On the Control of the Residual Circulation and Stratospheric Temperatures in the Arctic by Planetary Wave Coupling

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ABSTRACT

It is well established that interannual variability of eddy (meridional) heat flux near the tropopause controls the variability of Arctic lower-stratospheric temperatures during spring via a modification of the strength of the residual circulation. While most studies focus on the role of anomalous heat flux values, here the impact of total (climatology plus anomaly) negative heat flux events on the Arctic stratosphere is investigated. Utilizing the Interim ECMWF Re-Analysis (ERA-Interim) dataset, it is found that total negative heat flux events coincide with a transient reversal of the residual circulation and cooling of the Arctic lower stratosphere. The negative events weaken the seasonally averaged adiabatic warming.

The analysis provides a new interpretation of the winters of 1997 and 2011, which are known to have the lowest March Arctic lower-stratospheric temperatures in the satellite era. While most winters involve positive and negative heat flux extremes, the winters of 1997 and 2011 are unique in that they only involved extreme negative events. This behavior contributed to the weakest adiabatic downwelling in the satellite era and suggests a dynamical contribution to the extremely low temperatures during those winters that could not be accounted for by diabatic processes alone. While it is well established that dynamical processes contribute to the occurrence of stratospheric sudden warming events via extreme positive heat flux events, the results show that dynamical processes also contribute to cold winters with subsequent impact on Arctic ozone loss. The results highlight the importance of interpreting stratospheric temperatures in the Arctic in the context of the dynamical regime with which they are associated.

1. Introduction

During Northern Hemisphere (NH) winter, the stratosphere deviates significantly from radiative equilibrium because of the interaction of the stratospheric zonal-mean flow and planetary-scale waves, which propagate upward from the troposphere. The convergence of Eliassen–Palm (EP) flux by planetary-scale waves drives the equator-to-pole residual circulation that produces upwelling in the tropics and downwelling in high latitudes (Dunkerton et al. 1981; McIntyre and Palmer 1983; Andrews et al. 1987). The downward motion adiabatically warms the Arctic lower stratosphere and is opposed by radiative cooling.

In the NH the variability about the climatological mean state is large and coincides with variability in the upward propagation of planetary-scale waves from the troposphere. Newman et al. (2001) showed that interannual variability of eddy (meridional) heat flux near the tropopause controls the variability of the Arctic lower-stratospheric temperatures during spring. In particular, they showed that the 45°–75°N-averaged, 100-hPa eddy heat flux averaged from 15 January to 28 February is significantly correlated with the polar cap (60°–90°N)–averaged, 50-hPa zonal-mean temperature from 1 to 15 March.

Stratospheric sudden warming events are closely related to extreme time-integrated positive eddy heat flux

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events (Polvani and Waugh 2004). According to linear theory, such events imply enhanced upward wave propagation since the heat flux is an indicator of the direction of vertical wave propagation. Shaw and Perlwitz (2013) recently characterized the life cycle of extreme total (climatology plus anomaly) negative wave-1 heat flux events in the stratosphere. These events are associated with an eastward-wave phase tilt with height, which supports the interpretation from linear theory that the events are due to wave reflection, and a divergence of Eliassen–Palm flux in the Arctic lower stratosphere. So far, however, the impact of negative heat flux events on the residual circulation and temperatures in the Arctic has not been explored in detail.

Recent cold winters in the NH lower stratosphere and their connection with ozone loss have prompted a significant amount of research (e.g., Pawson and Naujokat 1999; Rex et al. 2004, 2006; Randel et al. 2009). During March 2011, record ozone loss was observed in the NH (Manney et al. 2011) together with extremely low Arctic lower-stratospheric temperatures. The March 2011 Arctic lower-stratospheric temperature was approximately 10 K below the climatological mean (Hurwitz et al. 2011). Hurwitz et al. (2011) showed that during February 2011, there was weak planetary wave driving of the stratosphere, which contributed to low March temperatures as expected from the Newman et al. (2001) relationship. The late winter period of 1997 was characterized by similar conditions [see Figs. 1a and 1b of Hurwitz et al. (2011)].

Weak planetary wave driving, as expressed by small time-integrated eddy heat flux values during a winter season, could arise from an enhanced number of negative heat flux events, which have a life cycle of approximately 20 days (Shaw and Perlwitz 2013), or from anomalously low positive heat flux values. Determining the connection between extreme eddy heat flux conditions and Arctic temperatures is key to improving our understanding of the link between stratospheric dynamics and ozone loss both in the real atmosphere and in stratosphere-resolving chemistry–climate models, which tend to underestimate temperature variability in the Arctic stratosphere (Hitchcock et al. 2009; Butchart et al. 2010).

Here we investigate the impact of total negative heat flux events on the residual circulation and Arctic lower-stratospheric temperatures during late winter using reanalysis data. We show that total negative heat flux events coincide with a transient reversal of the residual circulation leading to a weakening of the seasonally averaged adiabatic warming and thus cooling of the Arctic lower stratosphere. We subsequently highlight the role of negative heat flux events in the low temperatures observed during the late winter period of 1997 and 2011. The paper is organized as follows. Section 2 describes the data and methods. The connection between extreme heat flux events, residual circulation, and Arctic lower-stratospheric temperatures on daily and seasonal time scales are discussed in section 3. The results are summarized and discussed in section 4.

2. Data and methods

To explore the connection between extreme eddy heat flux events, the residual circulation, and temperatures in the Arctic lower stratosphere, we use the daily three-dimensional vertical and meridional wind and temperature from the Interim European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-Interim) dataset from 1979 to 2012 (Dee et al. 2011). We focus on the late winter period from 15 January to 15 March because planetary wave coupling is known to peak in late winter (Shaw et al. 2010; Perlwitz and Harnik 2003) and to be consistent with Newman et al. (2001). The results shown below were robust to extending the winter period back to 15 December. We consider total fields (not anomalies about the climatological seasonal cycle) to objectively identify negative heat flux events. We also found that the results based on ERA-Interim data are in good agreement with results from the Modern-Era Retrospective Analysis for Research and Application (MERRA) dataset (not shown).

