The Role of Inertial Instability in Cross-Hemispheric Coupling

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ABSTRACT: Recent studies suggest linkages between anomalously warm temperatures in the winter stratosphere, and the high-latitude summer mesopause. The summer temperature anomaly is manifested in the decline of polar mesospheric clouds. The 2-day wave is a strong-amplitude and transient summer feature that interacts with the background state so as to warm the high-latitude summer mesopause. This wave has been linked to a low-latitude phenomenon called inertial instability, which is organized by breaking planetary waves in the winter stratosphere. Hence, inertial instability has been identified as a possible nexus between the disturbed winter stratosphere, and summer mesopause warming. We investigate a sustained occurrence of inertial instability during 19 July–8 August 2014. During this period, stratospheric winter temperatures warmed by about 10 K, while a steep decline in polar mesospheric clouds was reported between 26 July and 6 August. We present, for the first time, wave driving associated with observed inertial instability. The effect of inertial instability is to export eastward momentum from the winter hemisphere across the equator into the summer hemisphere. Using a primitive equation model, we demonstrate that the wave stresses destabilize the stratopause summer easterly jet. The reconfigured wind profile excites the wavenumber-4 component of the 2-day wave, leading to enhanced warming of the summer mesopause. This work supports previous numerical investigations that identified planetary wave–driven inertial instability as a source of the 2-day wave.

KEYWORDS: Instability; Planetary waves

1. Introduction

Mesospheric variability and its relationship to sudden stratospheric warming has been the focus of numerous studies (Pancheva et al. 2006; Day et al. 2011; Zülicke and Becker 2013; Pedatella et al. 2014). These investigations arose in part from the desire to interpret interannual variations in polar mesospheric clouds (PMCs) (Siskind et al. 2011). PMCs (also called noctilucent clouds) are thin and ephemeral layers of ice water clouds that occur in very cold temperatures ($T < 150$ K) and at very high altitudes (~ 83 km). These conditions are found at the high-latitude summer mesopause, where equatorward and ascending motions associated with a gravity wave–driven global circulation result in the lowest temperatures in Earth's atmosphere (Garcia and Solomon 1985; Fleming et al. 1990). Because of their high sensitivity to upper-atmospheric water vapor and temperature (Thomas 1996), changes in PMC frequency and distribution are regarded as markers of mesospheric cooling trends, and possible global climate change (DeLand et al. 2007; Shettle et al. 2009). It is therefore important to sort out genuine long-term PMC changes from those related to aperiodic, year-to-year fluctuations.

Several studies indicate that PMC occurrence and brightness are anticorrelated with the temperature of the high-latitude winter stratosphere (Karlsson et al. 2007, 2009). Insofar as PMC occurrence is a proxy for mesopause temperature, and the stratospheric temperature is a measure of planetary wave (PW) activity, the reported correlation implies the imprinting of strong wintertime stratospheric PW activity upon the anomalously warm summer mesopause. Quite recently, Smith et al. (2020) argued that statistically, the summer mesosphere warming is an extension of the wave-driven winter hemisphere circulation that arises from global wind balance and mass continuity requirements. However, most theoretical investigations into this so-called interhemispheric coupling (IHC) mechanism have focused on the role of the gravity wave (GW)–driven circulation. Kőrnic and Becker (2010) and Karlsson and Becker (2016) attribute the summer mesopause warming to anomalies in the summer hemisphere circulation resulting from modified GW filtering in both hemispheres.

A different, and more direct mechanism for warming the cold summer mesopause involves the 2-day wave. This wave is a natural oscillation of Earth's atmosphere that amplifies in baroclinically and barotropically unstable regions of the middle atmosphere summer jets (Plumb 1983; Burks and Leovy 1986; Norton and Thuburn 1996; Gu et al. 2013). Momentum deposition to the mean flow by dissipating 2-day waves drives a hemispheric-scale circulation in the opposite sense to the GW-driven flow. Thus, the net effect of the 2-day wave is to warm the cold summer mesopause (Pendlebury 2012; France et al. 2018). Numerical experiments carried out by Pendlebury (2012) showed that 2-day wave forcing accounted for up to

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10% of interannual variability in polar summer mesopause temperatures in the Canadian Middle Atmosphere Model. Using a high-altitude forecast-assimilation system, [Siskind and McCormack \(2014\)](#) presented evidence of enhanced summertime 2-day wave activity during seasons of elevated winter PW activity and PMC decline.

[Orsolini et al. \(1997\)](#) examined the 2-day wave in the Met Office (UKMO) stratospheric analyses. They noted that at the stratopause, the wave coincided with occurrences of anomalous potential vorticity and inertially unstable structures in the PW breaking region (“surf” zone). Inertial instability flattens meridional shear near the equator, but intensifies the meridional curvature of the subtropical summer jet ([Hitchman et al. 1987](#)). Strong meridional curvature is a leading cause of barotropic instability ([Andrews et al. 1987](#)), which has been shown to be a source of the 2-day wave in the upper stratosphere ([Lieberman 1999](#)). [Orsolini et al. \(1997\)](#) therefore proposed that coupling of the 2-day wave to breaking PWs is facilitated by low-latitude inertial instability, which destabilizes the zonal wind profile in the upper stratosphere, thus creating a base state upon which the 2-day wave can grow. This mechanism was demonstrated in a mechanistic model by [Limpasuvan et al. \(2000\)](#).

The only known effort to investigate Orsolini’s and Limpasuvan’s mechanism observationally appears to be that of [McCormack et al. \(2009\)](#). These authors diagnosed inertial instability in a high-altitude numerical weather prediction model during a strong 2-day wave event in January 2006. They concluded that both PWs and inertially unstable structures acted to destabilize the zonal-mean zonal wind structure near the equatorial stratopause. This work, together with [Siskind and McCormack \(2014\)](#), presents prima facie evidence for linkages between 2-day wave intensity, PMC occurrence, and PW influences at low latitudes. However, these investigations all relied upon a limited collection of high-altitude analyses that, unfortunately, did not overlap with global observations of PMCs during years of strong-amplitude 2-day wave occurrence.

This paper reports the first observational confirmation of the IHC mechanism proposed by [Orsolini et al. \(1997\)](#) and modeled by [Limpasuvan et al. \(2000\)](#). Using satellite data, reanalysis, and output from a high-altitude forecast-assimilation system, we identify inertial instability during July–August 2014, when a strong 2-day wave was also present. We compute the zonal momentum force associated with inertial instability, and the response of the tropical zonal-mean zonal wind. We analyze the evolution of the observed zonal winds and 2-day wave spectrum, and isolate the sensitivity of the 2-day wave to tropical wind curvature. Our results show that inertial instability destabilizes the low-latitude summer easterly (westward) jet through transport of eastward momentum from the winter hemisphere across the equator at stratopause heights. This condition leads to amplification of the wavenumber-4 component of the 2-day wave, and subsequent high-latitude warming through its interaction with the mean flow.

