How does the quasi-biennial oscillation affect the stratospheric polar vortex?

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ABSTRACT

The stratospheric polar vortex is weaker in the easterly phase of the quasi-biennial oscillation (QBO-E) than in the westerly phase (QBO-W), but the mechanism behind the QBO's influence is not well understood. We argue firstly that the composite difference of the atmospheric state between QBO-E and QBO-W closely resembles the structure of the Northern Annular Mode (NAM), the leading empirical orthogonal function of stratospheric variability, including its wave components. Studies of dynamical systems indicate that many different forcings could give rise to this response, and therefore this composite difference does not provide information about the forcing mechanism. The transient response of the vortex to forcing by the QBO is probably much more informative, particularly on time scales shorter than the dynamical time scale of vortex variability, which is about a week. This response in a general circulation model is consistent with the proposed mechanism of Holton and Tan (1980) but does not show the signature of several proposed mechanisms in which the tropical lower stratospheric winds are not important. Our novel approach of examining the transient response to a forcing on short time scales may be useful in various other outstanding problems.

1. Introduction

The most prominent feature of the wintertime polar stratosphere is the westerly vortex that forms around the pole. The Northern Hemisphere (NH) winter vortex is more variable than its Southern Hemisphere counterpart, with breakdowns of the vortex known as major stratospheric sudden warmings (SSWs) happening about six times per decade on average (Charlton and Polvani 2007). As well as being theoretically interesting, understanding how external factors influence this variability may help improve seasonal forecasts of the NH troposphere as it has become realised that weakenings of the vortex give rise to a more negative tropospheric Northern Annular Mode (NAM) (e.g. Baldwin and Dunkerton 1999, 2001; Jung and Barkmeijer 2006).

Holton and Tan (1980) showed that the vortex is influenced by the quasi-biennial oscillation (QBO) (Baldwin et al. 2001; Gray 2010; Anstey and Shepherd 2013). The QBO is a phenomenon that dominates variability in the equatorial lower stratosphere whereby the zonal mean zonal wind (ZMZW) direction on a given pressure level alternates between being easterly and westerly, with the easterly and westerly wind regimes descending with time from the upper to the lower stratosphere. The QBO phase is normally defined as easterly (QBO-E) or westerly (QBO-W) according to the sign of the ZMZW in the lower stratosphere. The average period is 28 months. The vortex is weaker on average in the easterly QBO phase than in the westerly phase by over $10\,\mathrm{ms}^{-1}$ (Holton and Tan 1980; Pascoe et al. 2005). This "Holton-Tan (HT) relationship" has also been found in atmospheric models of varying complexity (e.g. O'Sullivan and Young 1992; Hamilton 1998; Gray et al. 2003; Calvo et al. 2007).

Understanding the mechanism behind this relationship is important for having confidence in observations of apparent non-linear interactions with other forcings, such as with the solar cycle (e.g. Labitzke 2005; Camp and Tung 2007) and the El Niño-Southern Oscillation (ENSO) (e.g. Garfinkel and Hartmann 2007; Wei et al. 2007) and in the seasonal timing of the effect, which is still not well reproduced by models – the only modelling study to report a statistically significant HT relationship in November, when the correlation between equatorial and vortex ZMZW is greatest in observations, is Anstey et al. (2010). It is also important for knowing what models must represent well in order to reproduce the HT relationship and exhibit realistic vortex variability.

The explanation for the HT relationship put forward by Holton and Tan (1980) involved the equatorial winds influencing the waveguide for extratropical planetary waves. Low-wavenumber stationary planetary waves dominate wave forcing of the extratropical NH stratosphere. The surface in the tropics where the ZMZW is zero (also referred to as the "zero wind line") is a critical surface for these waves. Holton and Tan (1980) referred to the work of Tung (1979) who argued this surface ought to reflect planetary waves back towards the pole. In QBO-E, the critical surface in the lower stratosphere is moved polewards into the NH subtropics, so Holton and Tan (1980) suggested this would concentrate wave activity in the NH polar region, weakening the vortex. Henceforth this will be referred to as the "Holton-Tan mechanism". The work of Tung (1979) was based on linear wave theory for which wave amplitudes are assumed to be small, unlike in the real stratosphere, but later work showed the critical surface ought to be reflecting of eddy zonal momentum flux in the time-averaged sense even if wave amplitudes become large (Killworth and McIntyre 1985). Physically this is because if the critical surface sustained absorption of eddy momentum flux then the critical layer would continually widen (Haynes 2003), which cannot happen indefinitely as its width cannot exceed the size of the Earth. (However, the critical layer may be absorbing at certain times as long as this is balanced by over-reflection at other times.) This depends on several assumptions, such as that the fluid motion is two dimensional and that vorticity is conserved in the region being mixed by planetary waves. These assumptions are not strictly met in the real stratosphere, where there is dissipation of wave activity by diabatic and viscous processes and motion is three dimensional; the critical surfaces in the tropics must be at least partially absorbing of eddy zonal momentum flux to be consistent with the observed overall convergence of the flux in the tropics on monthly time scales (Andrews et al. 1987).

Composite analysis has been used to test the HT mechanism. In observations the geopotential height (GPH) wavenumber-1 amplitude and upward component of the Eliassen-Palm (EP) flux (Andrews et al. 1987), which is commonly used as an indicator of planetary wave propagation, are greater in November and December in QBO-E, but in January and February these values are greater during QBO-W (although the difference in these months is not highly statistically significant) (Holton and Tan 1980; Ruzmaikin et al. 2005). Holton and Tan (1982) and Hu and Tung (2002) considered the January–February data not to be consistent with the HT mechanism.

Modelling studies using GCMs indicate that on average in winter there is a greater upward component of the EP flux into the high-latitude stratosphere from the troposphere and greater EP flux convergence in the stratosphere during QBO-E than during QBO-W (e.g. Hamilton 1998; Calvo et al. 2007), although the locations of these effects differ between models. Holton and Austin (1991) and O'Sullivan and Dunkerton (1994) found that the amplitude of planetary waves peaks faster in QBO-E in perpetual winter runs in primitive equation models. These results have been interpreted as being broadly consistent with the HT mechanism.

Kodera (1991) suggested that the ZMZW anomalies associated with the QBO meridional circulation (Baldwin et al. 2001) may also affect planetary wave propagation, but this has been given less attention until fairly recently. Ruzmaikin et al. (2005) suggested the meridional circulation may directly affect the vortex by advection of potential temperature.

Naoe and Shibata (2010) argued that according to the HT mechanism the mid-latitude lower stratospheric EP flux ought to be more poleward in QBO-E, yet this is not the case. Their analysis of composite differences of EP flux between QBO-E and QBO-W in a chemistry-climate model led them to argue that the QBO meridional circulation has an important role in the HT relationship and that the shift of the critical surface in the lower stratosphere is not important. Garfinkel et al. (2012) reached a similar conclusion by examining the transient response to nudging equatorial winds towards QBO-E in a general circulation model (GCM) without coupled chemistry. Yamashita et al. (2011) also argued that the HT mechanism was not consistent with composite differences of EP flux between QBO-E and QBO-W in their chemistry-climate model, and proposed that the southwards critical surface shift in the upper stratosphere is more important than the northwards shift in the lower stratosphere.

However, experiments with a primitive equation model showed that the vortex is more disturbed when equatorial winds are relaxed towards a constant easterly value at all heights, which would not be expected to produce a strong meridional circulation due to the lack of vertical wind shear in the tropics (Gray et al. 2003), or towards easterly jets (Gray et al. 2004). These results suggest that neither the meridional circulation nor a southwards shift of the upper stratospheric critical surface are necessary to produce the HT relationship. This raises the questions of whether GCMs behave differently to the primitive equation model or whether the effects of shifts in the critical surface at different heights do not combine linearly.

Section 2 of this paper describes the observational data, the GCM and the diagnostics we have used. In section 3 we argue that the composite difference of the atmospheric state between the QBO phases closely resembles the signature of the stratospheric NAM and that this is not likely to be helpful for understanding the mechanism behind the HT relationship, in observations or models. In section 4 we argue the full transient response of the vortex to imposing a QBO-E or QBO-W state at the Equator is likely to be much more informative, especially the response shortly after this forcing is applied. We examine this reponse in a GCM in section 5 and show the easterly acceleration of the ZMZW in the tropical lower stratosphere can directly cause increased EP flux convergence and zonal wind deceleration in the high latitude NH stratosphere, consistent with the HT mechanism. We do not see the responses in the eddy momentum flux predicted by other proposals in which the QBO meridional circulation or the upper stratospheric critical surface plays an important role.

We conclude based on these results and on the results of Gray et al. (2003, 2004) that the HT mechanism is the most likely explanation for the HT relationship (section 6).

2. Data and methods

a. Observational data

We use the ERA-40 reanalysis (Uppala et al. 2005) on standard pressure levels from September 1957 to August 2002 to examine the HT relationship in observations. Randel et al. (2004) found that ERA-40 matches stratospheric measurements of the zonal mean circulation derived from radiosonde, rocketsonde and lidar measurements quite closely and performs quite well compared to other analyses and re-analyses, although errors may be substantial in the upper stratosphere above about 5 hPa. Baldwin and Gray (2005) found the ERA-40 QBO to agree well with independent rocketsonde data. ERA-40 agrees well with other analyses in its representation of SSWs (Charlton and Polvani 2007).