The eddy heat flux, the residual circulation, and temperature are coupled via the thermodynamic energy equation. Following previous studies (e.g., Newman et al. 2001; Ueyama et al. 2013), we consider an approximation of the transformed Eulerian mean (TEM) thermodynamic energy equation in log-pressure coordinates on the sphere; for example,

$$\frac{\partial\overline{\theta}}{\partial t} + \overline{w}^*\frac{\partial\overline{\theta}}{\partial z} \approx \overline{Q}, \tag{1}$$

where $\overline{\theta}$ is the zonal-mean potential temperature, $\overline{w}^*$ is the vertical component of the residual circulation, $\overline{Q}$ is an effective diabatic source, and the remaining symbols have their usual meanings. The effective diabatic source term is defined as a residual and according to the full TEM thermal energy equation [see Eq. (3.5.2e) of Andrews et al. (1987)], includes both radiation and additional eddy terms. Recall that $\overline{w}^* = \overline{w} + \partial_\phi(\cos\phi\overline{v'\theta'}/\overline{\theta}_z)/a\cos\phi$ and hence includes a zonal-mean term and a term related to the meridional divergence of the eddy meridional potential temperature flux, which we will refer to as the eddy (sensible) heat flux. Here we focus on the impact of total negative eddy heat flux

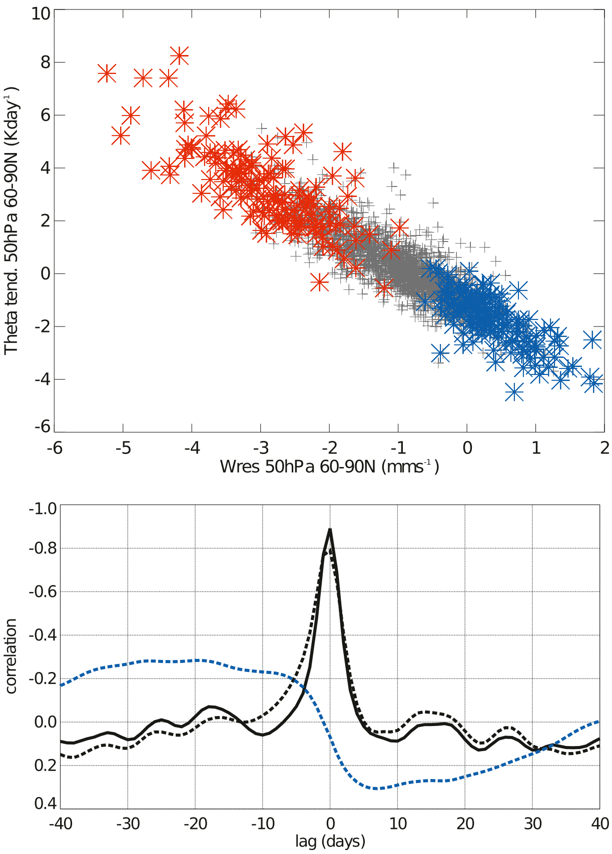

FIG. 1. (top) Scatterplot of the daily 15 Jan–15 Mar values of the potential temperature tendency vs $\overline{w}^*$ both averaged from 60° to 90°N at 50 hPa from 1979 to 2012. Red (blue) stars indicate extreme positive (negative) $\overline{v'\theta'}$ days (see text for definition) at 60°N. (bottom) Lag correlation of $\overline{w}^*$ and the potential temperature tendency at 50 hPa, both averaged from 60° to 90°N (black solid), of $\overline{v'\theta'}$ at 60°N and $\overline{w}^*$ at 50 hPa averaged from 60° to 90°N (black dashed), and $\overline{v'\theta'}$ at 100 hPa averaged from 45° to 75°N and the zonal-mean potential temperature at 50 hPa averaged from 60° to 90°N (blue dashed).

events on the residual circulation and temperatures in the Arctic lower stratosphere. In the analysis that follows we will refer to a polar cap average as a cosine-weighted average between 60° and 90°N.

3. Results

On a daily time scale the polar cap–averaged potential temperature tendency and $\overline{w}^*$ in the TEM thermal energy equation [see Eq. (1)] are significantly correlated (Fig. 1, top). The temporal correlation coefficient is -0.89 at zero-day lag with weaker values at positive and negative time lags (Fig. 1, bottom, black solid). The high correlation value suggests that the potential temperature tendency over the polar cap is directly related to the residual circulation. In particular, downward advection is associated with transient warming whereas upward advection is associated with transient cooling. Recall that the vertical residual velocity involves the sum of $\overline{w}$ and $\partial_\phi(\cos\phi\overline{v'\theta'}/\overline{\theta}_z)/a\cos\phi$. Consistently, the potential temperature tendency is significantly correlated with those individual terms and the lag-zero correlations with $\overline{w}$ and $\partial_\phi(\cos\phi\overline{v'\theta'}/\overline{\theta}_z)/a\cos\phi$ are 0.58 and -0.81, respectively (not shown).

Via a meridional integral, $\overline{w}^*$ can be directly connected to $\overline{v'\theta'}$. In particular, the polar cap–averaged $\overline{w}^*$ is by definition equal to the sum of the polar cap–averaged $\overline{w}$ and $(\cos\phi\overline{v'\theta'}/\overline{\theta}_z)/a$ at 60°N. The correlation coefficient between the $\overline{w}^*$ and $\overline{v'\theta'}$ at 60°N for lag zero is -0.86 (Fig. 1, bottom, black dashed), which is consistent with Newman et al. (2001) who noted a correlation coefficient of 0.92 using National Centers for Environmental Prediction (NCEP) data from 1979 to 1998.

Newman et al. (2001) showed that the 1 January–28 February, 45°–75°N-averaged 100-hPa eddy heat flux is significantly correlated with the 1–15 March polar cap–averaged 50-hPa temperature. Consistent with Newman et al. (2001), the time-lagged correlations between 45° and 75°N averaged 100-hPa $\overline{v'\theta'}$ and polar cap–averaged potential temperature maximize for positive lags from 5 to 38 days (Fig. 1, bottom, blue dashed). However, the maximum correlation values are much weaker than for the 50-hPa polar cap–averaged potential temperature tendency and polar cap–averaged $\overline{w}^*$ (blue dashed versus black solid). The results suggest that while the eddy heat flux is related to time-lagged polar cap temperature in the Arctic stratosphere, the daily evolution of $\overline{v'\theta'}$ is more closely related to the temperature tendency and implies that $\overline{v'\theta'}$ controls the temperature change over the late winter period.

