Austral Spring Stratospheric and Tropospheric Circulation Interannual Variability

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ABSTRACT

The relationship between the October (spring) total ozone column (TOC) midlatitude zonal asymmetry over the Southern Hemisphere (SH) and the stratospheric quasi-stationary wave 1 (QSW1) interannual phase variability is analyzed. Once contributions to the TOC from known global predictors, estimated with a multi-regression model, are removed, the residual TOC interannual variability is observed to be dynamically coupled to the stratospheric QSW1 phase behavior. The stratospheric QSW1 interannual phase variability, when classified according to specifically designed indices, yields different circulation patterns in the troposphere and stratosphere. High (upper quartile) index values correspond to a westward rotation of the midlatitude ozone trough and the stratospheric QSW1 phase, while low (lower quartile) index values represent their eastward-rotated state. These values can be associated with statistically different tropospheric circulation patterns: a predominantly single poleward tropospheric jet structure for high index values and a predominantly double-jet structure for low index values. For the latter, there is a higher daily probability of double-jet occurrence in the troposphere and a stronger stratospheric jet. These jet structures and their daily behavior are supported by significant synoptic-scale activity anomalies over SH mid- to high latitudes as well as changes in tropospheric quasi-stationary waves 1–3. The wave activity flux (W flux) diagnosis shows the contribution of active quasi-stationary waves in the observed tropospheric anomalies associated with high and low index values. With low index values, the quasi-stationary waves lead to a self-sustaining state of the stratospheric–tropospheric coupled system. With high index values, the overall mid- to high latitude circulation is associated with wave energy propagation from the tropical central Pacific into higher latitudes. Thus, during the austral spring, there are interactions between the troposphere and stratosphere, leading to the locally well-defined upward and downward propagation of wave anomalies, that is, significant upper troposphere (UT)–lower stratosphere (LS) interactions can occur within a spring month itself.

1. Introduction

The nature of the interannual ozone content variability at any given location is largely attributed to dynamics, especially through the interaction of tropospheric planetary waves with the stratospheric mean state (Fusco and Salby 1999; Hood and Soukharev 2005; Randel et al. 2002; Canziani et al. 2008). These planetary waves can be both transient and stationary. Stratospheric stationary waves have their source in the troposphere, and their variability can be found both in the troposphere and the stratosphere (Hirota and Sato 1969; Shiotani and Hirota 1985; Plumb 1989; Hio and Hirota 2002; among others).

The maximum amplitude of the ozone zonal asymmetry occurs during austral spring, in particular during October, together with a large stationary wave 1 (SW1; Labitzke and van Loon 1999; Grytsai et al. 2007). According to Moustaoui et al. (2003), the austral spring stratospheric mean state is dominated by an SW1, as can be observed in temperature and geopotential, with the maximum amplitude also in October. They argue that SW1 is responsible for the ozone distribution at mid- to high latitudes of the Southern Hemisphere (SH), in the low and middle stratosphere, because of associated vertical displacements that induce trace gas density variations there. They noted that SW1 does not experience phase variation year in, year out but without statistical evidence to support their conclusion.

However, during spring the SH monthly-mean total ozone column (TOC) as well as the lower stratosphere (30–100 hPa) dynamic variables exhibit transient quasi-stationary wave 1 (QSW1) variations, a behavior not
observed in the Northern Hemisphere (Malanca et al. 2005; Grytsai et al. 2007; Jiang et al. 2008). Visual inspection of SH October monthly-mean TOC fields for the period 1979–2005 mapped from the Total Ozone Mapping Spectrometer (TOMS) retrievals shows that the midlatitude TOC has an interannual phase variability throughout the period, which does not agree with the Moustaoui et al. (2003) assessment of SW1 stationarity.

In fact, Hio and Hirota (2002) argued, through a principal component (PC) analysis and subsequent zonal-mean vertical Eliassen–Palm (EP) flux compositing, that there exist interannual phase variations in geopotential stratospheric SW1 at 10 hPa (their first EOF), which would be primarily linked to the contribution of tropospheric synoptic-scale (wavenumber \( k \geq 4 \)) processes below 200 hPa, while interannual SW1 amplitude variability (their second EOF) is distinctly linked to planetary waves 1–3 in the troposphere and lower stratosphere. Nevertheless, they were unable to explain how synoptic-scale activity below 200 hPa is associated with the SW1 phase variability.