Following a description of the data and modeling resources, [section 3](#) provides an overview of the phenomenon of inertial

instability, and an example of its occurrence in satellite data and Modern-Era Retrospective Analysis for Research and Applications (MERRA-2) reanalysis during July 2014. We also examine the interaction of inertial instability with the background wind. [Section 4](#) describes the concurrent evolution of the 2-day wave in reanalysis, and the Navy Global Environmental Model (NAVGEOM) high-altitude forecast-assimilation system. NAVGEOM products are then used to compute the response of the cold summer mesopause to the 2-day wave. In [section 5](#) we explore the sensitivity of the 2-day wave to inertial instability in a primitive equation model. Conclusions appear in [section 6](#).

2. Data and models

Our study focuses on the 17 July–10 August 2014 interval, when an anomalous decline in PMCs was observed by the Cloud Imaging and Particle Size (CIPS) instrument on the Aeronomy of Ice in the Mesosphere (AIM) satellite. CIPS is a suite of four nadir-pointing cameras that image the atmosphere’s scattered UV brightness between 258 and 274 nm ([McClintock et al. 2009](#)). The PMC deficit was traced to a strong 2-day wave that increased the zonally averaged temperature ($\bar{T}$) on the poleward flank of the wave ([France et al. 2018](#)). The 2-day wave amplification was associated with quasigeostrophic instability of the equatorward flank of a stronger-than-average mesosphere easterly jet.

[France et al. \(2018\)](#) attributed the jet instability to a GW-driven IHC mechanism. This judgment was made on the basis of zonal-mean zonal wind ($\bar{U}$) and $\bar{T}$ anomalies that were consistent with the IHC paradigm of [Karlsson et al. \(2009\)](#). No observations of anomalous GW variance or stresses were presented. The present study differs fundamentally from [France et al. \(2018\)](#) insofar as we examine a separate IHC process—based on inertial instability—using observations, and observation-driven modeling. Our goals are to measure the stress exerted by inertial instability on the tropical easterly jet, and the sensitivity of the 2-day wave to the altered $\bar{U}$.

Our first task is to seek evidence for inertial instability in July 2014. The primary dataset for its detection is version 2 Sounding of the Atmosphere using Broadband Emission Radiometry (SABER) temperatures. These are retrieved from infrared emissions of CO₂ from approximately 17 to about 120 km, with a precision of between 1 and 2 K for an individual temperature profile in the regions of interest (~45–65 km) ([Remsberg et al. 2008](#)). The intrinsic vertical resolution of the SABER temperature retrieval is about 2 km, and is suitable for detection of structures with vertical wavelengths longer than 4 km. SABER data are reported and displayed on scaled log-pressure surfaces, with a constant scale height value (H) of 6.5 km in the mesosphere.

To compute the zonal momentum force associated with inertial instability, we require self-consistent definitions of low-latitude temperature and wind. Owing to the difficulty of deriving winds near the equator, we turn to the MERRA-2 ([Gelaro et al. 2017](#)). We identify inertial instability from instantaneous fields of temperature, geopotential height, horizontal winds and Ertel potential vorticity (EPV; defined in

section 3), provided every 3 h on 42 pressure levels (1000–0.1 hPa) at $1.25^\circ \times 1.25^\circ$ longitude and latitude spacing. MERRA-2 analyses are also used to examine the evolution of the 2-day wave in the stratosphere and lower mesosphere. For the wave analyses, we use instantaneous fields provided every 6 h between 1000 and 0.015 hPa.

We use a nonlinear, time-dependent primitive equation model to examine the atmospheric response to tropical wind forcing induced by inertial instability (Ortland and Alexander 2014). The vertical domain of the Ortland model is represented on discrete log-pressure surfaces [$\log(p) = z/H$]. The upper boundary corresponds to roughly 300 km geometric height. Horizontal structure is represented by spherical harmonics with 40 basis functions in latitude. Damping mechanisms include a sponge layer that begins at 17 scale heights, Newtonian cooling, an eddy diffusion parameterization, molecular diffusion, and an ion drag represented by Rayleigh friction above 105 km. The cooling rate peaks at 0.25 s^{-1} at the stratopause, then drops to 0.05 s^{-1} above 65 km. The eddy diffusion coefficient increases from a value of $1 \text{ m}^2 \text{ s}^{-1}$ at 15 km to $100 \text{ m}^2 \text{ s}^{-1}$ at 95 km.

To study the wintertime PW evolution we use version 4.2 retrievals of temperature and geopotential height from the National Aeronautics and Space Administration (NASA) Aura Microwave Limb Sounder (MLS). Temperature is inferred from emissions by molecular oxygen at 118 GHz. In the mesosphere (between 0.01 and 0.001 hPa), temperature profiles have a vertical resolution of 8–13 km, a precision of 2.2–2.5 K, and a 4–8 K cold bias (Livesey et al. 2018). MLS geopotential height is derived from temperature and pressure and integrated from a reference geoid; it has a precision of 45 m at 1 hPa and 110 m at 0.01 hPa (Livesey et al. 2018). Daily profiles are interpolated to a 2.5° latitude $\times$ 3.75° longitude grid to produce daily mean gridded fields, with the ascending and descending nodes averaged together in order to remove a large fraction of the tidal effects. MLS geopotential height data are used to compute daily zonal-mean gradient zonal winds using Eq. (3.1) of Hitchman and Leovy (1986). Wave amplitudes are calculated by fitting a sine wave to daily MLS geopotential height data around each latitude circle on constant pressure surfaces.

To examine the effects of the 2-day wave on $\bar{U}$ and $\bar{T}$, we make use of a high-altitude (0–100 km) global reanalysis of the 2014 austral winter (April–October 2014) generated with a high-altitude research prototype of the NAVGEM. Eckermann et al. (2018) provide details of this NAVGEM configuration, the reanalysis experiments that were performed in 2014, and validation comparisons with other analyses and mesosphere and lower-thermosphere (MLT) observations. The NAVGEM reanalysis assimilates version 4.2 MLS temperatures, version 2.0 SABER temperature, and O_2 microwave radiances from the Special Sensor Microwave Imager/Sounder (SSMIS) on three separate satellites (F17, F18, and F19) of the Defense Meteorological Satellite Program.