The connection between the polar cap–averaged $\overline{w}^*$ and $\overline{v'\theta'}$ at 60°N also holds for days where $\overline{v'\theta'}$ values are extreme. Here, negative extremes are defined by total negative $\overline{v'\theta'}$ values at 60°N, which represent the 7th percentile of the daily $\overline{v'\theta'}$ distribution. The positive extremes are subsequently defined by a similar percentile (e.g., 93rd percentile) to ensure the same number of extreme positive and negative days. In the scatter diagram (Fig. 1, top) extreme positive heat flux days are labeled with red stars (141 days) whereas extreme negative heat flux days are labeled with blue stars (142 days). The extreme negative days account for approximately 60% of the days with upward advection by the residual circulation and negative potential temperature tendency.

The nature of the extreme events can be illustrated by comparing composites of days with extreme heat flux values to the climatology. In this case we consider when the polar cap–averaged $\overline{v'\theta'}$ is extreme to avoid using one latitude, consistent with Newman et al. (2001). (Note the correlation between the 60°N and the 60°–90°N-averaged flux is 0.83.) Figure 2 shows the 15 January–15 March climatology (top) of $\overline{v'\theta'}$, zonal-mean residual circulation and potential temperature tendency, and composites for days with extreme negative (middle) and positive (bottom) eddy heat flux values. The climatology involves positive (poleward) $\overline{v'\theta'}$, consistent with upward wave propagation, equator-to-pole residual circulation with downwelling in the Arctic, and negligible potential temperature tendency (Fig. 2, top). The poleward residual circulation is driven by EP flux convergence (Fig. 3, top) via a balance with the Coriolis force. The EP flux convergence results from the dominance of the vertical convergence of vertical EP flux over the meridional divergence of the meridional EP flux.

During extreme negative $\overline{v'\theta'}$ days, the high-latitude residual circulation is reversed from its climatological direction (Fig. 2, middle). This behavior is consistent with EP flux divergence in the Arctic stratosphere (Fig. 3, middle), which is dominated by the vertical divergence of the vertical EP flux. There is negative (equatorward) $\overline{v'\theta'}$ consistent with downward wave propagation. The potential temperature tendency is negative in the Arctic stratosphere and positive in midlatitudes consistent with air being advected upward over the pole producing adiabatic cooling and advected downward in midlatitudes producing warming. The residual circulation is acting like a refrigerator, transporting air from cold polar regions to relatively warm midlatitudes regions. Finally, during extreme positive days, the residual circulation is very strong and there is enhanced downwelling in high latitudes (Fig. 2, bottom). The enhanced downwelling is consistent with enhanced EP flux convergence in the stratosphere (Fig. 3, bottom), which reflects an amplification of the climatological conditions. The potential temperature tendency is positive and largely opposite to that during extreme negative heat flux events.

a. Interannual variability

Our analysis reveals that negative eddy heat flux events have a strong impact on the residual circulation on short time scales. In particular, transient dynamical cooling disrupts the climatological-mean adiabatic warming associated with downwelling. The reversal of the high-latitude residual circulation is as extreme as the transient reversal of the zonal-mean zonal wind during stratospheric sudden warming events that are closely related to extreme positive heat flux events. Over a winter season, the temperature of the Arctic lower stratosphere depends on the time-integrated potential temperature tendency, which includes the effects of the residual circulation and diabatic processes [see Eq. (1)]. As discussed in the introduction there has been significant interest in winters that exhibit low temperatures in the Arctic stratosphere and their connection to Arctic ozone loss. Cold winters are conventionally thought to be associated with weak wave driving and weak dynamical warming, following the Newman et al. (2001) relationship. Here we explore the impact of total negative eddy heat flux events on the interannual time scale by examining the distribution of individual winters.

Figure 4 shows the distribution of daily 15 January–15 March polar cap–averaged $\overline{w}^*$ values (top) and $\partial_\phi(\cos\phi\overline{v'\theta'}/\overline{\theta}_z)/a\cos\phi$ (bottom) at 50 hPa from 1979 to 2012. The $\overline{w}^*$ values associated with extreme positive and negative heat flux events are indicated by red and blue stars, respectively [the stars indicate the same events as in Fig. 1 (top)]. The solid horizontal black lines indicate plus or minus one standard deviation about the mean value for the entire time series. On average, $\overline{w}^*$ is negative, indicating the well-known climatological downward motion of the residual circulation over the Arctic. The downward $\overline{w}^*$ is primarily due to the meridional convergence of $\overline{v'\theta'}$, which dominates over $\overline{w}$. The $\overline{w}$ contribution to $\overline{w}^*$ is upward on average, as can be inferred from the difference between the top and bottom panels in Fig. 4. Daily values of $\overline{w}^*$ that are slightly larger than one standard deviation (upper solid line) are in general positive and involve upward motion over the Arctic. These extreme conditions typically coincide with meridional divergence of $\overline{v'\theta'}$ (Fig. 4, bottom) and total negative $\overline{v'\theta'}$ values (blue stars). Note that upward $\overline{w}^*$ can also occur because of $\overline{w}$ but here we specifically focus on the role of the $\overline{v'\theta'}$.

Figure 4 also reveals that the majority of winters involve both positive and negative heat flux extremes. However, there are individual winters that can be characterized as involving only extreme positive (e.g., 1991, 2009) or only extreme negative (e.g., 1997, 2011) events. The latter winters are expected to have a very weak residual circulation and low temperatures resulting from a significant number of days with a reversed residual circulation and negative potential temperature tendency as shown in Fig. 2 (middle).

To understand the processes responsible for the late winter conditions that lead to low temperatures and potentially to ozone loss we integrate the thermal energy equation from 15 January to 15 March. Figure 5 shows the yearly values of the time-integrated vertical potential temperature advection by the residual circulation (red), potential temperature tendency (black), and the effective diabatic term (blue) from Eq. (1). On average, from 15 January to 15 March, the vertical potential temperature advection by the residual circulation