For analysis of our ERA-40 data we define the QBO as being in its easterly or westerly phase when the 5S–5N November–February mean ZMZW is easterly or westerly respectively at 50hPa, the pressure at which the correlation between the November–February mean ZMZW averaged over 5S–5N and that at (60N, 10 hPa) is greatest.

b. Northern Annular Mode index

We compare the QBO-E minus QBO-W EP flux and GPH differences to the EP flux signature of the stratospheric NAM, the leading empirical orthogonal function (EOF) of the extratropical stratosphere. We index the NAM by the leading principal component of the monthly-mean GPH north of 20N between 1–100 hPa, weighted by pressure and the cosine of the latitude, calculated using the method of Baldwin et al. (2009). This is similar to the NAM index of Thompson and Wallace (2000), but restricted to the stratosphere. We also reverse the usual sign convention, so that when the index is positive the vortex is weaker, to more easily compare the NAM signature with the QBO-E minus QBO-W differences. Then at each latitude and pressure, the EP flux components and its divergence are linearly regressed against the index, such that the presented signature corresponds to a one standard deviation increase in our NAM index. This is done separately for each calendar month. The correlation between this index and the leading principal components of GPH on individual pressure levels 10 hPa and 50 hPa is 0.97 or greater in each calendar month November–February and the correlation with that at 5 hPa is 0.75 or above, so this index captures variability throughout the NH stratosphere well.

c. GCM simulations

We have performed experiments using the Met Office HadGEM2-CCS GCM. This coupled ocean-atmosphere model has a well-resolved stratosphere, with 60 atmospheric levels in the vertical up to 84 km altitude (corresponding to a pressure of approximately 0.01 hPa) and atmospheric horizontal resolution 1.25° in latitude and 1.875° in longitude. The model includes parameterised orographic gravity wave drag up to 40 km height, using the scheme of Webster et al. (2003), and non-orographic gravity wave drag (NOGWD), using the scheme of Warner and McIntyre (1999) as implemented by Scaife et al. (2002). The NOGWD causes the model to exhibit a spontaneous QBO. The model does not include a chemistry scheme. For full model details see Martin et al. (2011). Osprey et al. (2013) found that HadGEM2-CCS exhibits a realistic stratospheric climatology and realistic variability.

Results from a 240-year pre-industrial control run, which excludes variability due to changing greenhouse gas concentrations, volcanic eruptions and solar variations, were used to confirm that the model reproduces the HT relationship reasonably well (section 5). The correlation between the November–February mean ZMZW averaged over 5S–5N and that at (60N, 10 hPa) is greatest for equatorial winds at 30 hPa, so the sign of the 5S–5N November–February mean ZMZW on this level is used to define the QBO phase for analysis of model data.

Section 5 presents results from experiments designed to examine the transient response of the vortex to nudging towards a QBO-E zonal wind pattern in the tropical stratosphere. We performed a "climatological tropical wind" (ClimEq) control run, which was set up identically to the 240-year pre-industrial control run except that the zonal wind in the tropical stratosphere was nudged towards the ERA-Interim monthly mean climatology between January 1979 and December 2010 (with the climatology at each model time step calculated by linear interpolation between the middle of each month). ERA-Interim is likely to have a better representation of the stratosphere above 10 hPa than ERA-40 (Simmons et al. 2007; Uppala et al. 2008), so it is used in preference to create target equatorial zonal wind profiles. The nudging had the effect of eliminating the QBO, but the mean and standard deviation of wintertime extratropical winds were not strongly affected. We then performed 120 "QBO-E" branch runs of length one month, taking initial conditions at January 1 and Feburary 1 of 60 different years from the ClimEq run (the first two years of this run were not used to allow the model to adjust to the nudging). In these runs the zonal wind in the tropical stratosphere was nudged towards the ERA-Interim climatology plus a typical QBO-E profile. The QBO-E profile was taken as the mean 3D zonal wind anomaly of the 30 months in ERA-Interim with the most negative anomalies in the 5S–5N mean ZMZW at 30 hPa, multiplied by a factor of 3 in order to raise the signal to noise ratio of the vortex response (figure 1) – the equatorial ZMZW anomaly does not become larger than that in observed QBO-E phases in the time scale of 8 days considered in section 5, so this just affects the rate at which equatorial winds adopt a QBO-E profile. This method is similar to that used by Garfinkel et al. (2012), but importantly we focus on the vortex response at shorter times after nudging towards the QBO-E profile is begun.

Nudging was carried out between 21.25S and 21.25N and was implemented by subtracting $\alpha(\phi)(u - u^T)$ from the change in the zonal wind calculated at the end of each time step at each gridpoint, where u is the zonal wind, u^T is the target nudging profile and ϕ is the latitude. The nudging parameter $\alpha(\phi)$ is given by

$$\alpha(\phi) = \frac{1}{20\,\mathrm{days}}\, e^{-2(\phi/16^\circ)^2}$$

between heights of 17.4–39.1 km (corresponding to pressure range 3.3–84.3 hPa). At one model level below and above this range (16.3 km, 103 hPa and 40.9 km, 2.6 hPa) α was set to one half its value within the range and is zero at other heights. The nudging code was adapted from that developed as part of the UKCA project (Telford et al. 2008). The vertical profile of the 5S–5N mean ZMZW differences between the QBO-E runs and the ClimEq run is very similar to that in figure 1 between about 3–100 hPa over the first 8 days, although the meridional width of the nudged QBO-E winds is only 2/3 that of the target profile due to the weakness of the nudging away from the Equator (not shown) – this would most likely cause the influence on the vortex to be weaker than for a perfect imposed QBO-E profile.

d. Diagnostic tools and statistical methods

We present the influence of the QBO on the EP flux $\mathbf{F} = (F^\phi, F^z)$ (Andrews et al. 1987), which is usually taken to show the negative of the zonal mean zonal momentum flux associated with zonal asymmetries. F^ϕ is analagous to the negative of the eddy zonal momentum flux $-\overline{u'v'}$ that was considered by Killworth and McIntyre (1985) in their analysis of the reflectivity of the critical line in their analytical model, where u' and v' are departures from the zonal mean zonal and meridional components of the wind respectively and the overbar indicates zonal averaging. We present the acceleration term $D_F = (\rho_0 a \cos\phi)^{-1}\nabla\cdot\mathbf{F}$ to show where the EP flux is convergent or divergent, where $\rho_0(z)$ is a reference density profile and a is the Earth's radius.

Statistical significances of the QBO-E minus QBO-W composite differences in the EP flux components and in the GPH were calculated according to a Monte Carlo permutation test. Each year was assigned to a surrogate QBO-E or QBO-W group at random, and the composite difference between these random groups was calculated. This was repeated 1,000 times to find the probability distribution of the differences at each grid-point under the null hypothesis that there is no dependence of these variables on the QBO, and the probability that the magnitude of the difference would exceed that of the difference in the data.

To test the statistical significance of the mean differences between the model branch runs nudged to QBO-E and the ClimEq run, a Monte Carlo bootstrap technique was used (Efron and Tibshirani 1993). At each gridpoint a surrogate data sample was generated according to the null hypothesis that the mean difference is zero but other moments of the true distribution of differences equal those in the data. The mean of the differences for all pairs of branch and control runs was subtracted from the difference for each pair and the results were resampled with replacement. The probability of the mean of this resampled data being larger than that for the real data was estimated using 1,000 data resamplings. All significance tests are two-tailed.

3. The observed influence of the QBO on the wave part of the stratospheric circulation

Figure 2 shows the QBO-E minus QBO-W composite difference in the monthly mean NH EP flux and D_F from November to February in ERA-40. In agreement with the findings of previous studies (e.g Dunkerton and Baldwin 1991; Ruzmaikin et al. 2005), the EP flux is more upward in November and December north of 55N, but this signal is not present in January and February, and it has been argued that the late-winter signal is not consistent with the HT mechanism (Holton and Tan 1982). In all months there is a poleward EP flux difference in the tropical lower stratosphere, indicating the equatorward flux is less in QBO-E. This is restricted to latitudes south of about 25N, and Naoe and Shibata (2010) and Yamashita et al. (2011) argued that this means reflection of eddy zonal momentum flux from the lower stratospheric easterlies in QBO-E cannot be directly influencing the vortex. There is greater EP flux convergence in the high-latitude stratosphere in November, and a region of greater convergence that moves downwards from the middle to lower stratosphere between December and February.

Figure 3 shows the regression of the EP flux against our NAM index, showing the anomaly associated with a weaker vortex (note our choice of sign of the index as explained in section 2b). There is a striking resemblance to the QBO-E minus QBO-W composite differences in each calendar month in the extratropics. The flux is more upward north of 55N in November and December and near 45N and 80N in January and more downward near 60N in January and February.

There is a similar correspondence between the QBO-E minus QBO-W composite difference and the NAM signature of GPH. Figure 4 shows the QBO-E minus QBO-W composite difference in the monthly mean NH 10 hPa GPH from November to February in ERA-40. The climatological eddy component, defined as the climatological GPH with the zonal mean subtracted, is also shown. Figure 5 shows the signature of the NAM in NH 10 hPa GPH, which bears a very good resemblance to the GPH differences between QBO-E and QBO-W. Anomaly correlations with the NAM signature north of 20N, calculated with gridpoints weighted by the square root of the cosine of the latitude, are indicated below the composite GPH differences in figure 4. The correlations are between 0.75 and 0.95 in November–January, with a lower correlation in February when the differences are not highly statistically significant. The correlations are all greater if only anomalies north of 60N are considered, so this is not simply arising from a direct influence of the QBO on the subtropics.

In November, the top left panel of figure 4 shows GPH is greater in QBO-E over the Canadian Arctic and less over northern Europe. This represents positive interference with the wavenumber-1 part of the climatological wave pattern, and so the wavenumber-1 amplitude is greater, whilst the wavenumber-2 amplitude is slightly less (figure 6). Over the course of winter, however, the pattern shifts so that more positive GPH is found over the Arctic and north Atlantic with lower GPH over the North Pacific in January and February. This gives weak destructive interference with the climatological waves, so wavenumber-1 and 2 amplitudes are both slightly reduced. Figure 6 also shows the change in the wavenumber-1 and 2 amplitudes at 60N if the NAM signature in GPH multiplied by the QBO-E minus QBO-W composite difference in the index is added to the climatological GPH, showing that the observed changes in the wave amplitudes in QBO-E versus QBO-W correspond closely to the seasonal evolution of the NAM signature.

Previous work has noted that the QBO-E minus QBO-W composite difference of ZMZW is very similar to the NAM signature (Dunkerton and Baldwin 1991; Kodera 1995; Ruzmaikin et al. 2005), but here we show that this is true for the EP flux and GPH wave amplitude differences as well, which is important given that these differences have been used to try to understand the mechanism of the QBO's influence.