As a result, the aim of this work is to examine and discuss the nature of the stratospheric SW1 interannual phase variability (QSW1 because of the observed variations) together with the SH midlatitude TOC zonal asymmetry evolution and their relationship with transient synoptic and quasi-stationary planetary wave activity and zonal circulation in the troposphere during October. The analysis shows that the temporal anomalies in the stratospheric QSW1 are linked with zonally localized quasi-stationary tropospheric wave packets, as derived from the wave activity flux (W flux), proposed by Takaya and Nakamura (2001). The interactions between tropospheric quasi-stationary waves, the synoptic-scale transient activity, and the resulting troposphere–stratosphere coupling and TOC variability are discussed. The analysis yields a strong coupling between lower stratospheric and tropospheric processes during spring.

2. Data and methodology

CH TOC \( 1^\circ \times 1.25^\circ \) monthly-mean fields, 1979–2005, were obtained from the TOMS dataset. This monthly-mean fields sequence has a gap between 1993 and 1995 because of a malfunction in the TOMS Meteor instrument. Geopotential height (GPH, m), temperature, vertical motion anomalies (omega; Pa \( \text{s}^{-1} \)), and horizontal \((U \text{ and } V)\) wind components (m \( \text{s}^{-1} \)) were obtained from the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis monthly- and daily-mean \( 2.5^\circ \times 2.5^\circ \) products. The dynamics dataset was extended through 2007 when available.

To determine objectively the TOC’s spatial variations during October, in the easterly and westerly displacements of the midlatitude/subpolar ozone minimum, it was necessary, first, to evaluate the contributions of primary global physical and chemical processes related to the ozone fields—that is, the 11-yr solar cycle (solar), the quasi-biennial oscillation (QBO), ozone depletion [as given by the equivalent effective stratospheric chlorine (EESC) loading], and the Brewer–Dobson (BD) meridional residual. In particular, the BD circulation proxy considered here is the cumulative vertical zonal-mean EP flux through 100 hPa, averaged between 45\(^\circ\)S and 75\(^\circ\)S, as discussed in Randel et al. (2002). The term cumulative implies a weighted mean of the monthly-mean vertical zonal-mean EP flux through 100 hPa, obtained from the Chemical and Dynamical Influences on Decadal Ozone Change (CANDIDOZ) project, beginning during April of each year under study, for the SH. Each year the cumulative flux is set to zero at the beginning of April, thus reproducing the stratospheric ozone meridional transport yearly cycle from the tropics into the high latitudes (Fioletov and Shepherd 2003). The cumulative EP flux is obtained according to Brunner et al. (2006) as follows:

\[
BD(t) = BD(t-1) \exp\left(-\frac{\Delta t}{\tau}\right) + EP(t),
\]

where \( BD(t) \) is the Brewer–Dobson proxy at time \( t \), as given by the final cumulative EP flux; \( \Delta t \) is the time step between \( t-1 \) and \( t \); \( \tau \) is the e-folding decay time, which equals 12 between April and September—that is, the TOC build-up phase—and 3 between October and March; and \( EP(t) \) is the CANDIDOZ 45\(^\circ\)–75\(^\circ\)S mean 100-hPa EP flux at time \( t \).