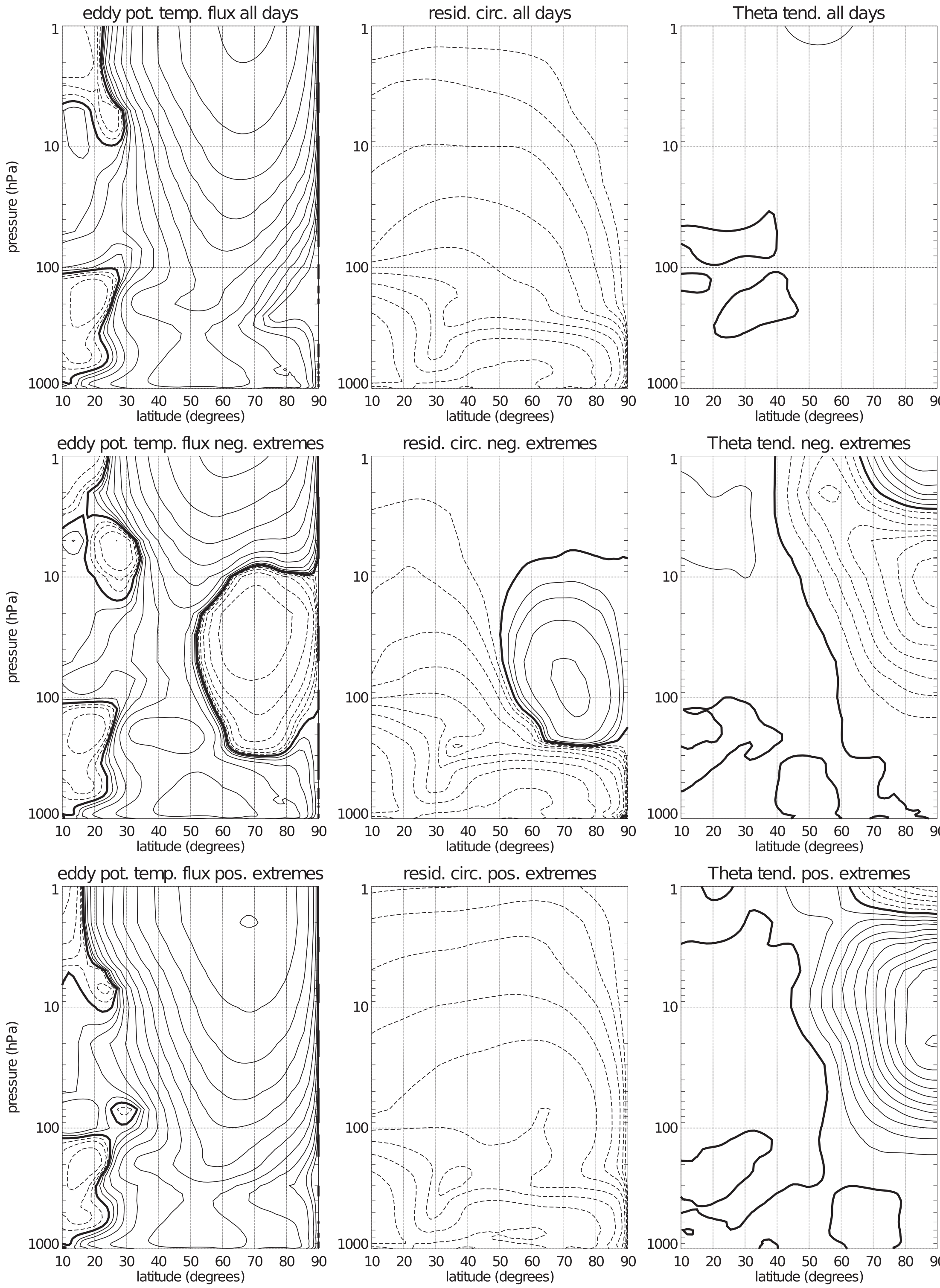

FIG. 2. (top) The climatological 15 Jan–15 Mar (left) $\overline{v'\theta'}$, (middle) residual circulation, and (right) potential temperature tendency. (middle) As in (top), but for extreme negative $\overline{v'\theta'}$ days. (bottom) As in (top), but for extreme positive $\overline{v'\theta'}$ days. See text for definition of negative and positive extremes. Contours are (left) $\pm[0.5, 1, 2, 4, 8, 16, 32, 64, \ldots]$ K m s^{-1} and (middle) $\pm(1 \times 10^9 \times [1, 2, 4, 8, 16, 32, 64, \ldots])$ kg s^{-1}, and (right) at an interval of ±1 K day^{-1}. Negative contours are dashed and zero contour is thick.