A correspondence between the leading EOF of a system and its response to an applied forcing is a commonly observed feature of dynamical systems, if the response is averaged over time scales that are long compared to the dynamical time scales. Palmer and Weisheimer (2011) illustrate that in the simple system of Lorenz (1963) (that which gives rise to the famous Lorenz butterfly attractor), applying a steady force in any direction in the xy-plane gives rise to a shift in the system's mean state which is very nearly aligned with the system's leading EOF, so the spatial pattern of the response closely resembles that of the leading EOF. Ring and Plumb (2008) found in a tropospheric GCM that the steady-state response to various mechanical and thermal forcings applied in the extratropics closely resembles the tropospheric NAM. Branstator and Selten (2009) examine the reasons why the response to greenhouse gas forcing in a tropospheric GCM is NAM-like. They conclude that it largely results from a linear effect whereby anomalies in the NAM tend to persist for a long time, so the NAM is a prominent pattern in natural variability and in the response to a forcing after time-averaging. This behaviour is also predicted by the fluctuation-dissipation theorem (Gritsun and Dymnikov 1999), which has been argued to apply approximately to the atmosphere (e.g. Leith 1975; Gritsun and Branstator 2007).

Invoking this behaviour, which seems robust in atmospheric models and also present in simpler systems like that of Lorenz (1963), can then explain the seasonal evolution of the pattern of EP flux differences in QBO-E and QBO-W, and supports the suggestion of Dunkerton and Baldwin (1991) that the QBO "excited a fundamental...mode of variability in the extratropical atmosphere". We do not know of any studies that have examined this phenomenon in the stratosphere. It does however seem consistent with previous literature identifying the stratospheric NAM as a prominent pattern that appears not just in response to the QBO but also to other important natural influences on the vortex (volcanic eruptions (e.g. Kodera 1995; Stenchikov et al. 2006), ENSO (e.g. Sassi et al. 2004) and the solar cycle (e.g. Kodera 1995; Labitzke 2005)).

Importantly, this behaviour implies that examining the response to a forcing averaged over a long time or using compositing does not in general yield information about the forcing mechanism. Observing an anomaly in response to an unknown forcing that resembles the leading EOF does not allow the nature of the forcing to be deduced if many different forcings can give rise to this response. Therefore, the HT mechanism is consistent with the vortex response to the QBO as it simply predicts there should be enough anomalous EP flux convergence in the high-latitude stratosphere in QBO-E to cause deceleration of the vortex, as is observed, and the above discussion indicates this may manifest itself as a modulation of the NAM in composite analysis. However, these observations could also be consistent with other mechanisms that predict a weakening of the vortex during QBO-E. Furthermore, structure in the extratropical QBO-E minus QBO-W composite differences seems to be primarily related to the NAM signature rather than the forcing mechanism, and is not a reliable indicator of the mechanism.

Another way to understand the difficulty in using these composite differences to infer the forcing mechanism behind the HT relationship is that the differences in the wave components of the flow have contributions not only from the QBO-E equatorial wind pattern but also the effect of the weaker vortex, as changes in the zonally symmetric component of the flow will cause changes in the wave components. What is required is a way of computing the changes due to QBO-E whilst the vortex state is close to constant.

4. Motivating examination of the short-term transient response

Here we argue that whilst the QBO-E minus QBO-W composite difference cannot be relied on to show the mechanism of the QBO's influence, the full time-dependent transient response should be much more useful.

If the forcing mechanism is simple (meaning it does not influence the system in question in a series of intermediate steps), then the response on short time scales following application of the forcing can be expected to show the mechanism clearly. Consider a system described by state vector $\mathbf{x}(t)$ that evolves according to equations

$$\dot{\mathbf{x}}(t) = \boldsymbol{\mathcal{L}}(\mathbf{x}(t), t), \tag{1}$$

where the equations may be non-linear and are explicitly time-dependent for generality. The equations governing atmospheric motion may be expressed in this form. Consider also a forced variant of this system described by state vector $\mathbf{x}'(t)$ which evolves according to similar equations with the addition of a state- and time-dependent forcing term,

$$\dot{\mathbf{x}}'(t) = \boldsymbol{\mathcal{L}}(\mathbf{x}'(t), t) + \mathbf{f}(\mathbf{x}'(t), t) \tag{2}$$

with $\mathbf{x}(0) = \mathbf{x}'(0) = \mathbf{x}_0$. In the context of the HT relationship, $\mathbf{x}'$ would represent the vortex state and $\mathbf{f}$ the influence of the QBO. Then as long as the difference between the state vectors $\delta\mathbf{x} = \mathbf{x}'(t) - \mathbf{x}(t)$ is analytic, which will be the case if both $\mathbf{x}(t)$ and $\mathbf{x}'(t)$ are analytic, $\delta\mathbf{x}(t)$ can be evaluated for small t by writing its Taylor series (in index notation using summation convention) to give

$$\delta x_a(t) \quad = \quad f_a(\mathbf{x}_0, 0)\, t + \left(\left.\frac{\partial \mathcal{L}_a}{\partial x_b}\right|_{(\mathbf{x},\, t)=(\mathbf{x}_0,\, 0)} \delta \dot{x}_b(0) + \left.\frac{\mathrm{d} f_a}{\mathrm{d}t}\right|_{(\mathbf{x}, t)=(\mathbf{x}_0,\, 0)} \right) \frac{t^2}{2} + \mathcal{O}(t^3). \tag{3}$$

Thus for short times, the difference between the systems is nearly proportional to the applied forcing at $t = 0$, when this is non-zero. State-dependence of the equations of motion (non-zero $\partial\mathcal{L}_a/\partial x_b$) acts to complicate the relationship between $\delta\mathbf{x}$ and the forcing as t increases and may be expected to become important on a time scale of the order of the system's dynamical time scale. (If $\mathbf{f}(\mathbf{x}_0, 0) = 0$, it can similarly be shown that $\delta\mathbf{x}(t)$ is proportional to $\mathrm{d}\mathbf{f}/\mathrm{d}t|_{(\mathbf{x},\,t)=(\mathbf{x}_0,\,0)}$ to $\mathcal{O}(t^2)$ when this is non-zero, which is relevant for the results of the nudging experiments in section 5.) Examining the system's response on short time scales shows the effect of the forcing before effects due to the change in the state of the system become large.

If the forcing mechanism unfolds in several steps then the short-term transient response would be expected to show the steps that develop up to a time scale of the order of the dynamical time scale, which may still give useful information for testing hypotheses.

For studies of the atmosphere, it is probably necessary to use a numerical model to evaluate the transient response. Using observations would require identifying two near-identical atmospheric states with different values of the forcing in question, which is practically impossible.

5. The transient response of the vortex to QBO-E forcing

Firstly we show the HT relationship in the HadGEM2-CCS GCM. Figure 7 shows the QBO-E minus QBO-W composite difference in the January–February mean ZMZW in ERA-40 and in the 240-year pre-industrial control run of HadGEM2-CCS – the monthly-mean difference is not statistically significant earlier in winter in HadGEM2-CCS. The model reproduces the weakening of the vortex seen in observations, with ZMZW differences that are somewhat smaller at high latitudes. The lack of a HT relationship in November and December in the model may be due to the equatorial winds simply having too weak an influence on the vortex, and it taking time for their impact to accumulate and give rise to an appreciable vortex response, following the suggestion of O'Sullivan and Young (1992). It could also be related to the model exhibiting slightly less total variability than in observations in early winter (Osprey et al. 2013), which is a common problem in stratosphere-resolving GCMs. As noted in section 1, few modelling studies have reported an HT relationship in early winter. The mechanism by which the equatorial winds influence the vortex in the model is likely to be qualitatively similar to the mechanism in the real atmosphere.

Figure 8 shows the QBO-E minus QBO-W composite difference in the monthly mean NH EP flux in January and February in the model. In these months the model reproduces the pattern of the observed influence of the QBO on the EP flux well, with greater upward flux in the high-latitude stratosphere in January in QBO-E and more poleward and downward flux in February, bearing in mind that the observed differences are not highly statistically significant. The EP flux differences are somewhat smaller than in ERA-40, however, consistent with the ZMZW differences being smaller.

We have examined the transient response of the vortex to nudging equatorial winds towards a QBO-E state on a time scale of about a week, which is comparable to the mid-stratosphere's dynamical time scale. Figure 9 shows the mean over our ensemble of QBO-E branch runs of the difference in the ZMZW between the branch runs and the ClimEq run. The QBO-E pattern of winds is visible in the tropics, with mid-latitude anomalies that arise due to the Coriolis force acting on the QBO meridional circulation. The extratropical winds in the lower stratosphere initially strengthen and this appears to be associated with the cells of the QBO meridional circulation. However, between days 5–8 following the start of the nudging there is a weakening of the upper stratospheric winds which descends with time, as expected from the HT relationship.

Figure 10 shows the mean difference in the EP flux between the branch runs and the ClimEq run. On days 1–2 between ∼15–30N the EP flux is more equatorward in the lower stratosphere and poleward in the middle and upper stratosphere. Between days 3–8, however, EP flux differences become statistically significant at higher latitudes. The EP flux is more poleward in the lower stratosphere and more equatorward in the upper stratosphere, extending to the polar region by days 5–6. This is associated with increased EP flux convergence in the high-latitude stratosphere between $\sim$2–20 hPa, which is consistent with the weakening of the winds in this region in figure 9. The initial increased absorption of EP flux in the lower stratosphere in days 1–2 next to the critical surface followed by increased reflection seems consistent with the "Stewartson–Warn–Warn" solution for the critical layer discussed by Killworth and McIntyre (1985).

