250 mb streamfunctions from the December + January six-year average. The same field for the model prediction with all December and January inputs is shown in Fig. 9(e) and for the observations in Fig. 3(c). The contour interval is $5 \times 10^4 \text{m}^2\text{s}^{-1}$.

It is clear from these figures that the change in the transient eddies is crucial to the change in the December + January and February mean flows. However, in section 2(c) it was noted that in February there was much enhanced diabatic cooling in the Caribbean in particular. The 250 mb divergence for December + January and February for the region 20°S–30°N, 90°W–0°W are shown in Figs. 12(a) and (b) respectively. The convergence in the Caribbean in February is consistent with the sinking and adiabatic warming in that

Figure 11. The change in 250 mb streamfunction from December + January to February for (a) the observations, (b) the full vorticity model and (c) the vorticity model with zero transient vorticity flux convergence for both periods. The contour interval is $5 \times 10^4 \text{m}^2\text{s}^{-1}$ and only the region from 90°W to 90°E is shown.
region that balanced the diabatic cooling. The divergence near the equator and the eastern and western coasts of S America also increases significantly. 250 mb is rather below the main tropical convective outflow level but Fig. 12 strongly suggests that in February there was a marked increase in the local Hadley-like cell with northward divergent motion at this level. The Climate Analysis Center bulletins give 200 mb velocity potential figures which also suggest such a strong intensification.

In order to judge the importance of the change in mean divergence in this region, a new mean divergence field was created by merging the December + January field in this area with the February field elsewhere. The difference between the February mean divergence field and this modified field is shown in Fig. 12(e). The small values outside the region are due to spectral smoothing. The difference in the model solutions for the full and modified February divergences, both having February zonal flow and transient fluxes, is given in Fig. 13(a). It is seen that the change in mean divergence over the

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Figure 12. The 250 mb divergent flow for (a) December + January and (b) February in the region 90°W–0°W, 20°S–30°N. The contours are those of divergence, smoothed by the same $\Psi^4$ filter as in Fig. 7. The contour interval is $1 \times 10^{-6}$ s$^{-1}$ with the zero contour not drawn and negative contours dashed. The arrows, drawn on a 10° grid, represent the divergent component of the wind, the scale being such that the additional arrow corresponds to 5 m s$^{-1}$. (c) The difference between the February divergence field and a modified field. The modified field is obtained by transplanting into the February divergence the December + January field in the region 90°W–0°W, 20°S–30°N and performing the same smoothing operation mentioned above. Because of the smoothing, the modification extends slightly outside the transplant region.

Figure 13. (a) The difference between the 250 mb streamfunction given by the full vorticity equation model for February and that given by this model with the modified February divergence. (b) The streamfunction response to just the divergence shown in Fig. 12(c). The contour interval is 2.5 $\times 10^{6}$ m$^{2}$ s$^{-1}$ (half that in previous figures) and only the region from 90°W to 90°E is shown.
tropical region of interest from December + January to February is consistent with a large rise in streamfunction (and pressure) in the N Atlantic though not the blocking signature of Figs. 11(a) and (b).

The response to the divergence modification alone (i.e. that shown in Fig. 12(c)) using the February zonal flow and zero transient fluxes is exhibited in Fig. 13(b). The signature is qualitatively similar to that in Fig. 13(a) but quantitatively it is very different. This emphasizes that the forcings in the vorticity equation are large enough that the response is not linear.

4. DISCUSSION

The large and very sudden change in the northern hemispheric flow from December + January to February has been illustrated. The variety of the half-monthly average flows in the region of the Rockies did not lend much credence to the idea that the large February block in the eastern Atlantic could be a result of a resonant interaction with mountain forcing. Vorticity equation diagnostics suggest that the differences in the zonal flow are not directly important, though such calculations cannot rule out their possible indirect role in changing the transients for example. It is the behaviour of the transients that is found to be crucial. Study of daily IPV pictures and also E vector diagnostics (Hoskins et al. 1983) for all transients and high-pass transients alone (not shown here) strongly suggests that the change is predominantly in the synoptic time-scale transients rather than a manifestation of one of the low-frequency barotropic modes of Simmons et al. (1983). If this change is totally internal to the middle latitude atmosphere it is then inherently unpredictable. However, both column-averaged diabatic heating and 250mb divergences suggest the possible importance of the region 10°S–30°N, 90°W–45°W. Vorticity equation diagnostics suggest that the change in 250mb divergence in this area is consistent with enhanced anticyclonic flow in the Atlantic. It is possible that in February 1986 this region provided the catalyst that made the blocking process, in which the transients are vital, much more likely to occur. This conclusion is similar to that of Kok and Opsteegh (1985) who were diagnosing the global response to the 1982–83 El Niño event. However, in that case the crucial role played by the tropics is self-evident.

A subsidiary question is to ask why the change in mean divergence should have occurred in the tropical American region. It is unlikely that it occurred because of the blocking in the eastern N Atlantic. The s.s.t. charts in the Climate Analysis Center Bulletins (1985; 1986a, b) show a region of colder than normal water in the Caribbean and anomalously warm water in the S Atlantic throughout the period. The maximum anomaly in the Caribbean was more than −0.5K in December but tended to weaken as the season progressed. At the same time the S Atlantic anomaly increased so that in February the 1K anomaly line reached to within a few degrees of the equator. A clue to the marked change in the divergent flow in this region may be given by the work of Hastenrath and Heller (1977) and Moura and Shukla (1981). North-east Brazil obtains its rainfall almost entirely in March–April when the ITCZ is in its extreme southerly position. The seasonal southward extension of the rains and the local Hadley cell are enhanced when there is anomalously warm water to the south and cold water to the north. It is possible that the change discussed in this paper is similarly an s.s.t. effect which only becomes important in the latter part of the northern winter when the seasonal migration of rainfall brings it in phase with the s.s.t. anomalies. In this manner, an anomaly in a boundary condition having a long time-scale could make a quite sudden impact on the flow of the atmosphere.
Another possibility that cannot be discounted is that the strong 30–60-day event that occurred in the tropical Indian and west Pacific Oceans in January acted as a trigger for the changes that occurred at the end of January in the central American and N Atlantic regions. Weickman et al. (1985) have given evidence of the global nature of the upper tropospheric streamfunction anomalies that are on average associated with such events.

Study of this one season has suggested what ingredients must be included in dynamical models for predicting on the monthly to seasonal time-scales. GCMs must be able to reproduce faithfully the life-cycles of individual synoptic systems and their feedback onto a blocking flow or a storm-track. Simpler models must include a parametrization of the transient flux convergences which also reproduces these differing feedbacks. They will also probably have to be nonlinear to handle the large changes in the forcing terms. Changes in zonal flow characteristics may be of lesser importance. It is clear that the models must respond to (and may have to predict the evolution of) s.s.t. anomalies and that this response may have to be one that is sensitive to the relative positions of the ITCZ and anomalous s.s.t.s. A good representation of 30–60-day oscillations in the tropics, including their variability, may also be a necessity for predicting changes of regime such as the one that occurred during this season.

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