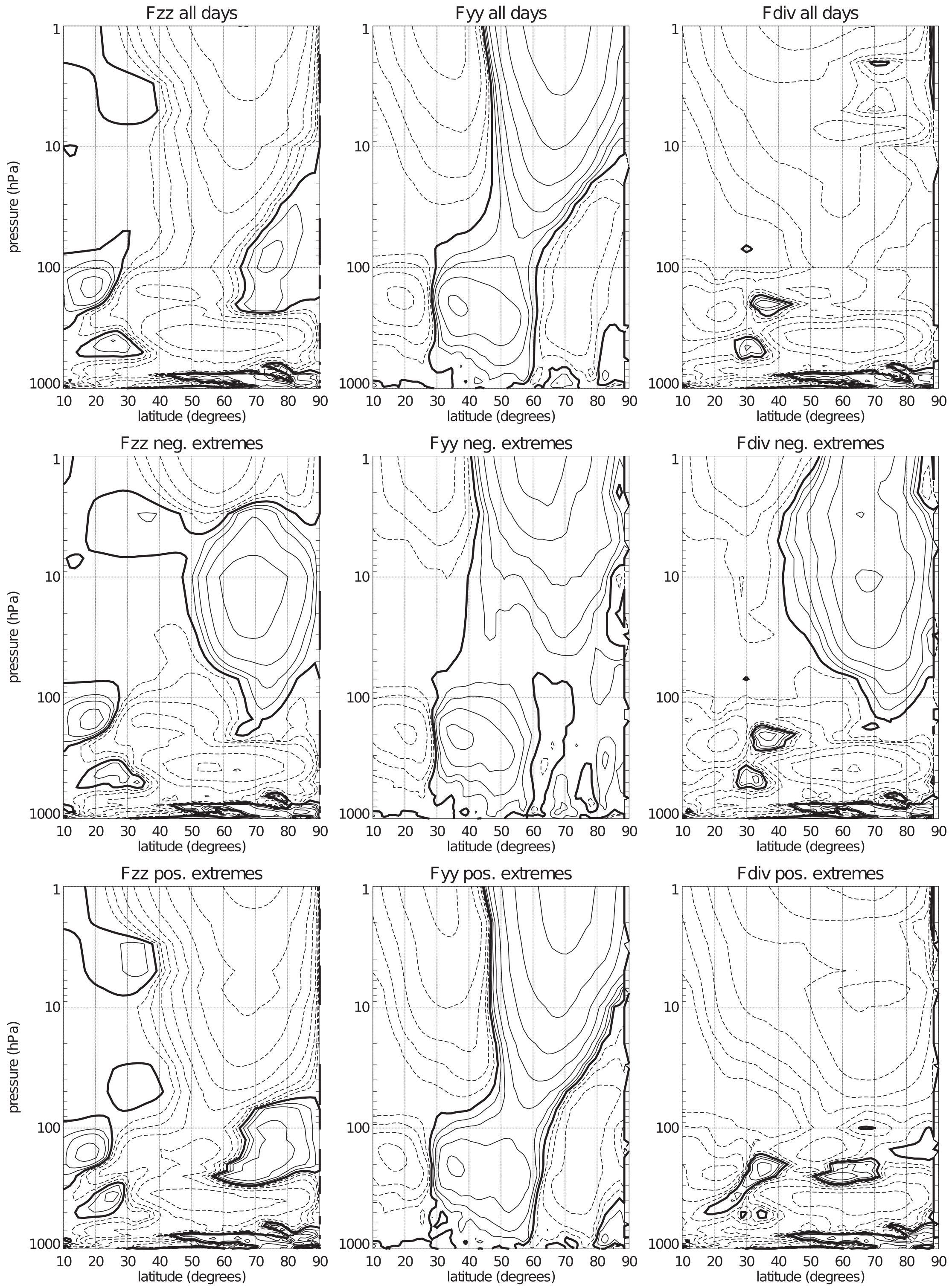

FIG. 3. (top) The climatological 15 Jan–15 Mar (left) vertical divergence of vertical EP flux, (middle) meridional divergence of the meridional EP flux, and (right) EP flux divergence. (middle) As in (top), but for extreme negative $\overline{v'\theta'}$ days (see text for definition). (bottom) As in (top), but for extreme positive $\overline{v'\theta'}$ days (see text for definition). Contours are $\pm[0.5, 1, 2, 4, 8, 16, 32, 64, \ldots]$ m s^{-1} day^{-1}. Negative contours are dashed and zero contour is thick.

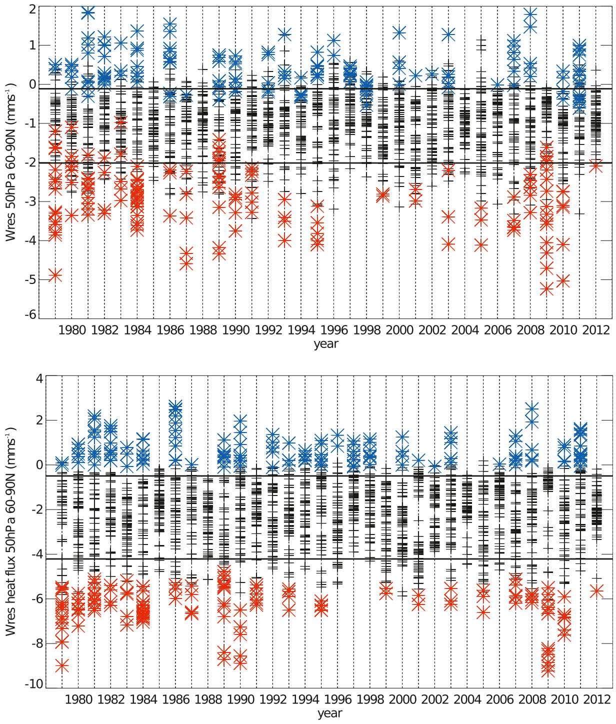

FIG. 4. Yearly distributions of the 15 Jan–15 Mar (top) $\overline{w}^*$ and (bottom) $\partial_\phi(\cos\phi\overline{v'\theta'}/\overline{\theta}_z)/a\cos\phi$ averaged from 60° to 90°N at 50 hPa from 1979 to 2012. Colors are as in Fig. 1 (top). The horizontal black lines indicate plus or minus one standard deviation about the mean value for the entire time series.

produces warming of the Arctic lower stratosphere and is opposed by cooling due to the effective diabatic term. The impact of having only negative extreme heat flux events during an individual winter can be seen in the time-integrated vertical advection by the residual circulation, which exhibited the lowest values in the satellite era during 1997 and 2011. Note that the effective diabatic term was also anomalous during 1997 and 2011. However, anomalous values also occurred in other years such as 1996 and 2000, suggesting diabatic processes alone are not responsible for the anomalous late-winter temperatures during 1997 and 2011.

The final contribution to the temperature changes is the time-integrated potential temperature tendency (Fig. 5, black bars), which is weakly positive on average. During 1997 and 2011, the integrated potential temperature tendency is negative, which is very anomalous. There exist other years when the tendency is negative (e.g., 2002, 2004, and 2006). It can be shown, however, that these winters are associated with different preconditioning. Figure 6 shows the 15 January–15 March time-integrated potential temperature tendency versus the potential temperature on 15 January, which are significantly correlated with a correlation coefficient of

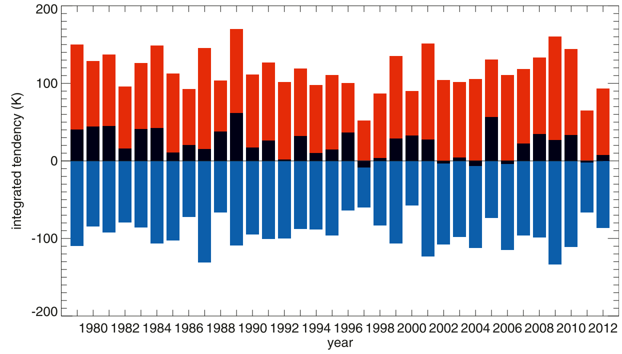

FIG. 5. Yearly values of the 50-hPa, 60°–90°N-averaged, and 15 Jan–15 Mar time-integrated TEM thermal energy equation [see Eq. (1)]: potential temperature tendency (black), vertical advection by the residual circulation (red), and effective diabatic term (blue).