NAVGEM fields are provided at a 1-h time cadence, resulting from 6-h analyses with forecasts in the intervening 5 h. Two-day wave fields are computed from space–time Fourier analyses of 6-day sliding windows. The 2-day wave is

identified as the sum of zonal wavenumbers 3 and 4 with a period of 2 days, and zonal wavenumber 4 with a period of 1.5 days.

3. Inertial instability during July 2014

Inertial instability occurs where the absolute vorticity is *anomalous*, or of opposite sign to the planetary vorticity f (defined as $2\Omega \sin\phi$). A simple, zonally averaged paradigm is presented in section 7.5 of Holton (2004), and section 13.2 of Boyd (2018). In a resting atmosphere, a north–south displacement of an air parcel on a constant pressure surface experiencing a restoring force proportional to the planetary vorticity f will return to its point of origin, tracing out a circular, anticyclonic trajectory. However, if a parcel embedded in a laterally sheared background wind $\bar{U}(y)$ is similarly displaced, it is restored by a force proportional to the *absolute* vorticity $f - \bar{U}_y$ (Holton 2004). If the magnitude of $\bar{U}_y$ exceeds that of f , the parcel does not return to its original position, but is embedded within an *inertially unstable* environment, and accelerates further away in latitude. This condition is often realized in narrow longitude bands on the winter side of the equator, where f is small, and breaking PWs generate strong positive gradients of $\bar{U}$ (Hitchman et al. 1987; Knox and Harvey 2005).

Dunkerton (1981) examined inertial instability numerically with respect to a weakly viscous, linearly sheared $\bar{U}(y)$ in the tropical middle atmosphere. Steady-state (i.e., marginally stable) solutions are manifested as coherent, vertically stacked thinly layered (“pancake”) structures, confined to the region of anomalous absolute vorticity (Dunkerton 1981; Hitchman et al. 1987). Several studies have documented features in planetary-scale (zonal wavenumbers 1–6) temperatures and trace gases that closely match the pancake structuring described by Dunkerton (1981): confinement to inertially unstable zones within 20° of the equator, and vertical wavelengths between 6 and 15 km (Fritts et al. 1992; Hitchman et al. 1987; Hayashi et al. 1998; Smith and Riese 1999; Rapp et al. 2018; Harvey and Knox 2019).

Figure 1 provides the context for the present investigation, as revealed in MLS geopotential heights, $\bar{T}$, and gradient $\bar{U}$. Strong PWs 1 and 2 appear in the Southern Hemisphere during July–August 2014. At 1 hPa (~ 48 km), peak PW 1 amplitudes occurred between 25 and 29 July, after which the values decreased. PW 2 attained peak amplitudes around 15 July and 3 August. An increase in $\bar{T}$ is observed poleward of 40°S during 17 July–2 August, accompanied by a modest weakening (20 m s^{-1}) of eastward $\bar{U}$. At low latitudes, PW1 amplification coincides with penetration of the summer hemisphere westward $\bar{U}$ across the equator, to 30°S .

Figure 2 shows the EPV derived from MERRA-2 corresponding to 48 km on 18 July 2014. EPV is a generalization of the barotropic absolute vorticity $f - \bar{U}_y$ to three dimensions. It is defined as $-g(\zeta_\theta + f)\partial\theta/\partial p$, where θ is the potential temperature, ζ_θ is the relative vorticity on an isentropic (constant θ) surface, and p is pressure (Holton 2004). Physically, EPV is a measure of the absolute circulation of an air parcel enclosed between two isentropic surfaces. This quantity is conserved for dry adiabatic, frictionless flows, and thus can be thought of as a

FIG. 1. (top) MLS geopotential (left) PW1 and (right) PW2 amplitudes at 1 hPa (approximately 48 km). (bottom left) $\bar{T}$ at 1 hPa. (bottom right) $\bar{U}$ at 1 hPa, calculated as the gradient wind.

dynamical tracer. Figure 2 shows that in the low-latitude winter hemisphere, narrow tongues of low (high) EPV are drawn poleward (equatorward), leading to strong spatial gradients in potential vorticity. Together, Figs. 1 and 2 suggest comingling of summer and winter hemisphere air at low latitudes between 17 and 25 July 2014.

The criterion for inertial instability is met where the sign of EPV is opposite to the planetary vorticity f : negative (positive) in the Northern (Southern) Hemisphere. These occurrences, marked as the hatched areas in Fig. 2, are seen as far as 40°S, but more generally attend breaking PWs protruding into low latitudes. Figure 3 shows SABER temperature structure (on the left) as a function of scaled log pressure and latitude at 170°W averaged between 19 and 21 July 2014 in the heart of the inertially unstable zone. The data have been interpolated to a 1 km log-pressure grid, and filtered of vertical wavelengths longer than 15 km. Thinly layered vertical oscillations are observed from between 25 and 60 km, with a prevailing vertical wavelength of roughly 10 km. Temperature excursions range between ± 6 K. These structures are very stable from day to day, and are observed over much of July and August 2014 (not shown). A comparison with MERRA-2 is shown in the right panel of Fig. 3. MERRA-2 temperatures have been averaged from the 00, 06, 12, and 18 Z files over the 19–21 July interval, interpolated in altitude, and filtered identically to SABER data. The two analyses show similarly vertically stacked structure

at the equator between 25 and 60 km. MERRA-2 temperature excursions are generally stronger than SABER.

The morphology of the layered features in both SABER and MERRA-2 temperatures, and their persistence, are consistent with pancake structures identified by Hitchman et al. (1987), Hayashi et al. (1998) and Smith and Riese (1999) in satellite

FIG. 2. MERRA-2 EPV on 18 Jul 2014 plotted on the 2000 K isentropic surface, corresponding roughly to 48 km. Hatching indicates regions of anomalous EPV that are inertially unstable.

FIG. 3. (left) SABER and (right) MERRA-2 temperature at 170°W, averaged between 19 and 21 Jul 2014.

temperatures. Based on the similarities between MERRA-2 and SABER analyses, we interpret the concurrent structuring in vertically filtered MERRA-2 temperatures as inertial instability, and proceed to examine the corresponding wind structures using MERRA-2. Both panels of Fig. 4 indicate MERRA-2 temperature shading, averaged between 18 and 20 July. The left panel overlays MERRA-2 perturbation zonal winds (contours) and vectors in the y - z plane. Contours in the right panel are vertical winds. Horizontal and vertical winds exhibit vertically layered structure, with a wavelength of approximately 10 km.