These processes were selected following Mäder et al. (2007). While it is usually assumed that they are either longitudinally independent or that possible longitudinal differences can be disregarded using potential vorticity (PV) equivalent latitude, the approach applied here needs to preserve the dynamically induced longitudinal differences while removing all other sources of variability. Thus, all calculations were carried for each gridpoint series. The respective contributions were estimated using a multiregression statistical model with seasonality, as given by Ziemke et al. (1997) and updated by Brunner et al. (2006) as follows:

\[
TOC(\phi,\lambda,t) = TOC(\phi,\lambda,t) + \beta(\phi,\lambda,t)EESC(t) + \gamma(\phi,\lambda,t)QBO(t) + \delta(\phi,\lambda,t) \cdot \text{solar}(t) + \eta(\phi,\lambda,t)BD(t) + \varepsilon TOC(\phi,\lambda,t),
\]
where $\phi$ is latitude; $\lambda$ is longitude; $t$ is the time index in months ($t = 1, \ldots, 312$), beginning January 1979; $\text{TOC}(\phi, \lambda, t)$ is the TOC 12-month seasonal harmonic fit (12, 6, 4, and 3 months); $\text{EESC}(t)$ is the equivalent effective stratospheric chlorine loading time series; $\text{QBO}(t)$ is the zonal-mean equatorial zonal wind at 30-hPa time series from the NCEP–NCAR reanalysis; $\text{Solar}(t)$ is the 10.7-cm solar flux (F10.7) time series, provided as a service by the National Research Council Canada; $\text{BD}(t)$ as in Eq. (1); $\beta(\phi, \lambda, t)$, $\gamma(\phi, \lambda, t)$, $\delta(\phi, \lambda, t)$, and $\eta(\phi, \lambda, t)$ are the corresponding multiple regression coefficients calculated as harmonics of the annual cycle (12, 6, and 4 months) and $\varepsilon \text{TOC}(\phi, \lambda, t)$ is the TOC residual field for each month, that is, the dynamically induced TOC variations.

Daily atmospheric data were time filtered to focus on the high-frequency (HF) synoptic scale (2–8 days), using a bandpass filter. October monthly-mean HF wave activity $Z^2$ fields at 300 hPa are calculated according to Canziani et al. (2008).

The anomalous W flux was calculated following Takaya and Nakamura (2001) and Nishii and Nakamura (2004), which is defined as

$$
W = \frac{P}{2|U|} \left\{ U(\nu'^2 - \psi'v') + V(-u'v' + \psi'u') \\
U(-u'v' + \psi'u') + V(u'^2 + \psi'u') \\
f_c R_o N^2 H_o \left[ U(\nu'T' - \psi'T') + V(-u'T' - \psi'T') \right] \right\},
$$

(3)

October TOC field (Oct $\varepsilon \text{TOC}$) is considered. The standard deviation (SD) field Oct $\varepsilon \text{TOC}$ was estimated to determine the areas where its interannual variability is largest. It has maximum values south of Madagascar, close to 60°S–65°S, 30°–70°E (Fig. 1a), highlighting an area with behavior distinctly different from the regression model output, with a higher sensitivity to other processes not contemplated by it. The associated $\varepsilon \text{TOC}$ time series averaged for grid points with peak standard deviation (centered at 61°S, 40°E) was used to build an index for TOC variability not explained by the multiregression model: the VARMAX index (Fig. 1b). To obtain a dimensionless index, VARMAX, the $\varepsilon \text{TOC}(61°S, 40°E)$ time series median $[-1.67 \text{ Dobson units (DU)}]$ is subtracted from it and then divided by the interquartile range [the absolute difference between the lower (q1 = −19.13 DU) and upper (q3 = +23.64 DU) quartiles] throughout the period sampled.

Low and high VARMAX values were characterized by the upper and lower quartiles. The corresponding composite mean fields were calculated according to this criterion. Figure 2a shows the mean October TOC field over the SH for the sample. Figures 2b and 2c show the composites for October $\varepsilon \text{TOC}$ fields identified with high and low VARMAX values, respectively. There is almost a 180° phase shift in the residual anomalies between the high- and low-index composites: a flip-flop behavior can be distinctly observed in these VARMAX composite residual fields, with a meridional axis close to 60°W–120°E, which is not related to the regression model predictors considered. The magnitude of the residual above the high-latitude Indian Ocean ($\sim$60°–70°S, 90°–120°E) is close to 50 DU—that is, approximately 15%–20% of TOC—while changes in the South Pacific, where a
secondary maximum is located near 65°–70°S, 90°–170°W, and are at most approximately 20 DU. Thus, the VARMAX index captures the midlatitude eTOC phase variations. Note that these residuals, when added to the mean TOC field describe reasonably well the zonal interannual phase variability observed in monthly October asymmetric TOC fields during the years sampled.