The more poleward EP flux from the tropics to the high latitudes is the signal that Naoe and Shibata (2010) argued should be present if the HT mechanism is correct. Our results are thus consistent with the HT mechanism and indicate that increased reflection of eddy zonal momentum flux in QBO-E may directly affect the polar stratosphere. Thus the HT mechanism cannot be ruled out of playing a part in the HT relationship as argued in recent studies. Note that this signal is unlike the QBO-E minus QBO-W composite differences in the EP flux in any calendar month (figure 2), illustrating the necessity of examining the transient response to understand the influence of the QBO.

However, it is not just the tropical lower stratospheric easterly wind acceleration that matters – the westerly acceleration above 10 hPa appears to cause a more equatorward EP flux that decreases its convergence at high latitudes, reducing the QBO influence there. Figure 11 shows the EP flux differences using the "acceleration scaling" defined by Gray et al. (2003), to indicate the size of the acceleration associated with the flux. By days 7–8 there is more poleward flux in the QBO-E runs at 1 hPa and above which appears associated with the tropical ZMZW at these heights becoming more easterly (figure 9). Easterly anomalies at these altitudes are also observed in QBO-E in winter (Pascoe et al. 2005). Although the magnitude of this momentum flux is small, it is associated with EP flux convergence and deceleration of the ZMZW in the mid-latitude upper stratosphere. If the vortex state is sensitive to ZMZW changes in this region then this provides a way for tropical upper stratospheric wind changes to contribute to the HTR, as suggested by Gray et al. (2001a,b).

These results do not indicate that the EP flux convergence in the high-latitude stratosphere arises due to other recently proposed mechanisms discussed in section 1 – the EP flux differences have the opposite sign to those predicted by Naoe and Shibata (2010) and Yamashita et al. (2011) and its convergence in mid-latitudes has opposite sign to that predicted by Garfinkel et al. (2012). It is not clear if there are reasons why the signatures of these mechanisms would take more than 8 days to appear, but these results make it seem less likely that any of these mechanisms are dominant.

The greater reflection of EP flux in the lower stratosphere in the QBO-E runs occurs despite the zero wind line (ZWL) shifting both poleward and equatorward on different levels below 10 hPa. As remarked in section 2c, the QBO profile is too narrow meridionally, and the ZWL shift may have been more pronounced if this were not the case. The ZMZW=c lines show a poleward shift on average at all levels between 15–80 hPa by up to a few degrees for $-5\,\mathrm{m/s} \leqslant c \leqslant -2\,\mathrm{m/s}$. The greater reflection in the QBO-E runs may therefore be associated with waves with small negative phase velocities. As the critical surfaces are partially absorbing (section 1), it may also be due to EP flux associated with stationary waves being affected by the ZMZW south of the ZWL. It is not clear that the ZWL should be considered fundamental given it is unclear whether wave phase velocities can be defined in the subtropics, and the EP flux may be affected by tropical ZMZW in a region of finite width.

The QBO meridional circulation has roughly the same strength relative to the equatorial ZMZW differences in our branch runs as in the free-running model, indicating that if this circulation were having a substantial direct effect on the vortex then it should be evident in our results. The maximum 10S–10N mean difference in the downwelling of the residual meridional circulation (Andrews et al. 1987) between the branch runs and the ClimEq run over days 7–8 after branching, which is at 15 hPa, is 1.8×10^{-4} m/s. The QBO-E minus QBO-W difference in this quantity divided by two in the pre-industrial control run is

1.3×10^{-4} m/s. The ratio of these quantities is 1.4. The ratio of the peak equatorial wind differences in the lower stratosphere between the branch runs and the ClimEq run averaged over days 7–8 (-20 m/s) and the QBO-E minus QBO-W composite peak equatorial wind differences in the pre-industrial control run divided by two (-11 m/s) is 1.8. These ratios are quite similar, indicating that the meridional circulation has about the correct strength in our experiment.

The ensemble-mean results we present show the average effect of nudging towards a QBO-E state but may hide sensitivity to the initial conditions. The theory of critical layers only indicates they may be absorbing for finite time periods and the influence of the QBO may be different in such periods. The QBO influence may also be reduced if critical surfaces exist in mid-latitudes. Gray et al. (2003) suggest the QBO may have a weaker influence if the tropospheric wave forcing is very strong or very weak as well.

The signal loses statistical significance after day 8, which appears to be because the standard deviation of the differences between the runs grows exponentially with time (not shown), as expected from exponential perturbation growth in a chaotic system such as the atmosphere. The ZMZW and EP flux differences between the branch and ClimEq runs grow approximately linearly with time over the first week, and so the loss of statistical significance is consistent with this signal continuing to grow but the noise growing more quickly so that the signal to noise ratio decreases with time. The magnitude of the noise must saturate after some time at the level of the climatological variability. The fact that there is an HT relationship in the 240-year pre-industrial control run implies that the signal to noise ratio would be large enough again after several months for the signal to be statistically significant, implying that the signal would continue to increase in time. Seeing the full evolution of the vortex response to QBO-E forcing in this model would likely require a prohibitive amount of computing resources, and similar experiments with computationally cheaper models may be helpful to understand the transient response fully. The transient response to nudging towards QBO-E presented by Garfinkel et al. (2012) appears to be more statistically significant, but the model they used has unrealistically low vortex variability, which may be part of the reason for the difference.

It should be considered whether the use of nudging in the tropical stratosphere would cause differences between the coupling between the tropics and extratropics in our model runs and in the real stratosphere. Previous studies have used nudging of equatorial winds to examine the HT relationship and have obtained realistic results (e.g. Hamilton 1998; Garfinkel et al. 2012), implying that the nudging does not drastically interfere with the coupling between the tropics and extratropics. The nudging we have used would tend to dampen wave activity in the tropical region with time scales longer than a couple of weeks. The most unrealistic effect of the nudging may be to create a QBO profile that is too narrow meridionally, which may weaken the extratropical response. We see no reason why these effects would qualitatively change the interaction between the tropics and extratropics.

The atmospheric response to an applied forcing at short times following application of the forcing is indicative of the whole forcing mechanism only if the steps of the mechanism unfold on a time scale less than about the system's dynamical time scale, so mechanisms whose early stages are consistent with our results but which also involve subsequent steps cannot be ruled out. Our results also do not rule out the meridional circulation having a role through modifying the background state through which the eddy zonal momentum propagates. Our experiments also show the effect of nudging towards QBO-E from a state having close to climatological equatorial winds – as the winds approach a full-strength QBO-E state, the vortex response may depend non-linearly on further increases in the strength of the QBO-E profile. This could be explored in a similar way to our method, using a control run nudged towards a QBO-E state with branch runs nudged towards a stronger QBO-E state. In our runs the equatorial ZMZW differences do reach the approximate magnitude of the anomalies in the QBO-E phase in observations in the time scale considered here, however, allowing the possibility that non-linearity will show an effect, so we do not expect that the effects of non-linearity with respect to the equatorial winds would greatly change our results. Our results leave open the question of whether the HT mechanism can account quantitatively for the HT relationship.

The key innovative part of this work has been to examine the stratospheric response to QBO forcing on time scales that are shorter than or of the order of the mid-stratosphere's dynamical time scale of $\sim$1 week, so we see the response to QBO-E forcing alone before the circulation has evolved. If our EP flux differences are averaged over the first 16 days, as done by Garfinkel et al. (2012), then the poleward EP flux difference from the tropics to high latitudes is no longer apparent.

6. Summary and conclusions

We have discussed investigations into the mechanism by which the QBO influences the stratospheric polar vortex. Understanding this mechanism is important for improving seasonal forecasts of the vortex, and therefore also of the troposphere (e.g. Baldwin and Dunkerton 2001), for understanding non-linear interactions of this relationship with other forcings such as the solar cycle (e.g. Labitzke 2005) and ENSO (e.g. Garfinkel and Hartmann 2007) and understanding the seasonal timing of the effect. Previous observational and modelling studies, when considered together, have not come down firmly in favour of any of the proposed mechanisms.

We have shown that composite differences of the wave components of the stratospheric circulation between QBO-E and QBO-W are very similar to the signature of the NAM, the leading EOF in the stratosphere. This behaviour is qualitatively similar to that of other dynamical systems, in that different forcings applied to the same system are found to give a response similar to that system's leading EOF. We have argued that this implies that QBO-E minus QBO-W composite differences are not likely to be informative about the mechanism behind the HT relationship, since many different mechanisms could give rise to the NAM-like vortex response.

We then showed that the full transient response of a system to a given forcing should be much more informative about the forcing mechanism. The system's transient response on short time scales after application of the forcing will show the steps of the forcing mechanism that unfold up to about one dynamical time scale, about one week in the mid-stratosphere, and will closely resemble the forcing if the forcing mechanism essentially involves only one step.

Our examination of the first few days of the transient response of the vortex to nudging of equatorial stratospheric winds towards a QBO-E state in the HadGEM2-CCS GCM indicates that the EP flux becomes less equatorwards between the tropics and high latitudes in the lower stratosphere, there is greater convergence of the EP flux in the high-latitude stratosphere between about 2–20 hPa and that the westerly wind in this region decelerates. This is consistent with the hypothesis that more easterly winds in the tropical lower stratosphere cause greater reflection of eddy zonal momentum flux towards the polar stratosphere, causing deceleration of the westerly winds, as suggested by Holton and Tan (1980). Our results do not show the signatures of the mechanisms suggested by Naoe and Shibata (2010), Yamashita et al. (2011) or Garfinkel et al. (2012) who proposed that the role of the lower stratospheric zero wind line is less important than that of the QBO meridional circulation or the upper stratospheric zero wind line. Combining this with the results of Gray et al. (2003, 2004), which indicate that the meridional circulation and the shift of the zero wind line in the upper stratosphere (in the opposite sense to that in the lower stratosphere) are not essential to produce the HT relationship, we conclude that the mechanism of Holton and Tan (1980) is the most likely explanation for the HT relationship, and that the behaviour of HadGEM2-CCS is consistent with the behaviour of the primitive equation model used by Gray et al. (2003, 2004) in showing that easterly acceleration of tropical winds leads to deceleration of the ZMZW at high latitudes. However, it is not clear if the role of the zero wind line is fundamental and the total anomalous poleward EP flux depends on ZMZW changes throughout the depth of the tropical stratosphere. We have not ruled out a role for the meridional circulation through modifying the background state through which the eddy zonal momentum flux propagates, or mechanisms that include multiple steps that unfold on a time scale greater than about a week.