Sharp meridional wind gradients are observed at the boundaries of the inertially unstable zones (10°N and 25°S). The vertical winds track the temperature perturbations very closely, with rising (sinking) motions within the cold (warm) perturbations. The antiphase relationship between T' and w' was predicted by Hitchman et al. (1987) from the balance of vertical motion by radiative damping. Maximum convergence (divergence) is found at the zero line separating the warm (cold) centers that overlie cold (warm) centers, or where downward (upward) motion overlies upward (downward) motion. From continuity, the strongest meridional winds are expected along the zeroes of the T' and w' perturbations. This near-quadrature relationship between v' and T implies that the net heat transport within pancake structures is very weak.

Between 40 and 60 km, southward winds generally coincide with westward winds, while northward winds overlap eastward winds. The net horizontal momentum flux $u'v'$ for pancake structures therefore implies transport of eastward momentum from the winter hemisphere to the summer hemisphere. Peak eastward and westward winds within the pancake structure occur near the zero temperature and vertical motion contours.

Thus, the net vertical transport of horizontal momentum ($\overline{u'w'}$) is expected to be quite weak.

We quantify the effects of pancake structures on the mean wind using the Eliassen–Palm (EP) flux (denoted by $\mathbf{F}$), which is a measure of wave “activity” (Andrews 1987). Wave activity is a quantity that is quadratic in the amplitude of the perturbation, and conserved in the absence of forcing and dissipation. An exact expression in the primitive equation system on log-pressure surfaces appears in section 3 of Andrews (1987). Here, we focus not on the wave activity per se, but on two useful properties of $\mathbf{F}$ for waves that are small amplitude (i.e., at least an order of magnitude weaker than the $\bar{U}$). First, the divergence of $\mathbf{F}$ per unit mass describes the force contributed by the waves to the zonally averaged momentum budget. Second, for waves that are quasigeostrophic (i.e., weakly divergent), $\mathbf{F}$ is oriented parallel to the wave activity group velocity. Positive (negative) EP flux divergence indicates a wave source (sink) (Andrews 1987). These properties apply to the 2-day wave, to be discussed in section 4.

The components of $\mathbf{F} = (0, F^\phi, F^z)$ are given by

$$F^\phi = \rho_0 a \cos\phi \left(\bar{U}_z \overline{v'\theta'}/\bar{\Theta}_z - \overline{v'u'} \right), \quad (1)$$

$$F^z = \rho_0 a \cos\phi \left\{ \left[f - (a \cos\phi)^{-1} (\bar{U} \cos\phi)_\phi \right] \overline{v'\theta'}/\bar{\Theta}_{0z} - \overline{u'w'} \right\}. \quad (2)$$

The EP flux divergence is defined as

$$\nabla \cdot \mathbf{F} = (a \cos\phi)^{-1} \frac{\partial (F^\phi \cos\phi)}{\partial \phi} + \frac{\partial F^z}{\partial z}. \quad (3)$$

The zonally averaged momentum budget in the middle atmosphere is described by

FIG. 4. (left) MERRA-2 temperature and wind at 170°W, averaged between 18 and 20 Jul 2014. Color shading indicates temperature. Contours are zonal wind (solid is eastward, dotted is westward, at intervals of 3 m s⁻¹). Green arrows indicate the wind speed and direction in the y–z plane. Vertical winds are scaled by 100 m s⁻¹. Legend vector is 5 m s⁻¹. (right) Temperature (shaded) and vertical wind (contours; cm s⁻¹).

$$\frac{\partial \bar{u}}{\partial t} + \left[(a \cos \phi)^{-1} (\bar{U} \cos \phi)_{\phi} - f \right] \bar{v}^* + \frac{\partial \bar{u}}{\partial z} \bar{w}^* = (\rho_0 a \cos \phi)^{-1} \nabla \cdot \mathbf{F} \quad (4)$$

(Andrews et al. 1987). Equation (4) tells us that the zonal wave force upon the background atmosphere (as described by the EP flux divergence per unit mass) is balanced by the Coriolis force upon a global “residual” meridional wind ($\bar{v}^*$), advection of $\bar{U}$ by $(\bar{v}^*, \bar{w}^*)$, and the tendency of $\bar{U}$. Technically, the residual circulation $(\bar{v}^*, \bar{w}^*)$ is the component of the Eulerian mean circulation $(\bar{v}, \bar{w})$ that is not canceled by the convergence of the PW heat flux $(\bar{v}\theta'/\Theta_z)$. Physically, however, $(\bar{v}^*, \bar{w}^*)$ approximates the material zonally averaged motion, and is equivalent to the actual mass motion in the case of steady, conservative quasigeostrophic waves (Dunkerton 1978).

The EP flux vectors for pancake structures (not shown) are predominantly oriented in the y direction. This outcome was anticipated in the discussion of Fig. 4. The horizontal winds (u' and v') are in phase, leading to a strong F^{ϕ} , while the near-quadrature relationships between v' and T' and between u' and w' result in nearly nonexistent F^z . Figure 5 shows the EP flux divergence per unit mass of MERRA-2 pancake structures between 30°S and 30°N, and 20 and 60 km scaled heights. The pattern reflects the horizontal divergence of the momentum flux. The winter side of the equator between 30 and 50 km is characterized mostly by westward (negative) wave forcing. The summer flank of the equator is characterized by eastward acceleration out to

10°N, due to convergence of eastward momentum. Thus, pancake structures export eastward momentum from the winter hemisphere subtropics to the tropics, and across the equator into the summer subtropics. Acceleration values are on the order of individual meters per second per day, and reach maximum values between 40 and 50 km.

Figure 6 shows the change to $\bar{U}$ computed by Ortland’s nonlinear, time-dependent primitive equation model (described in section 2). A zonally symmetric version of the model was relaxed to a time-dependent $\bar{T}$ and $\bar{U}$ obtained from a smooth merging of the following datasets: daily MLS $\bar{T}$ and $\bar{U}$ from 20 to 95 km; daily $\bar{U}$ and $\bar{T}$ from MERRA-2 below 20 km; the Horizontal Wind Model (HWM) climatology (Drob et al. 2015), and the Mass Spectrometer and Incoherent Scatter (MSIS) radar model temperature climatology above 90 km (Hedin 1991). The relaxation to the specified zonal-mean background occurs at a rate of 0.5 day⁻¹.

The model is then run again, with forcing due to pancake structures represented by the EP flux divergence per unit mass in Fig. 5, and a slower relaxation rate for the wind (0.05 day⁻¹). The change in $\bar{U}$ is the difference in the two model runs, contoured in the top panel of Fig. 6. Close to the equator (where f is weak or nonexistent), the mean flow responds through $\partial \bar{U} / \partial t$. The evolution of $\bar{U}$ thus closely reflects the pancake structure EP flux divergence shown in Fig. 5. Winds on the summer flank of the equator increase eastward on the order of 6 m s⁻¹, while winds on the winter flank acquire a westward component of about 5 m s⁻¹.