Moustaoui et al. (2003), in a case study using Microwave Limb Sounder data from the Upper Atmosphere Research Satellite (UARS) mission for the late austral spring of 1993, have shown that the stratospheric QSW1 perturbs the isentropic surfaces, redistributing ozone in such a way that the ozone fields closely mimic the observed pattern in the stratospheric GPH field. Hence, a relationship between the VARMAX index’s temporal evolution and that of a planetary wave variability index could provide a link between the observed residual ozone variability and stratospheric dynamics.

Figure 3a shows the GPH stationary waves 1–3 climatology at 30 hPa, estimated using harmonic analysis. The pattern points to a dominant QSW1 but not solely a wave 1 structure, as previously noted by Quintanar and Mechoso (1995). Brahmananda Rao et al. (2004) show that the October 30-hPa geopotential quasi-stationary waves 1, 2, and 3 have amplitude ratios close to 10, 3, and 1, respectively. Figure 3b highlights its interannual variability, as given by its standard deviation. Two regions of high variability can be observed, with peak variability centered near 65°S, 57.5°E to the southeast of southern Africa and the other at 65°S, 120°W to the west of southern South America. The comparison with TOC residuals in Fig. 1b shows significant coincidence in the areas of largest standard deviation, particularly in the area to the southeast of southern Africa. Note that while the TOC residual shows a much larger variability over the Indian Ocean, the GPH standard deviation values are rather close on both sides of Antarctica.

A measure of the interannual phase displacements of these stationary waves can be estimated using the difference between grid points A and B in the monthly mean October GPH. The series, which can be obtained from both filtered waves 1–3 and raw data, with few if any differences, is the 30-hPa planetary wave maximum difference (30 PW) index. The dimensionless 30-PW
index (calculated in a manner similar to VARMAX) is shown in Fig. 3c, overlaid with VARMAX for comparison. Positive values imply that the stationary wave pattern’s negative lobe is westward rotated toward South America, while negative values imply this lobe is rotated toward South Africa. Since the correlation between the VARMAX and the 30-PW indices is 0.83 (99% significant), most of the October variability at this stage could be mainly attributed to interannual changes in the longitudinal state of stratospheric planetary wave 1.

Although Moustaouï et al. (2003) demonstrated the relationship between the vertical QSW1 structure and the resulting TOC spatial distribution, they concluded, without any justification, that the SW1 did not show interannual phase changes. In contrast, the current analysis shows that a considerable fraction of the interannual TOC variability occurs primarily in response to QSW1 phase changes. Furthermore, it is now possible to infer, using 30 PW, an approximate $\varepsilon$TOC behavior during the missing years 1993–95, effectively allowing for the study of the complete 1979–2007 period. This result also suggests that the multiregression model can be improved by adding a QSW1 phase index as a predictor, such as 30 PW, at least for October.

The 30-PW series shows both a significant interannual variability and a change in trend before and after the early 1990s. Before the early 1990s, there is a positive trend, which, if considered between 1982 and 1991, has a maximum correlation with a positive linear trend of
0.74 (99% significance level). Before 1982, there appears to be a period with negative trend; however, it is too short to evaluate statistically. After 1991, the negative trend shows a 0.38 correlation, significant at 90%. Grytsai et al. (2005) presented an SH planetary waves evolution for the period 1979–2003. Their wave intensity study for waves 1–3 shows that wave 1 has a different variability before and after the early 1990s, approximately. Thus, the negative trend also corresponds to the eastward migration of the mean state of planetary waves 1–3 after the early 1990s, in agreement with the TOC observations (Malanca et al. 2005). However the long-term evolution of the observed interannual variability requires further analysis, which is beyond the scope of the current study.