To our knowledge, using the transient response of the atmosphere to a forcing on time scales of a few days to understand the forcing mechanism is novel and may be of use in various other outstanding problems, such as for understanding the downward influence of the stratosphere on the troposphere (e.g. Baldwin and Dunkerton 2001) or the dynamical influence of the solar cycle (e.g Labitzke 2005).

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List of Figures

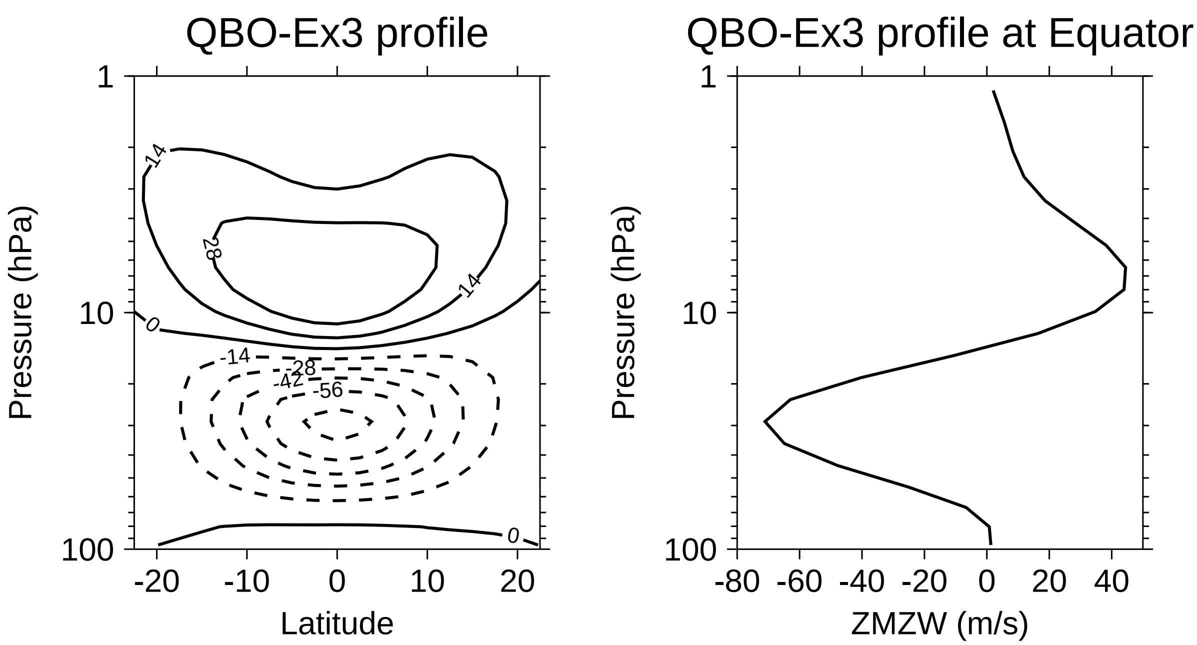

FIG. 1. Left: the zonal mean of the target "QBO-E times 3" zonal wind profile used in the GCM experiments with nudged equatorial winds. Right: the 5S–5N mean of this profile.

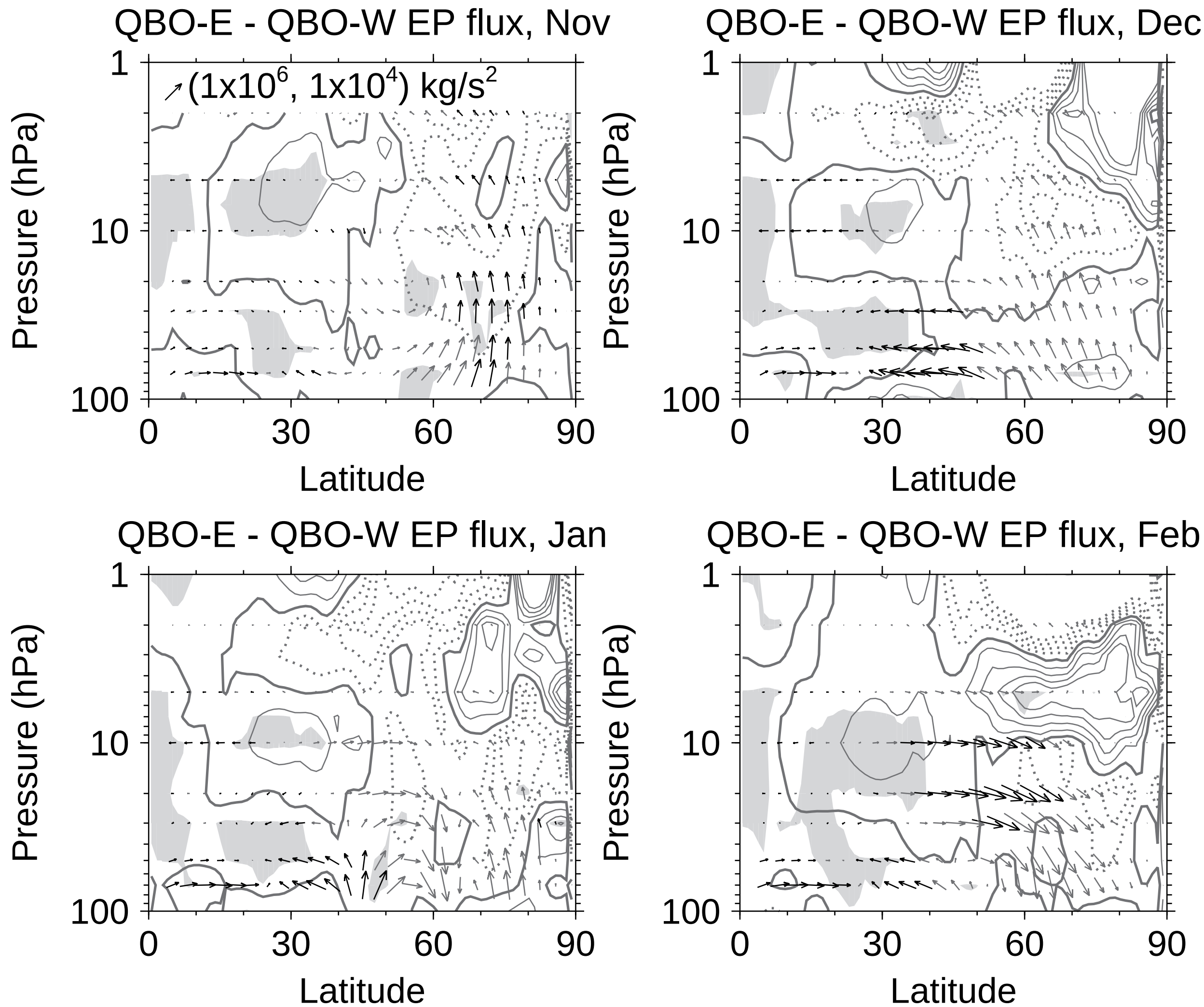

Fig. 2. QBO-E minus QBO-W composite differences in ERA-40 of the wintertime EP flux (arrows, black where either the F^{ϕ} or F^{z} differences are statistically significant above the 95% level and grey otherwise) and D_F (contours, plotted at 0.5 m/s/day intervals, with negative contours dotted and a thickened zero contour). The EP flux is shown at pressures 2, 5, 10, 20, 30, 50 and 70 hPa and every 3.75° in latitude. A reference arrow is shown in the top left plot along with its (F^{ϕ}, F^{z}) values. Shading shows where D_F differences are statistically significant above the 95% level.