FIG. 5. EP flux divergence per unit mass of MERRA-2 pancake structures (contours; $\text{m s}^{-1} \text{ day}^{-1}$) averaged between 18 and 20 Jul 2014.

The center plot in Fig. 6 shows $\bar{U}$ from the control (black) and experimental (red) runs at 45 km. The curves in the bottom panel indicate meridional shear (black) and curvature (red) of the contoured $\bar{U}$ difference pattern within 20° of the equator. The result of adding pancake structure forcing to the control run is to reduce $\bar{U}$ shear on the summer flank of the equator, and to enhance curvature. These effects stem from the enhancement of the eastward wind (“westerlies”) at the equator. On the southern flank of the equator, the increase in $\bar{U}$ shear represents a reduction of the negative shear that prevailed in the absence of the pancake structures.

The response of MERRA-2 $\bar{U}$ to inertial stability is seen most clearly near 40 km. Evolution of equatorial $\bar{U}$ at 39 km between 19 July and 14 August 2014 is shown in Fig. 7. A modest increase in $\bar{U}$ appears between the equator and 10°S , sandwiched between westward winds. The effects of inertial instability are highlighted by forming the difference of $\bar{U}$ from its values on 14 July. This evolution is shown in the middle panel of Fig. 7. An increase in eastward $\bar{U}$ is observed between 5°S and 10°N , while westward acceleration occurs on the winter flank. The $\bar{U}$ change on 25 July 2014 is plotted in the bottom panel (solid black curve), matched to the date in Fig. 6. The eastward wind enhancement peaks at 5 m s^{-1} , while the westward winds in winter hemisphere increase more sharply, by 15 m s^{-1} . Curvature of the change in $\bar{U}$ on 25 July (plotted in red) has a peak negative value of $8 \text{ m}^{-1} \text{ s}^{-1}$ at the equator, and positive regions between 5° and 10° on either side of the equator. The evolution of MERRA-2 $\bar{U}$ and its curvature are highly consistent with the model predictions shown in Fig. 6.

4. Two-day wave evolution

We have shown that the evolution of low-latitude MERRA-2 $\bar{U}$ between mid-July and early August 2014 is consistent with

FIG. 6. (a) Response of primitive equation model zonal wind (denoted $\delta\bar{U}$) on 20 Jul 2014 to EP flux divergence per unit mass shown in Fig. 5. (b) $\delta\bar{U}$ at 43 km from model run including pancake structure EP flux divergence (red curve). Control $\bar{U}$ at 43 km (black curve). (c) $\delta\bar{U}_y \times 10^5 \text{ s}^{-1}$ at 43 km (black curve). $\delta\bar{U}_{yy} \times 10^{11} \text{ m}^{-1} \text{ s}^{-1}$ at 43 km (red curve).

forcing by inertial instability, insofar as equatorial wind shear is reduced, and enhanced meridional curvature develops in low latitudes. We now examine the stability of baroclinic and/or barotropic perturbations to the summer hemisphere winds. Unstable baroclinic eddies amplify by extracting available potential energy from the mean flow, while barotropically unstable perturbations feed on the kinetic energy of the zonal mean flow. A necessary condition for instability of quasigeostrophic perturbations is that the meridional gradient of the zonally averaged quasigeostrophic potential vorticity ($\bar{q}$) reverse sign in the domain (Andrews et al. 1987):

$$\bar{q}_y = 2\Omega \cos\phi/a - \bar{U}_{yy} + \bar{U}_z \epsilon/H - \epsilon \bar{U}_{zz} - \epsilon_z \bar{U}_z, \quad (5)$$

where ϵ is the square of the ratio of f to the buoyancy frequency N . Equation (5) states that quasigeostrophic instability can occur in regions of positive meridional and vertical $\bar{U}$ curvature, negative vertical shear, and strong vertical gradients of static stability. It will be shown that for the case examined in this paper, all but the last term in (5) contribute to negative $\bar{q}_y$ somewhere in the summer hemisphere. The term involving the static stability gradient (ϵ_z) is nearly an order of magnitude smaller than the curvature terms, and generally negative, and is thus a stabilizing element. The unstable configuration of the

FIG. 7. (a) Evolution of low-latitude $\bar{U}$ in MERRA-2 at 39 km between 16 Jul and 15 Aug 2014. (b) Difference between (a) and $\bar{U}$ on 15 Jul 2014 (denoted $\delta\bar{U}$) at 39 km. (c) $\delta\bar{U}$ (solid curve) and $\delta\bar{U}_{yy} \times 10^{-11} \text{ m}^{-1} \text{ s}^{-1}$ (red curve) at 39 km on 25 Jul 2014.

midlatitude summer easterlies is generally attributed to GW drag (Norton and Thuburn 1996; Ern et al. 2013).

Our focus is on the instability of the low-latitude easterly jet, where enhanced positive curvature results from inertial instability, as shown in Figs. 6 and 7. Figure 8 shows the evolution of MERRA-2 $\bar{U}_{yy}$ and $\bar{q}_y$ in July and August 2014 at 43 km, a level transecting both the region of pancake structure forcing (as seen in Fig. 6), and the core of the stratopause subtropical easterly jet (discussed subsequently in relation to Fig. 11). At this altitude, the combined vertical shear ($\bar{U}_z$) and curvature ($\bar{U}_{zz}$) terms are an order of magnitude weaker than the remaining terms, and are not shown.

A channel of positive meridional $\bar{U}_{yy}$ (top panel) that is observed poleward of 9°N at the beginning of July migrates equatorward, receding northward after 4 August. The middle panel shows $\bar{q}_y$, which closely tracks the y -curvature pattern. The unstable region (negative $\bar{q}_y$) dips from 10° to 6°N on 27 July, disappears briefly, and reappears on 5 August, and migrates northward for the remainder of August. The bottom panel shows the evolution of inertial instability EP flux divergence per unit mass. The increase in positive $\bar{U}_{yy}$ and negative $\bar{q}_y$ aligns with eastward pancake structure forcing between the equator and 7°N during 17–25 July, and 4–12 August. Inertial instability appears to be a plausible source of the low-latitude background instability.

FIG. 8. (a) Evolution of MERRA-2 $\bar{U}_{yy} \times 10^{11} \text{ s}^{-1} \text{ m}^{-1}$ during July–August 2014 at 43 km. (b) Evolution of $\bar{q}_y \times 10^{11} \text{ s}^{-1} \text{ m}^{-1}$ during July–August 2014 at 43 km. (c) Divergence of pancake structure EP flux per unit mass ($\text{m s}^{-1} \text{ day}^{-1}$).