−0.81. The 2002, 2004, and 2006 winters were very warm in early January owing to stratospheric sudden warmings [see Table 1 in Cohen and Jones (2011)], suggesting that the conditions were very disturbed and that the subsequent cooling during late winter period was primarily associated with radiative relaxation. However, the 15 January conditions during 1997 and 2011 (highlighted in blue) were close to the mean and subsequently cooled from 15 January to 15 March. Thus, these two years are clear outliers in the otherwise close relationship between time-integrated potential temperature tendency and the potential temperature on 15 January. This suggests that the winters of 1997 and 2011 represent a very different dynamical regime where negative $\overline{v'\theta'}$ events play an important role in the temperature change over the late winter season.

b. *The winters of 1997 and 2011*

In the following we study the winters of 1997 and 2011 in more detail. They are particularly interesting because they exhibited the lowest March polar cap–averaged lower-stratospheric temperatures in the Arctic during the satellite era [see Figs. 1 and 2 of Hurwitz et al. (2011)]. The anomalous conditions during late winter of 1997 and 2011 can be seen in the daily evolution of $\overline{v'\theta'}$ at 60°N, the polar cap–averaged $\overline{w}^*$ at 50 hPa, and polar cap–averaged geopotential height at 10 hPa (Fig. 7). The polar cap–averaged geopotential height is a proxy for the northern annular mode—for example, the state of the polar vortex (Baldwin and Thompson 2009). The evolution of $\overline{v'\theta'}$ during both winters (top, dashed) is mostly below the climatological average (top, solid). Consistently, the evolution of $\overline{w}^*$ during both winters (middle, dashed) is mostly above the climatological average (middle, solid) and the zonal-mean geopotential height (bottom, dashed) is mostly below the

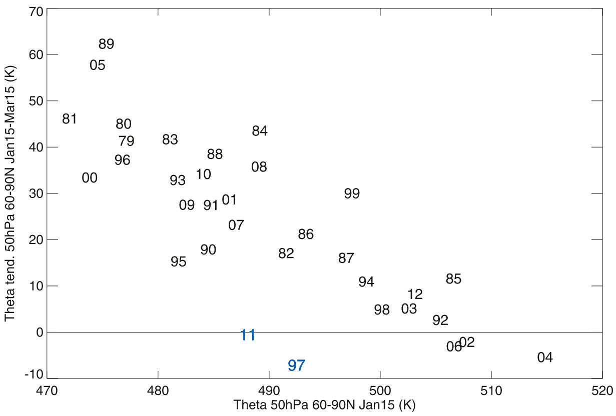

FIG. 6. The time-integrated potential temperature tendency from 15 Jan to 15 Mar vs the potential temperature on 15 Jan, both averaged from 60° to 90°N at 50 hPa. Years 1997 and 2011 are indicated in blue.

climatological average (bottom, solid), indicating an anomalously strong vortex. During 2011 (1997), 23% (13%) of the days between 15 January and 15 March involved negative $\overline{v'\theta'}$ and positive $\overline{w}^*$. This implies that many days during both years exhibited the behavior seen in the composite of extreme negative days, including a reversed residual circulation and a dynamically driven negative potential temperature tendency. Note that the negative heat flux events coincide with a strengthening of the polar vortex consistent with the life cycle analysis of Shaw and Perlwitz (2013).

The impact of the anomalous $\overline{v'\theta'}$ during 1997 and 2011 on the overall character of these winters, including the zonal-mean $\overline{v'\theta'}$, residual circulation, and zonal wind becomes obvious in Fig. 8. The shading indicates the region where the quantity is outside one standard deviation of the yearly time series. When compared to the climatological-mean $\overline{v'\theta'}$ (see Fig. 2), the midlatitude $\overline{v'\theta'}$ is weak, consistent with Newman et al. (2001) and Hurwitz et al. (2011), but the high-latitude heat flux is close to the mean. Note that the behavior in high latitudes involves an average over negative and positive heat flux events (see Fig. 7). The residual circulation in the Arctic stratosphere is close to zero, implying a lack of wave-driven downwelling, which dominates the climatology. Recall that the 1997 and 2011 winter seasons have the smallest time-integrated vertical advection by the residual circulation over the Arctic in the observational record (see red bars in Fig. 5). As a result of the weak residual circulation, the Arctic temperature is below climatology (not shown) and the stratospheric polar vortex is anomalously strong (Fig. 8, right) during both years. In addition, the polar vortex exhibits a maximum in the midstratosphere with negative vertical zonal wind shear in the upper stratosphere, which is known to create a reflecting surface (region of zero squared vertical wavelength) that is favorable for wave reflection (Perlwitz and Harnik 2003; Shaw et al. 2010). The structure of the polar vortex during the winters of 1997 and 2011 is consistent with the reflective configuration of the polar vortex discussed by Perlwitz and Harnik (2004) that favors wave reflection over wave absorption.

Overall, the winters of 1997 and 2011 are unique for the following reasons. First, as discussed previously, these winters belong to a limited set that have no positive extreme heat flux events (no red stars in Fig. 4). Second, vertical advection for nonextreme days (black plus signs in Fig. 4) was anomalously weak (greater than one standard deviation); note, however, that it was also low during 1984 and 2009 (not shown) when extreme low temperatures were not observed, suggesting that this is not a sufficient condition. Third, during the winters of 1997 and 2011, the number of extreme negative heat flux days was anomalously high as discussed above. The enhanced occurrence of extreme negative heat flux events in the absence of extreme positive events contributes to an anomalously weak time-integrated heat flux. Finally, consistent with the dominance of negative heat flux extremes, the time-integrated potential temperature tendency during these winters was negative and the vertical advection by the residual circulation over the Arctic was the lowest in the observational record (see Fig. 5).

4. Summary and discussion

The impact of total (climatology plus anomaly) negative eddy heat flux events on Arctic lower-stratospheric temperatures and the residual circulation is analyzed. The study is motivated by Shaw and Perlwitz (2013), who isolated the dynamical impacts of total (climatology plus anomaly) negative wave-1 heat flux events and showed that they were largely consistent with wave reflection in the stratosphere, although other processes could also be involved (e.g., stationary–transient interactions, resonance, instability). The goal of the present study has been to better understand the thermodynamic impact of total negative heat flux events, including their potential role in recent cold winters in the Arctic stratosphere.