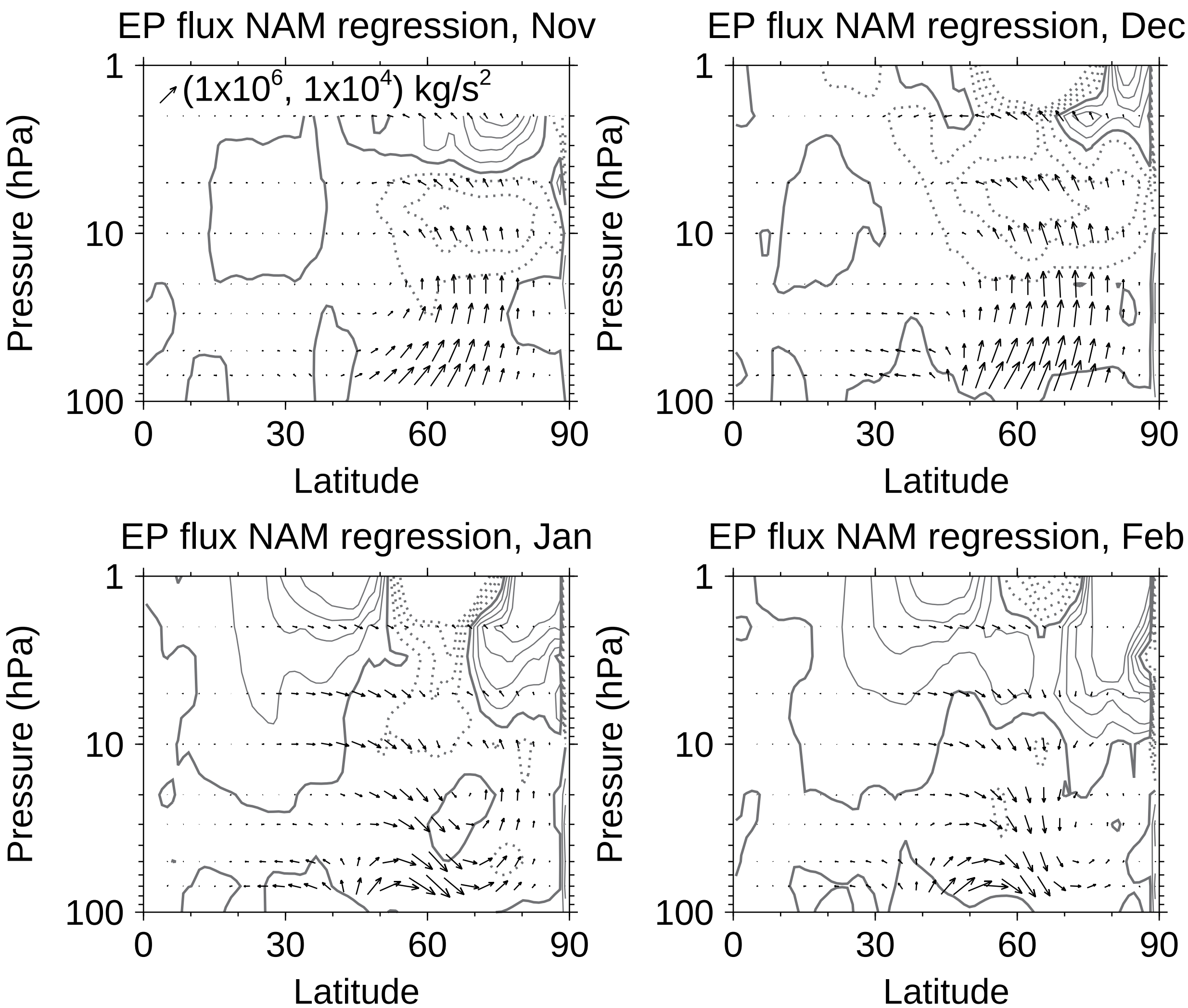

FIG. 3. Regression of the EP flux and D_F onto our NAM index in ERA-40, showing the anomaly associated with a weaker vortex with our choice of sign of the index, plotted as for the QBO-E minus QBO-W composite differences in figure 2. There is a good correspondence between these regression patterns and the composite differences.

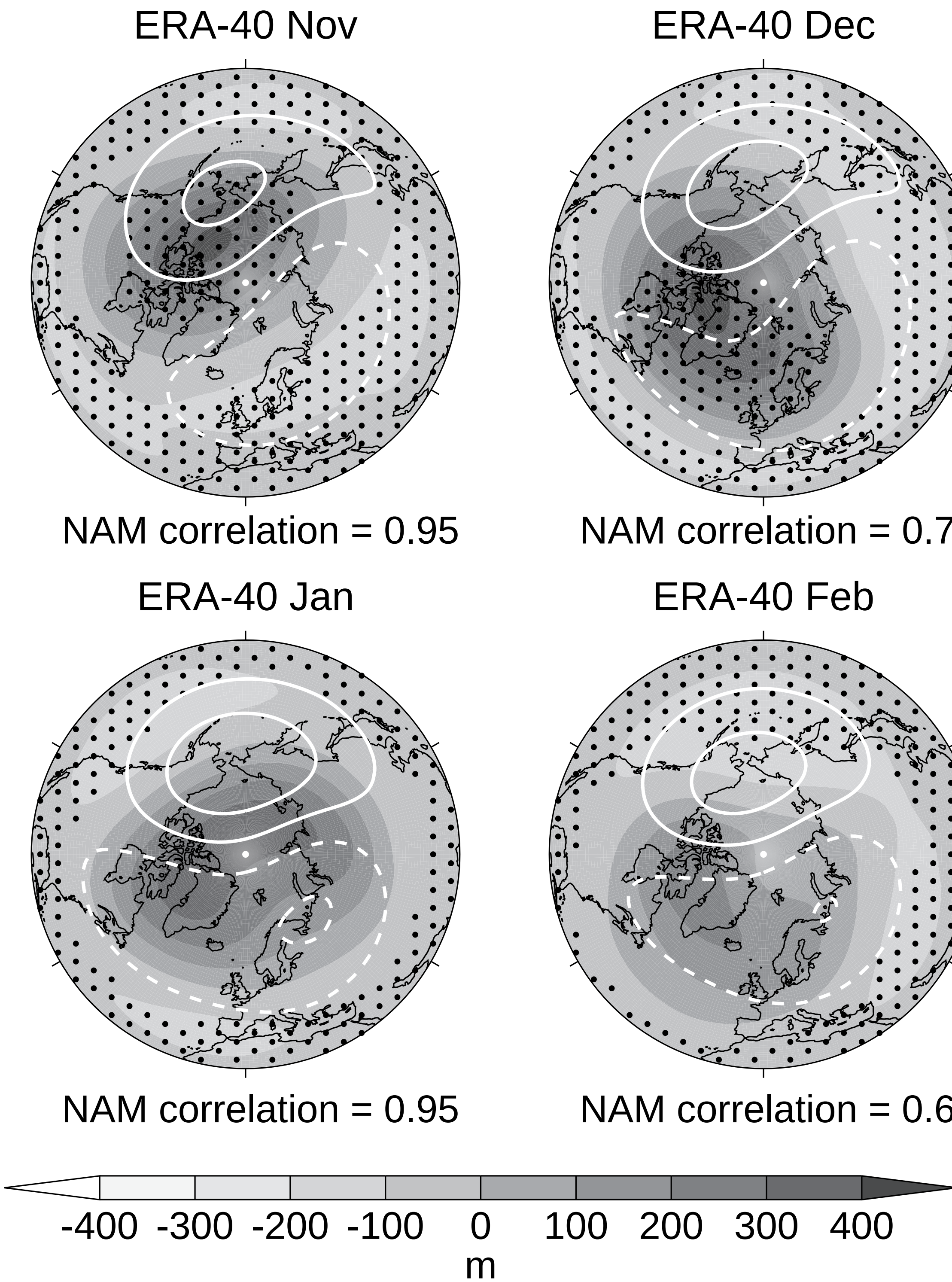

FIG. 4. QBO-E minus QBO-W composite differences in ERA-40 of the wintertime GPH at 10 hPa (greyscale) north of 20N. White contours show the climatological zonally asymmetric component of GPH with contour values ± 200 and ± 600 m with negative contours dashed. NAM correlation values indicate the anomaly correlation of the composite differences with the NAM signature north of 20N shown in figure 5. Stippling shows where GPH differences are statistically significant above the 95% level.

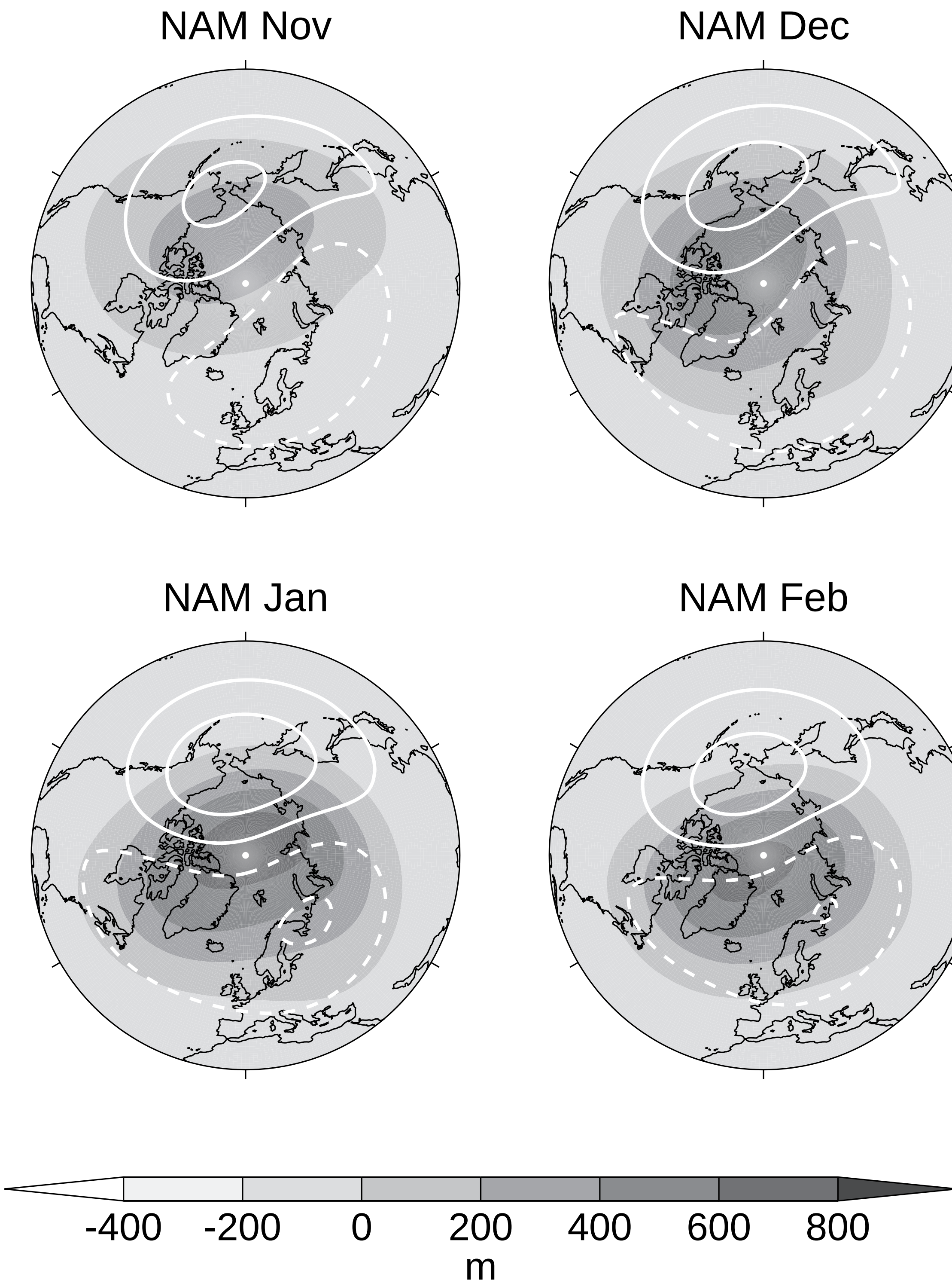

Fig. 5. Regression of 10 hPa GPH north of 20N onto our NAM index in ERA-40, showing a close resemblance to the QBO-E minus QBO-W composite differences in figure 4. White contours show the climatological zonally asymmetric component of GPH with contour values ±200 and ±600 m with negative contours dashed.