Many studies have identified the 2-day wave with baroclinic and barotropic instability of the summer easterly jet (Plumb 1983; Norton and Thuburn 1996; Lieberman 1999; Merzlyakov and Jacobi 2004; Siskind and McCormack 2014; France et al. 2018). The 2-day wave is actually a packet moving westward with a midlatitude zonal wind speed of approximately 60 m s^{-1} , most often comprised of a zonal wavenumber (s) 3 with a period near 2.1 days, and $s = 4$ with a period near 1.8 days. $s = 4$ is generally associated with instability (Pfister 1985; Lieberman 1999), while $s = 3$ is a free oscillation that is strongly influenced by $\bar{U}$ (Salby 1981). These waves have a westward zonal phase speed, and propagate vertically only in regions where the background wind is lightly eastward relative to the wave's zonal phase speed.

The evolution of the 2-day wave during July–August 2014 is shown in MERRA-2 temperatures in Fig. 9. In both wavenumbers 3 and 4, the 2-day temperature wave initiates in the tropical stratosphere (bottom panel), and migrates to the middle latitudes with altitude. Wavenumber $s = 3$ amplifies rapidly between 15 and 25 July, and dissipates abruptly. Wavenumber $s = 4$, whose presence is weak during July, rapidly amplifies following the collapse of $s = 3$.

Figure 10 shows typical temperature structure for $s = 3$ and 4 with a period of two days retrieved from MERRA-2 and NAVGEM analyses. Both datasets show excellent agreement

FIG. 9. Evolution of 2-day wave MERRA-2 T' amplitude. (left) $s = 3$. (right) $s = 4$. Heights are (top) 76 km (contour interval is 1 K), (middle) 57 km (contour interval is 0.5 K), and (bottom) 46 km (contour interval is 0.3 K).

below 80 km, and demonstrate the need for the NAVGEM high-altitude analyses in order to evaluate the effects of the 2-day wave on the PMC region (~ 83 km) that lies above the MERRA-2 ceiling. Comparison of Fig. 10 with the contoured values of $\bar{U}$ in Fig. 11 indicates that the 2-day wave exists in regions where westward $\bar{U}$ exceeds 20 m s^{-1} (i.e., $\bar{U} < -20 \text{ m s}^{-1}$). Thus, the wave is strongly asymmetric about the equator, with no significant amplitude in the winter hemisphere (where $\bar{U} > 0$). We note that $s = 4$ peaks at approximately 30°N and 65 km, about 10° closer to the equator and 10 km lower than $s = 3$.

Figure 11 shows NAVGEM $\bar{U}$ on 27 July 2014, when $s = 3$ and $s = 4$ were both present. Black hatched regions denote negative $\bar{q}_y$. The arrows show the direction of the normalized 2-day wave EP flux vector $\mathbf{F}$, computed by applying Eqs. (1)–(3) to perturbations that are the sum of $s = 3$ with a period of 2 days, and $s = 4$ with periods of 2 and 1.5 days. Also plotted (in white) are the points where the phase speed of the 2-day wave coincides with $\bar{U}$.

The key features of Fig. 11 are as follows:

- 1) $\mathbf{F}$ is oriented mostly horizontally equatorward of 30°N , indicating that transfer of wave energy there is effected through the meridional convergence of horizontal momentum flux.
- 2) At midlatitudes between 60 and 80 km, $\mathbf{F}$ points upward, indicating energy transfer through the convergence of meridional heat transport.

- 3) A quasigeostrophic instability condition ($\bar{q}_y < 0$) exists within the core of the westward wind system in the summer hemisphere. This region is collocated with the 2-day wave, as previously noted when comparing Figs. 10 and 8.
- 4) The “critical line”—where the phase speed of the 2-day wave equals $\bar{U}$ —lies within the unstable region, lending support to the association of instability with wave “over-reflection” at the critical line (Lindzen and Tung 1978).
- 5) $\mathbf{F}$ emanates (diverges) from the area between 10°N at 45 km and 40°N at 65 km.

The juxtaposition of features 3, 4, and 5, together with the properties of quasigeostrophic $\mathbf{F}$, suggest that the 2-day wave has a source in the unstable region in the subtropical stratopause (45–65 km) where $\mathbf{F}$ is divergent. The origin of the instability in that latitude–altitude regime is meridional curvature of $\bar{U}$, as suggested in Fig. 8.

We present in Fig. 12 the zonal forcing of the 2-day wave (i.e., its EP flux divergence per unit mass), the two-dimensional circulation ($\bar{v}^*$, $\bar{w}^*$) induced by the forcing, and the $\bar{T}$ response. The calculations are made with a quasigeostrophic model of the zonally averaged circulation developed by Garcia (1987), and adapted by Lieberman (1999) and France et al. (2018) to 2-day wave observations. Two-day wave driving is eastward in the lower mesosphere (55–70 km) and westward in the MLT

FIG. 10. Two-day wave T' amplitude as a function of latitude and altitude averaged over 3 days centered on 27 Jul 2014. (left) $s = 3$. (right) $s = 4$. (top) MERRA-2. (bottom) NAVGEM.

(70–100 km). This forcing is balanced primarily by the Coriolis force, through northward $\bar{v}^*$ where the wave driving is westward (negative). From mass continuity, the residual mean vertical wind ($\bar{w}^*$) is upward on the equatorward flank of the wave forcing, and downward on the poleward flank. The $\bar{T}$ balances the downward (upward) motion through adiabatic heating (cooling). Using a Newtonian damping coefficient of 0.5 day^{-1} produces a 2-day wave-induced warming between 2 and 3 K in the high-latitude mesopause.

Figure 13 shows the evolution of the PMC frequency of occurrence averaged between 70° and 80°N , derived from the CIPS instrument. Also shown is the evolution of $\bar{T}$ induced by 2-day wave forcing throughout the same latitude range, between 80 and 85 km in geometric height. PMC decline starts on 26 July, and reaches a maximum on 6 August, during peak 2-day wave warming. Comparison with Fig. 9 shows that the warming and PMC decline were initiated when the 2-day wave transitioned from $s = 3$ to $s = 4$. During this time, the low-latitude westward jet was unstable, apparently due to forcing by inertial instability (see Fig. 8).