To examine the GPH anomaly phase variability and its links in the troposphere, composite anomalies associated with high and low 30-PW values are analyzed. The upper- and lower-quartile index values are considered, as for VARMAX. Figure 4 shows the high and low 30-PW years GPH anomaly field composites at 30 (Figs. 4a and 4b), 100 (Figs. 4c and 4d), and 300 hPa (Figs. 4e and 4f). Dark/darker (lighter/light) gray-shaded areas correspond to statistically significant (90%/95%) positive (negative) temporal anomalies, as given by the unequal variance t test. Figures 4a and 4b show the stratospheric QSW1 to be the dominant feature in the stratospheric anomalies. The differences between low and high 30-PW anomaly fields show the flip-flop behavior, in correspondence with eTOC (Figs. 3a and 3b). This further argues for the dynamic nature of TOC residual changes. For stratospheric QSW1-dominated anomalies during high index values, the stronger positive temporal anomalies occur close to Antarctica over the Indian Ocean (−65°–70°S, 45°–90°E), with amplitudes of greater than 3500 m at 30 hPa and 1000 m at 100 hPa (Fig. 4a), while a weaker negative anomaly appears over the Bellinghausen Sea/Drake Strait (−60°S, 70°–80°W), with amplitudes close to −2000 and −800 m at 30 and 1000 hPa, respectively. For low index values, the pattern is inverted and the negative lobe is now found where the high-index positive lobe was, almost at the same location, over the Indian Ocean. Peak amplitudes are −3000 and −1200 m at 30 and 100 hPa, respectively, while the now positive lobe has moved west with respect to the previous high-index negative one. It maximizes near 60°S, 90°–100°W, with peak values greater than 2500 and 800 m at 30 and 100 hPa, respectively. Note that the 30 and 100 hPa anomalies are vertically collocated over the Indian Ocean; however, the 100-hPa anomalies in the Western Hemisphere are found to the east of the 30 hPa ones, over the western South Atlantic. Thus, although the 30-hPa anomalies are almost in antiphase across the South Pole—that is, mostly a QSW1 structure—at 100 hPa the anomalies are more distorted, suggesting contributions from other planetary waves. Although the standard deviation used to define the 30-PW index has approximately the same amplitude in both areas, the 30-hPa composites yield an amplitude range over the Indian Ocean sector larger than that observed over the South Pacific, at high latitudes, in qualitative agreement with the eTOC spatial differences. Such an interhemispheric difference suggests that even at 30 hPa, despite the significant QSW1 contribution, higher-order planetary waves also contribute to the anomaly fields.

The upper-troposphere 300-hPa anomaly fields, composited with the stratospheric 30 PW index, show statistically significant anomalies with a more complex pattern (Figs. 4e and 4f). For the high index values, there is a large positive GPH anomaly with statistically significant areas extending from Ross Bay, Antarctica (−65°S, 180°), to Patagonia in southern South America (50°S, 70°W) and a somewhat weaker negative, significant anomaly extending between Australia and Antarctica along the Indian Ocean toward South Africa at mid- to high latitudes. Another negative, though not significant, anomaly appears to the north of Weddell Sea (−55°S, 20°W). In the Pacific subtropics, there appear to be some negative anomalies too. The low-index anomaly composites yield an almost similar, but inverted, pattern, where positive anomalies are now negative and vice versa. However, although the anomaly patterns’ sign inversion is distinct over the South Atlantic and Indian Oceans, the patterns over the South Pacific vary.

This is further confirmed when the 300-hPa anomaly fields are Fourier decomposed into waves 1–3 components (Fig. 5). At high latitudes, waves 1 and 2 are of the same order and together explain most of the variance. Tropospheric wave 1 points to a 120° eastward phase rotation (Figs. 5a and 5b), almost in antiphase with stratospheric wave 1, in agreement with Quintanar and Mechoso (1995). Wave 2 shows a virtual flip-flop behavior, with negative (positive) anomalies becoming positive (negative) (Figs. 5c and 5d). Wave 3 also shows a flip-flop behavior. Thus, these two waves also exhibit an interannual phase variability similar to wave 1 (Figs. 5e and 5f). These results suggest that the observed composite differences between low- and high-phase index values at least span the lower stratosphere (LS) and upper troposphere (UT).