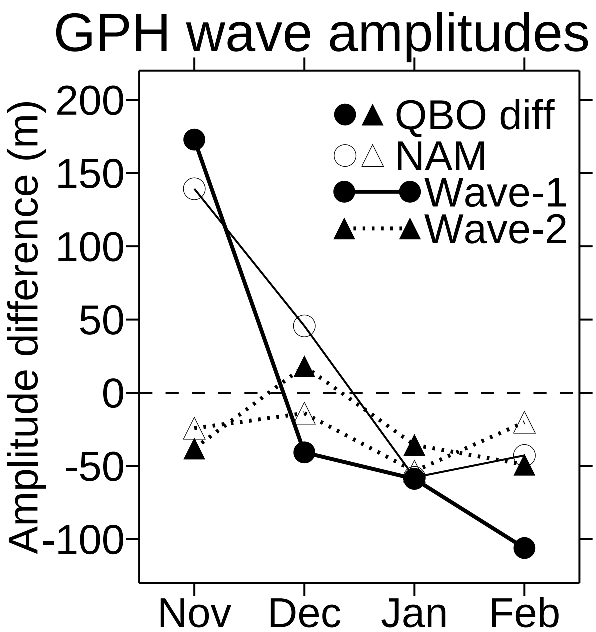

Fig. 6. Lines with filled symbols show the QBO-E minus QBO-W composite differences of GPH wavenumber-1 amplitude (solid lines and circles) and wavenumber-2 amplitude (dotted lines and triangles) at 60N and 10 hPa for months November–February. Lines with unfilled symbols show the differences resulting from adding the NAM signature in GPH multiplied by the QBO-E minus QBO-W composite difference in the NAM index to the climatological GPH, showing that the main qualitative features of the QBO-E minus QBO-W differences in GPH wave amplitudes are largely explained just by QBO modulation of the NAM.

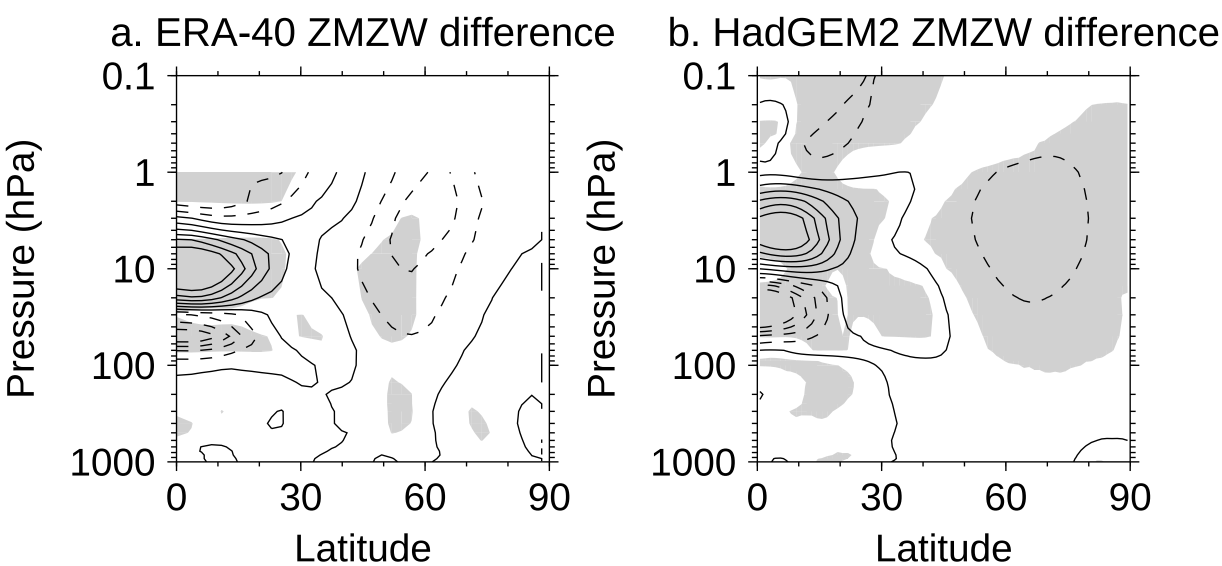

FIG. 7. QBO-E minus QBO-W composite differences of January–February mean ZMZW in ERA-40 (a) and in HadGEM2-CCS (b). HadGEM2-CCS reproduces the differences in ERA-40 reasonably well, with smaller high-latitude differences. Contour levels are every 4 m/s with negative contours dashed and the zero contour thickened.

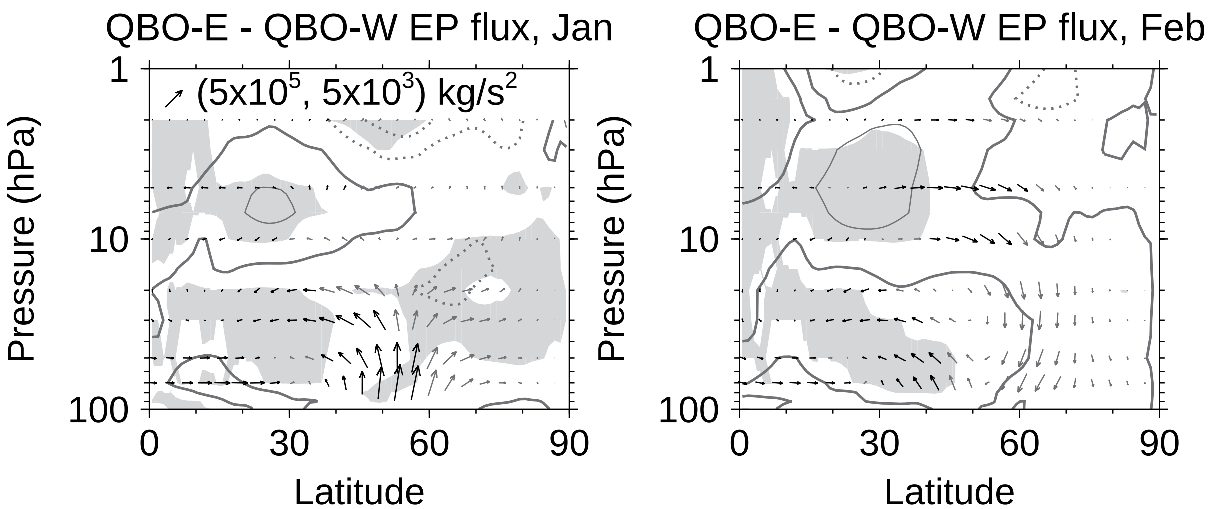

FIG. 8. As in figure 2 but for HadGEM2-CCS, showing only January and February when the differences are statistically significant. In these months HadGEM2-CCS reproduces the pattern of QBO-E minus QBO-W composite EP flux differences in ERA-40 quite well.

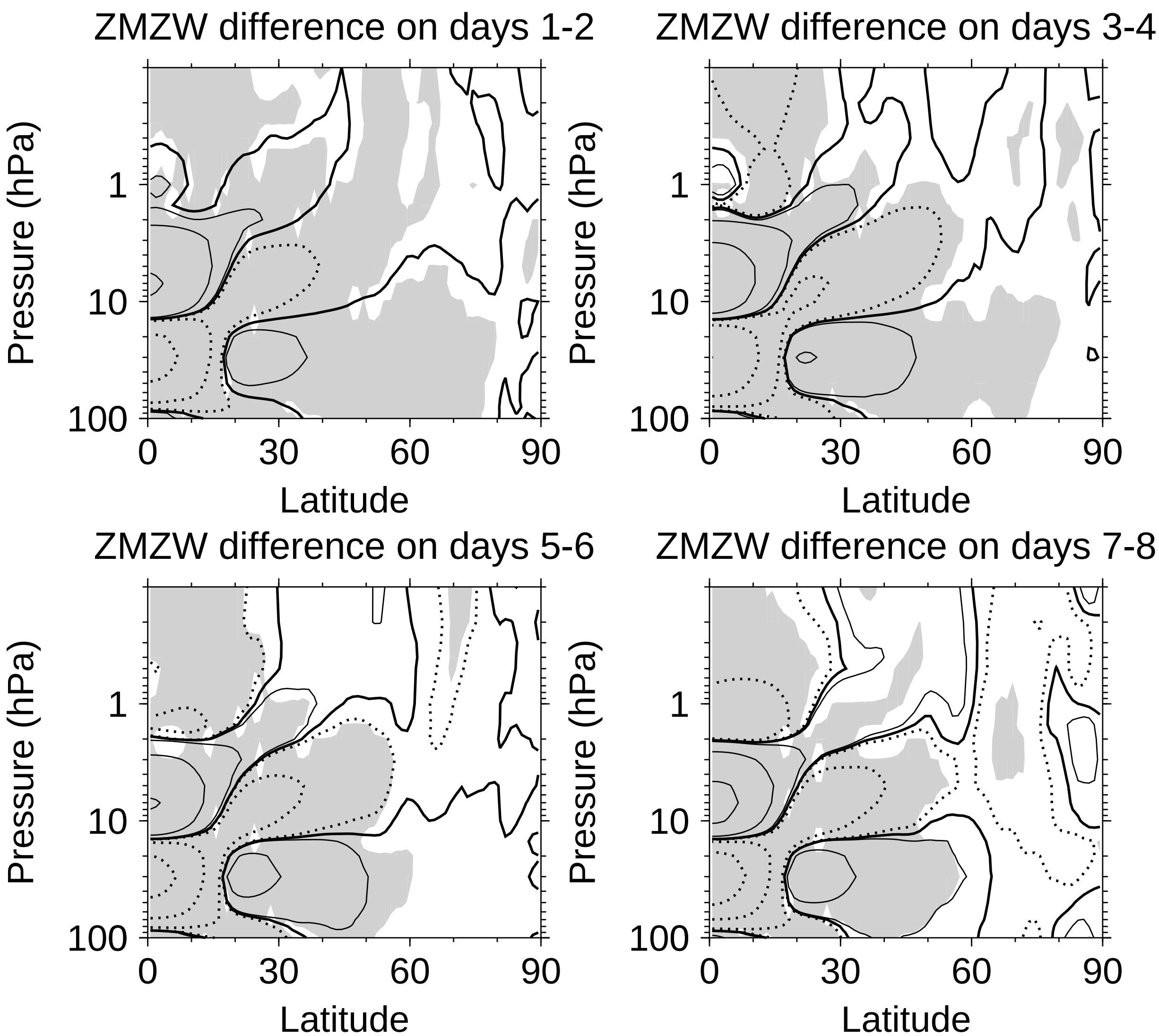

FIG. 9. Ensemble mean ZMZW differences between branch runs nudged towards the QBO-E equatorial wind profile and the ClimEq run averaged over each two day interval up to 8 days following branching. Contours are at 0, ± 0.02, ± 0.2, ± 2 and ± 10 m/s with negative contours dotted and the zero contour thickened. Shading shows where differences are statistically significant at the 95% level.