It is well known that the variability of the time-integrated eddy (meridional) heat flux controls the temperature variability of the Arctic lower stratosphere (Newman et al. 2001). Here we have shown that the daily high-latitude eddy heat flux controls the vertical advection of air by the residual circulation over the Arctic. More specifically, we have found that on daily

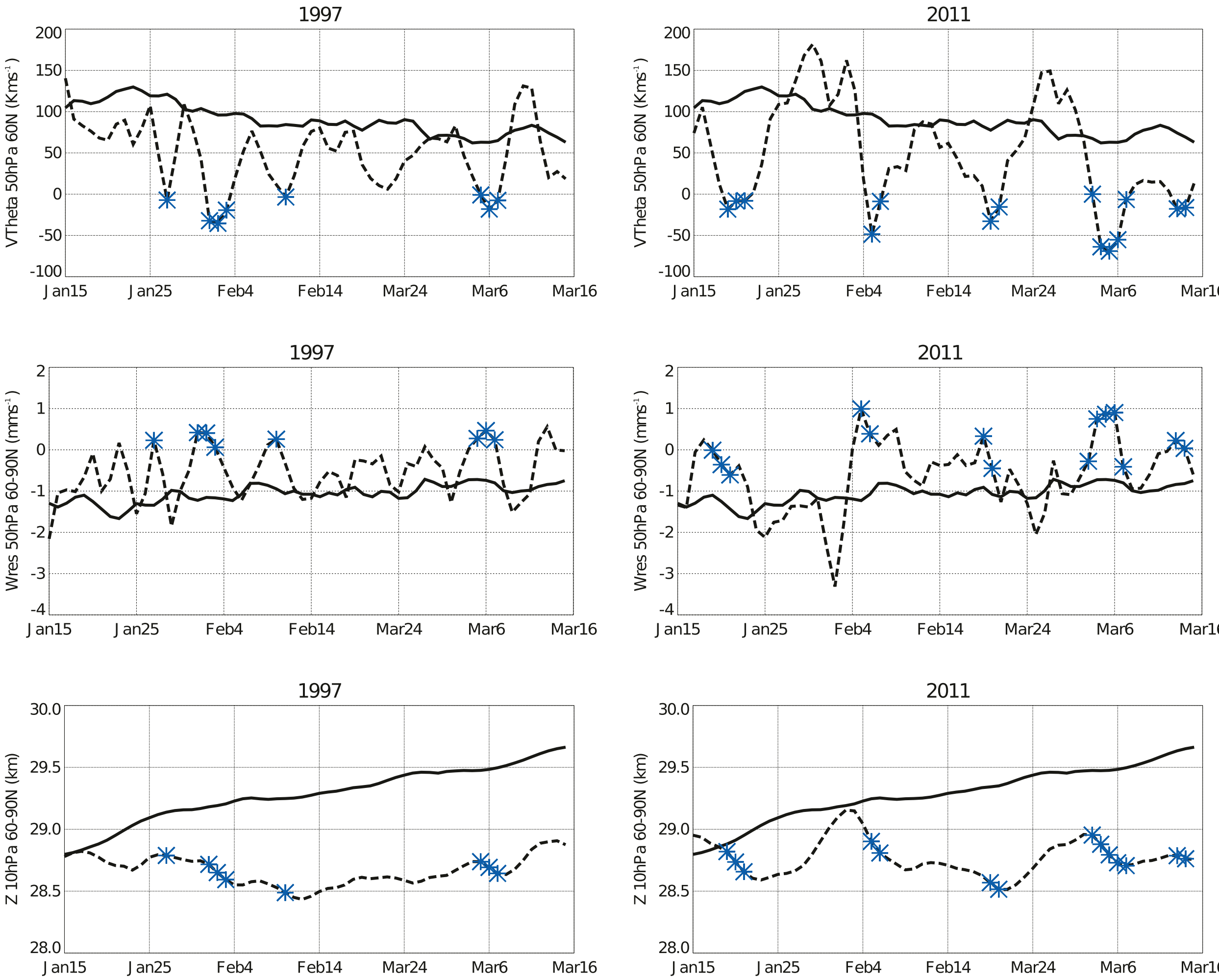

FIG. 7. (top) Daily evolution of $\overline{v'\theta'}$ at 60°N and 50 hPa, (middle) polar cap–averaged $\overline{w}^*$ at 50 hPa, and (bottom) polar cap–averaged geopotential height at 10 hPa for (left) 1997 (dashed), (right) 2011 (dashed), and the climatological evolution from 1979 to 2012 (solid).

time scales the heat flux is highly correlated with the potential temperature tendency over the Arctic such that during extreme negative eddy heat flux events, there is transient dynamical cooling balanced by upward advection by the residual circulation. The extreme positive eddy heat flux events are associated with transient dynamical warming and downward advection. Ueyama et al. (2013) noted that short-time-scale positive and negative eddy heat flux anomalies are associated with anomalous transient warming and cooling, respectively. Here we have shown in absolute terms that transient cooling is associated with a total negative eddy heat flux.

Extreme eddy heat flux events have a strong impact on the structure of the residual circulation and potential temperature tendency. We have shown that during extreme negative events, the eddy heat flux is equatorward and thus reversed from its climatological mean state. According to linear theory, the reversal in sign of the eddy heat flux indicates a reversal of the direction of wave propagation, which is normally upward. It is known that extreme positive events are associated with a reversal of the zonal-mean wind—for example, a stratospheric sudden warming. Here we have shown that total negative heat flux events produce a reversal of the residual circulation. The reversed circulation is due to EP flux divergence and transports air from the Arctic lower stratosphere to midlatitudes, producing a significant cooling tendency throughout the Arctic stratosphere. This behavior is consistent with the circulation acting like a refrigerator in which air is transported from cold polar regions to relatively warm midlatitude regions. Individual winters have different combinations of positive and negative extreme events. Most winters involve both extreme positive and negative events; however, certain winters, such as 1997 and 2011, only involved