The elements of IHC set out by Limpasuvan et al. (2000) linking the 2-day wave to the wintertime stratosphere appear to all be in place between 19 July and 6 August: 1) inertial instability (organized by winter PW breaking) driving of subtropical $\bar{U}$ curvature satisfying baroclinic/barotropic instability

conditions in the summer subtropics; 2) 2-day wave amplification and its heating of the summer mesopause. We note, however, that the $s = 3$ two-day wave is a free oscillation of the atmosphere (Salby 1981). It therefore seems reasonable to ponder the need for inertial instability-induced wind curvature as a trigger. In the following section we explore the sensitivity of the 2-day wave to wind curvature at the tropical stratopause.

5. Sensitivity of 2-day wave to tropical $\bar{U}$

In the previous section we laid the observational groundwork for direct linkage between inertial instability, tropical wind destabilization, 2-day wave amplification, and summer mesopause warming. However, generation of the 2-day wave—a major element of the IHC—may occur without any direct source, owing to its identity as a free oscillation. To what extent, then, does inertial instability truly influence 2-day wave occurrence and amplitude? In this section we explore the sensitivity of the 2-day wave spectrum to the location and intensity of tropical $\bar{q}_y$.

Ortland's nonlinear, time-dependent primitive equation model (described in section 2) is used to examine the atmospheric response to tropical wind forcing induced by inertial instability. The model simulates wave propagation through background $\bar{U}$ and $\bar{T}$ that are constrained to daily, data-specified values. The constraint of the model background

FIG. 11. NAVGEM $\bar{U}$ on 30 Jul 2014 (contour interval is 10 m s^{-1}). Red arrows are normalized $\mathbf{F}$ vectors. Black hatched regions denote negative $\bar{q}_y$. F_z has been multiplied by 100 for greater visibility. White symbols indicate locations where zonal phase speed of the 2-day wave equals $\bar{U}$.

implies that the zonal-mean evolution of the model is largely governed by observations. This strategy is a crude representation of the forces that shape the zonal-mean structure of the atmosphere, since there is no attempt to reproduce these forces in the model. One implication is that the zonal-mean evolution is partly shaped by the actual 2-day wave that was present in the atmosphere.

The 2-day wave is seeded in the following manner: Nonlinear momentum and heat advection are computed from the full spectrum of waves resulting from a historical model run driven by tropical and midlatitude heating. In all of the experiments described below, the model is forced by using these advection terms as sources of momentum or thermal energy. The only other forcing applied in the model is a relaxation of $\bar{U}$ and $\bar{T}$ to the background zonal mean described in section 3, as it is varied daily over Julian days [aka days of year (doy)] 180–270 (29 June–27 September 2014). The advective forcing is only applied over the first 10 days, during which it is gradually ramped up and down. Because this is a spectral model, the time tendencies are computed separately for each zonal wavenumber. For our experiments, this range is restricted to the zonal mean, $s = 3$, and $s = 4$. This configuration is known as quasi linear, in which the waves only interact with the mean flow, but not with each other.

Simulations in which the model was relaxed to the background described in section 3 did not spontaneously produce a 2-day wave. To stimulate the spontaneous generation, the background was modified by deepening the westward $\bar{U}$ by 10 m s^{-1} in a Gaussian shaped region centered at 45°N and 70 km, with half width of 10° latitude, and half width of 15 km in the vertical. This enhancement resulted in stronger curvature and a deeper negative minimum in $\bar{q}_y$. We refer to the simulation using this modified background as the control (CTRL) run, illustrated for day 195 (14 July) in the left panel of Fig. 14.

The CTRL run, seeded with the small waves excited by the advective forcing, spontaneously produced a strong $s = 3$ two-day

wave in the midlatitude summer hemisphere (not shown), with amplitude first exceeding 1 K mid-June, and peaking at around 7 K in mid-July, after which it fades by early August. There is no significant $s = 4$ response in CTRL. However, further modification to $\bar{U}$ that mimics the effects of inertial instability did produce an $s = 4$ two-day wave. The modification (SMOD) consists of enhancing the curvature of $\bar{U}$ at the summer subtropical stratopause. An example for day 195 (15 July) is shown in the right panel of Fig. 14.

In SMOD, the $s = 3$ two-day wave is essentially unchanged except for a slight increase in the growth rate. However, this simulation produces a large $s = 4$ disturbance, that is at first confined to the easterly wind regimes in the subtropical stratopause. During this stage it exhibits the structure of a barotropically unstable, symmetric Rossby wave (RW). This mode is illustrated in the left panel of Fig. 15, for day 205 (24 July). It consists of two counterrotating cells offset about 5°N of the equator, exhibiting a net positive flux of horizontal momentum ($\bar{u}'v'$) at the equator and subtropics. Its EP flux vectors (not shown) are nearly horizontal, emanating from 15°N and $35^\circ\text{--}45^\circ\text{N}$, and directed toward the Southern Hemisphere. By the end of July (day 212) the RW is replaced by a Rossby–gravity wave (RGW) that fills the Northern Hemisphere. The $s = 4$ reaches an amplitude of about 4 K in early August, after $s = 3$ has died away.

Further light is shed on the evolution of $s = 4$ by performing two more simulations: CTRL4 and SMOD4. These experiments use the two backgrounds from the two experiments already described, but with the $s = 3$ suppressed by setting their time tendencies in the model to zero. In the CTRL4 simulation, $s = 4$ two-day wave with RGW structure forms early, growing steadily within the summer hemisphere easterlies (westward winds) from day 205 (24 July) to day 245 (2 September), when it reaches a peak amplitude of about 6 K (shown in the right-hand panel of Fig. 15). This simulation illustrates that the $s = 3$ two-day wave suppresses the $s = 4$ wave in the control run. It does so by modifying and stabilizing the mean flow before $s = 4$ can grow upon the unstable configuration.

The SMOD4 simulation produces both the $s = 4$ RW and RGW modes. However, relative to SMOD (that included $s = 3$), the RGW mode appears 20 days earlier, on day 215 (3 August), compared to day 235 (23 August) in SMOD. Without the suppression of the $s = 4$ by the $s = 3$ two-day wave, the tropical RW mode is a less necessary precursor to the RGW mode. When $s = 3$ is artificially suppressed, the RGW manifold evolves within the subtropical curvature much earlier.

To summarize, we have found that the $s = 3$ two-day wave is relatively insensitive to perturbations upon a control $\bar{U}$ with sufficient midlatitude curvature. By contrast, $s = 4$ launches and evolves in the model summer hemisphere only when $s = 3$ is suppressed, or when curvature of $\bar{U}$ is enhanced at the subtropical stratopause. In the latter circumstances, $s = 4$ initiates as a RW at the low latitude stratopause, and subsequently evolves into a RGW only when the $s = 3$ life cycle was complete.