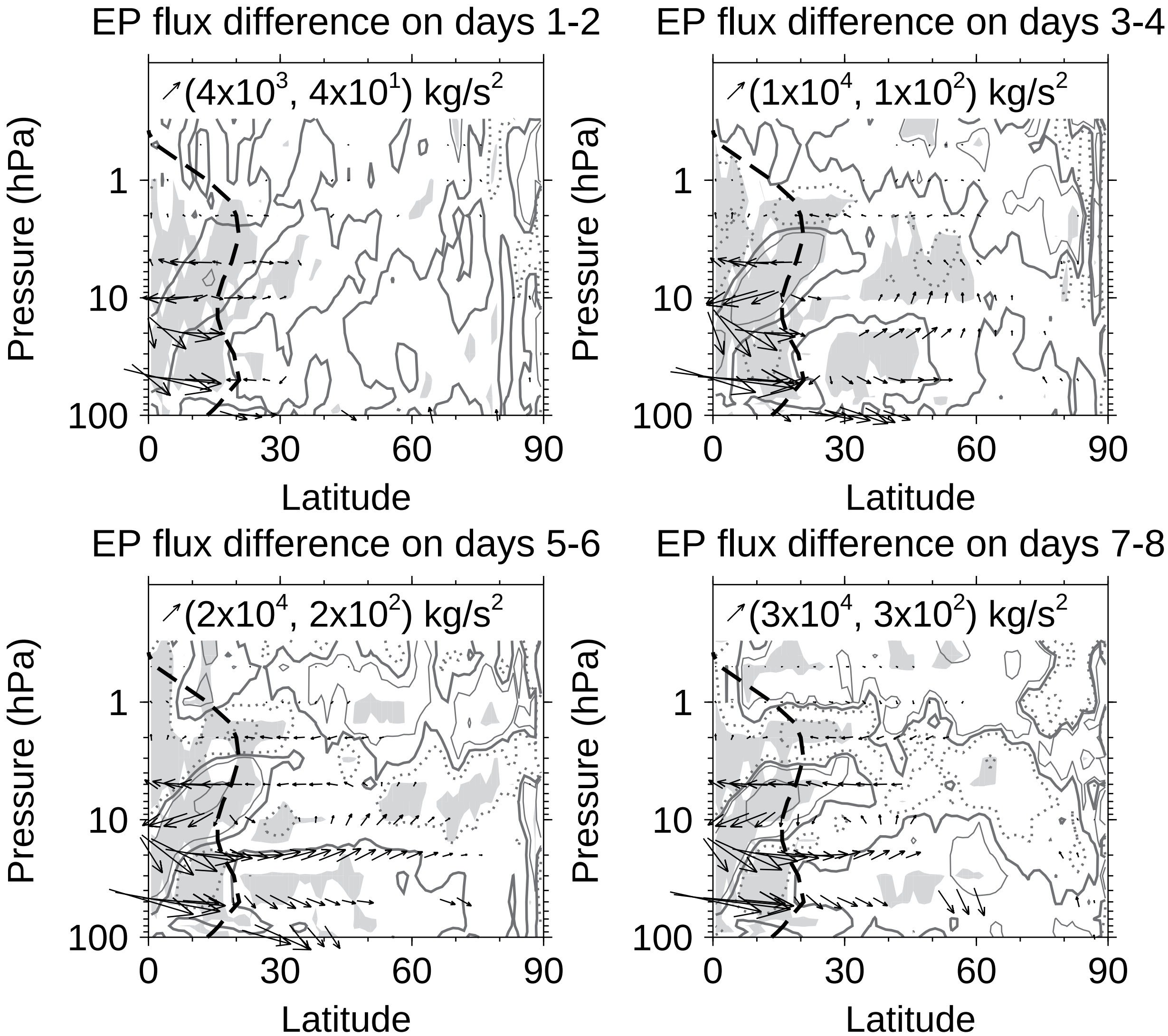

Fig. 10. Ensemble mean differences between branch runs nudged towards the QBO-E equatorial wind profile and the ClimEq run averaged over each two day interval up to 8 days following branching in EP flux (arrows, shown only where either the F^{ϕ} or F^{z} differences are statistically significant above the 95% level) and D_F (contours, at 0, ± 0.02 and ± 0.1 m/s/day, with negative contours dotted and the zero contour thickened). The thick dashed line shows the zero wind line in the days 1–8 mean ZMZW in the ClimEq run. The EP flux differences are shown at pressures 0.5, 1, 2, 5, 10, 20, 40 and 100 hPa and every 3.75° in latitude. A reference arrow is shown in the top left of each plot along with its (F^{ϕ}, F^{z}) values. Shading shows where D_F differences are statistically significant at the 95% level.

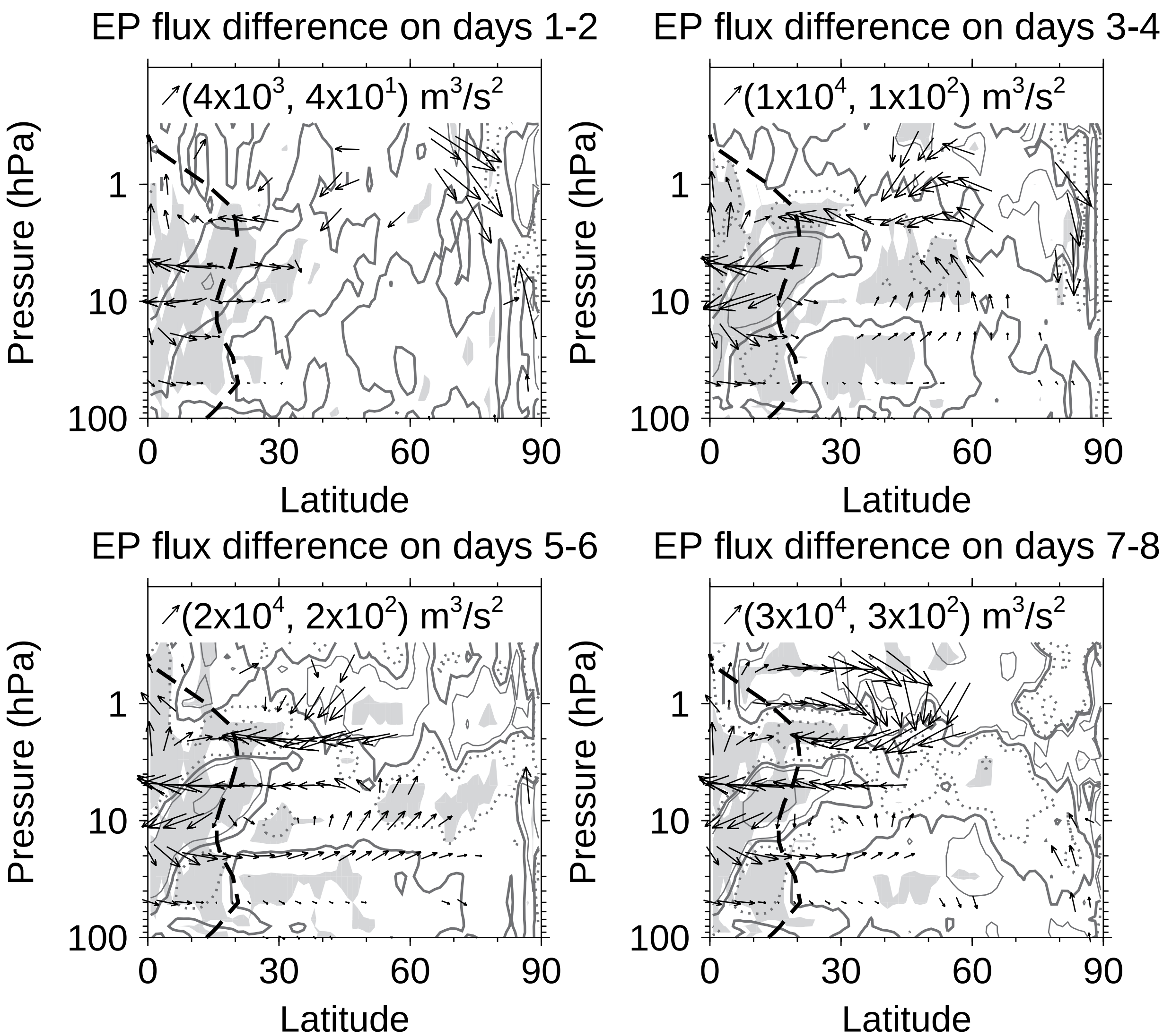

FIG. 11. As in figure 10 but using "acceleration scaling" of the EP flux vectors as defined by Gray et al. (2003), which indicates the zonal acceleration associated with the flux. A reference arrow is shown in the top left of each plot along with the (F^{ϕ}, F^{z}) values it would have at the Equator at 10 hPa.