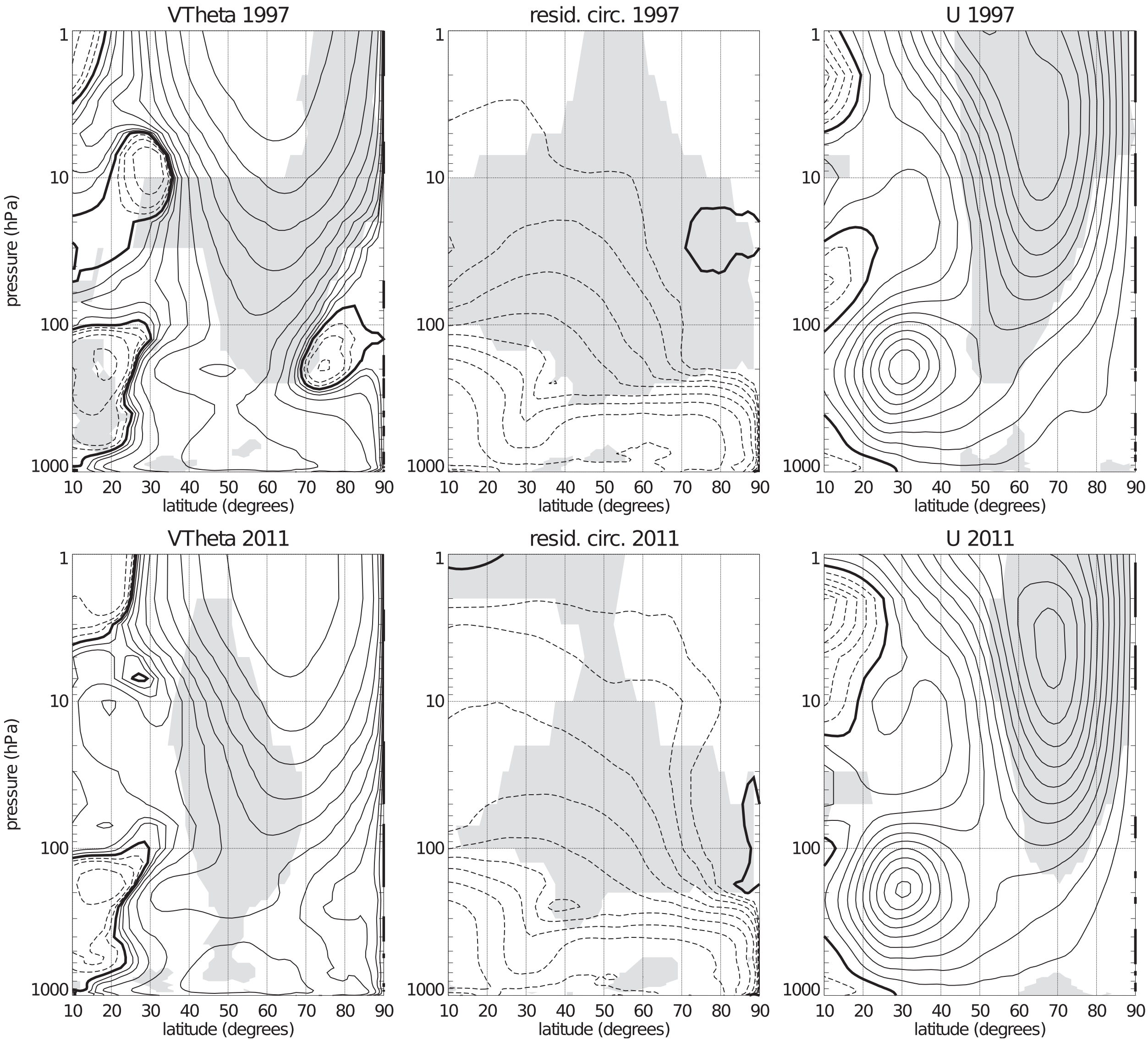

FIG. 8. The 15 Jan–15 Mar-averaged (left) eddy meridional heat flux, (middle) residual circulation, and (right) zonal-mean zonal wind during (top) 1997 and (bottom) 2011. Contours are (left) $\pm[0.5, 1, 2, 4, 8, 16, 32, 64, \ldots]$ K m s^{-1} and (middle) $\pm(1 \times 10^9 \times [1, 2, 4, 8, 16, 32, 64, \ldots])$ kg s^{-1}, and (right) at an interval of ± 5 m s^{-1}. Shading indicates where the field is (left) less than one standard deviation and (middle),(right) greater than one standard deviation. Negative contours are dashed and zero contour is thick.

extreme negative events. The occurrence of multiple negative extreme events during an individual season weakens the seasonally averaged adiabatic warming and produces an anomalously weak (but positive) or (even) negative time-integrated potential temperature tendency.

The results provide a new interpretation of the winters of 1997 and 2011, which have the lowest March Arctic polar cap–averaged temperature in the satellite era (Hurwitz et al. 2011). Previous work suggested that the extremely low temperatures occurred because of weak wave driving (Newman et al. 2001; Hurwitz et al. 2011), suggesting radiative rather than dynamical control. The new interpretation discussed here involves the occurrence of extreme eddy meridional heat flux events. The winters of 1997 and 2011 are anomalous because of enhanced occurrence of negative events representing the negative extreme tail of the distribution as well as because of the absence of extreme positive events. This behavior translates into the weakest dynamical warming in the satellite era and a negative potential temperature tendency over the late winter season that involves the episodic reversal of the residual circulation. We found that diabatic effects alone could not account for the extreme temperatures during late winter. The results establish that dynamical processes can contribute to cold winters via their impact on the residual circulation and temperature tendency.

A remaining question is the connection between vertical wave propagation, temperature, and stratospheric-ozone chemistry. Previous authors have noted dynamical contributions to anomalously low ozone in the Arctic (e.g., Fusco and Salby 1999; Randel and Wu 2002; Tegtmeier et al. 2008). Tegtmeier et al. (2008) showed that dynamical and chemical contributions are equally important in the Arctic. Heterogeneous chemical loss and a late final warming were the two major reasons for the low ozone during March 2011 (Manney et al. 2011; Strahan et al. 2013). A recent attribution study of the March 2011 Arctic ozone loss event by Isaksen et al. (2012) showed that chemical ozone loss accounted for 23% of the ozone anomaly whereas anomalous transport accounted for 76%. Note that the extreme negative eddy heat flux events observed during March 2011 could contribute to ozone loss via a weakening of the residual circulation, which leads to weakened transport, a lowering of Arctic temperatures, a strengthening of the polar vortex, and thus a delayed final warming. The detailed coupling between wave propagation and chemistry in the stratosphere requires further investigation. Applying the diagnostics used in this paper to chemistry–climate models in order to better understand their known temperature biases (Hitchcock et al. 2009; Butchart et al. 2010) is work in progress.

The present results highlight the impact of negative heat flux events on the residual circulation and Arctic stratospheric temperatures but did not address the conditions that lead to such events. Harnik (2009) suggested that short-time-scale positive heat flux pulses from the troposphere are more likely to lead to wave reflection. A better understanding of the tropospheric conditions that produce heat flux pulses is needed to improve understanding of winter variability in the Arctic stratosphere. It has been proposed that cold winters are getting colder because of the increase in greenhouse gas concentrations (Rex et al. 2004, 2006). However, the role of changes in the dynamical configuration of the atmosphere, which includes changes in tropospheric wave forcing and stratospheric basic state, are not well understood. The winters of 1997 and 2011 clearly represent a unique dynamical regime but do not constitute a trend. The current results highlight the importance of interpreting past and future climate change trends in wintertime Arctic temperatures and ozone in the context of the dynamical regime with which they are associated.

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