Several elements of this simplistic model are pertinent to the observations. These include the injection of higher $\bar{U}$ curvature at the model low latitudes (a proxy for the real-time effects of

FIG. 12. (a) EP flux divergence per unit mass of the 2-day wave. (b) $\overline{\Delta T}$ induced by the 2-day wave. (c) $\overline{v^*}$ induced by the 2-day wave. (d) $\overline{w^*}$ induced by the 2-day wave.

inertial instability), the initiation of $s = 4$ at the subtropical stratopause, and its rapid amplification following the collapse of $s = 3$ (seen in Fig. 9). We conclude that inertial instability is a likely source of the $s = 4$ two-day wave between 25 July and 4 August 2014.

6. Summary and conclusions

The goal of this study was to obtain observational confirmation of a mechanism described by Orsolini et al. (1997) and Limpasuvan et al. (2000) that identified inertial instability as a source of the 2-day wave in the summer mesopause. This process is a crucial link in an interhemispheric teleconnection relating anomalous warming of the summer mesopause to the disturbed polar vortex in the winter stratosphere. Observationally, this teleconnection manifests as a correlation between negative PMC occurrence anomaly, and positive stratospheric T anomaly in the opposite hemisphere (Karlsson et al. 2007).

We examined inertial instability, and the evolution of tropical $\overline{U}$ and the 2-day wave during 25 July–5 August 2014. Evidence for inertial instability is seen in SABER and MERRA-2 temperatures as persistent regions of anomalous EPV, and vertically thin, layered “pancake structures” over the central Pacific. These features transport eastward momentum from the winter hemisphere into the low-latitude

summer hemisphere. We computed their EP flux divergence per unit mass, a measure of wave forcing. The observed pancake structures extract eastward momentum from the tropical mean flow on the winter flank of the equator, and accelerate the tropical mean flow eastward between 10°S and 10°N. The net effect of this forcing is to reduce shear and enhance curvature of $\overline{U}$ between 5° and 15°N.

FIG. 13. PMC occurrence frequencies averaged over 70°–80°N (solid curve; left y axis). $\overline{T}$ induced by the 2-day wave, averaged over the same latitudes, and 80–85 km geometric altitude (dashed curve; right y axis).

FIG. 14. $\bar{U}$ on day 195 plotted as a function of latitude and altitude. Dashed red contours indicate negative $\bar{q}_y$. White contour is the 2-day wave critical line. Contour interval is $10^{-11} \text{ m}^{-1} \text{ s}^{-1}$. (left) Control run. (right) Modified control run with enhanced subtropical $\bar{U}$ curvature, and instability at the summer subtropical stratopause.

MERRA-2 data reveal a channel of subtropical negative $\bar{q}_y$ (an index of quasigeostrophic instability) whose source is positive $\bar{U}$ curvature during much of the July–August 2014 interval. During this period, the 2-day wave develops a strong amplitude, that transitions at the end of July from $s = 3$ to an unstable, forced $s = 4$. The 2-day wave induces a hemispheric-scale summer circulation that warms the mesosphere poleward of 40°N . The PMC region ($70^\circ\text{--}80^\circ\text{N}$ and $80\text{--}85 \text{ km}$ altitude) warms by approximately 2 K , while PMC occurrence frequency drops from 60% to 5% .

To investigate the sensitivity of the 2-day wave to subtropical stratospheric $\bar{U}$, we perturbed $\bar{U}$ and $\bar{q}_y$ within a primitive equation model. The $s = 3$ component of the 2-day wave, a free oscillation, is relatively insensitive to an initially unstable summer easterly $\bar{U}$. The $s = 4$ component, however, was quite responsive to the wind changes. Specifically, $s = 4$ developed within an unstable midlatitude region only when $s = 3$ (along with its stabilizing influence on $\bar{U}$) was eliminated. A similar responsiveness of $s = 4$ under these conditions was reported by Limpasuvan et al. (2000). When negative $\bar{q}_y$ extends to the

subtropical stratopause, $s = 4$ initiates within a narrow tropical latitude band, and subsequently expands into the summer hemisphere MLT. In our model, the intensification of negative $\bar{q}_y$ is a proxy for the effects of inertial instability on the tropical stratopause $\bar{U}$ curvature. Our numerical experiments therefore underscore the role of inertial instability in seeding a $s = 4$ two-day wave that can attain observed amplitudes (10 K) in the MLT, and warm the PMC region.

Our study confirms the role of inertial instability as an agent that can couple the 2-day wave in the summer hemisphere to breaking stratospheric winter PW activity, that is presumed to be the source of the instability. This work, however, is confined to an individual case that has been interpreted within a highly restricted modeling framework. Further confidence in the inertial instability-driven IHC must be established by investigating additional occurrences, that link both the boreal and austral cold summer mesopauses with their wintertime counterparts. The effects of inertial instability on the 2-day wave-driven warming of the summer mesopause must also be placed in the context of other processes. Modeling studies (Karlsson and Becker 2016)

FIG. 15. The $s = 4$ two-day wave from SMOD4 experiment. (left) Geopotential height and horizontal winds, as a latitude vs longitude plot, on day 205 at 47 km . (right) Altitude–latitude cross section of T amplitude, on day 245.

have demonstrated that warming of the summer mesopause results from reduction of global GW driving during periods of strong winter PW activity. A recent analysis by Smith et al. (2020) indicates that, statistically, the anomalous warm summer mesopause is unrelated to specific wave activity in the summer hemisphere, and is merely the extension of a hemispheric-scale circulation induced by heightened PW driving in the winter stratosphere.

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The authors dedicate this paper to the memory of William Albert Lahoz (1960–2019).

“Let your ideas be understood without making them complicated or obscure.” —Miguel de Cervantes

Data availability statement. SABER temperatures are archived at <http://saber.larc.nasa.gov>, and at saber-gats-inc.com/custom/Temp_O3_H2O/v2.0/. MLS v4.2 data are available from the NASA Goddard Space Flight Center Earth Sciences (GES) Data and Information Services Center (DISC) at <https://mls.jpl.nasa.gov/data/>. MERRA-2 data are available at MDISC, managed by the NASA GES DISC at <https://gmao.gsfc.nasa.gov/reanalysis/MERRA-2/>. Inertial instability EP fluxes, 2-day wave forcing, PMC evolution, and the induced residual circulation are available through Open Science Framework at <https://osf.io/85mzt/>. Hourly NAVGEM forecast-assimilation files for July–August 2014, and output from CTRL, CTRL4, SMOD and SMOD4 experiments are also hosted at <https://osf.io/85mzt/>.

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