Inspection of the stratosphere’s zonal asymmetries (not shown) shows results in agreement with Grytsai et al. (2007) for TOC and Agosta and Canziani (2010) for GPH. These authors conclude that the negative lobe in the zonal asymmetry, over the South Atlantic, undergoes the largest longitudinal variations, while the
FIG. 4. Composite geopotential anomalies for 30-PW (a),(c),(e) high- and (b),(d),(f) low-index months at (a),(b) 30 hPa (CI = 500 m), (c),(d) 100 hPa (CI = 200 m), and (e),(f) 300 hPa (CI = 150 m). Dark/darker (lighter/light) gray-shaded areas correspond to statistically significant (90%/95%) positive (negative) temporal anomalies as given by an unequal variance $t$ test.
FIG. 5. Fourier harmonic decomposition of the 300-hPa GPH anomalies for (a),(c),(e) high and (b),(d),(f) low 30-PW indices into waves (a),(b) 1 ($k = 1$), (c),(d) 2 ($k = 2$), and (e),(f) 3 ($k = 3$). Dark/darker (lighter/light) gray-shaded areas correspond to weak/strong positive (negative) GPH anomalies (CI = 50 m), not statistically significant.
positive lobe, over the South Pacific, remains virtually over the same location, albeit with some changes in its size. This implies that the lobes observed here behave as wave packets, including at least waves 1–3 components, with QSW1 becoming more prominent with height.

**b. QSW1 phase variations: Observed UT–LS linkages**

Hio and Hirota (2002) argued, through a PC analysis and subsequent zonal-mean vertical EP flux compositing, for the existence of links between stratospheric QSW1 and tropospheric synoptic-scale processes, though they did not establish how this could occur. They showed that the composited vertical EP flux associated with the longitudinal changes in GPH QSW1 at 10 hPa (their first EOF) appear to be linked to the contribution of tropospheric synoptic-scale processes below 200 hPa, while QSW1 interannual amplitude variability (their second EOF) would be distinctly linked to planetary waves 1–3 variability in the troposphere and lower stratosphere. However, the physical links between tropospheric synoptic-scale processes and the observed longitudinal changes remained unclear.

They did observe differences in the composite zonal wind field at 300 hPa, which presented a double-jet structure over the Indian Ocean and western South Pacific for negative values of their first EOF. Furthermore, their analysis of poleward momentum flux at 400 hPa for waves 4–12, which can be linked to the SH storm-track pattern, shows differences between their positive and negative first EOF values over Australia, Oceania, and the central South Pacific. However, in an EOF analysis of the vertical structure of tropospheric zonal winds, based on winds zonally averaged over the Eastern Hemisphere (0°–180°) east of the prime meridian, with the resulting EOFs representing single- and double-jet vertical structures, they were unable to reproduce, through compositing, the two distinct stratospheric structures, suggesting a far more complex interaction between the troposphere and stratosphere.

Figure 6 shows the zonal wind composite fields in the upper troposphere (300 hPa) and lower stratosphere (100 and 30 hPa) over the SH. Dark/darker (lighter/light) shading implies statistically significant (at the 90%/95% level) enhanced (weakened) zonal winds with respect to the sample climatology. The stratospheric polar vortex and the tropospheric polar and subtropical jets composites show significant differences between high- and low-value indices.

At 30 hPa the significant areas represent the QSW1 anomaly pattern together with the associated differences between high- and low-index composites. For low index values, the polar vortex jet, around the edge of Antarctica, appears to extend eastward into the Pacific sector of the Southern Ocean, well beyond the date line (180°). Polar vortex jet strength values over the Indian Ocean are also stronger. However, for high index values, the composite jet extends with values above 45 m s⁻¹ across the South Atlantic. At midlatitudes the same QSW1 structure can be observed for the regions with significant wind enhancement (decrease). These significant anomalies extend at least to 100 hPa, where they are now found mostly over the Indian Ocean and in the vicinity of southern South America. Note again the eastward displacement, at 100 hPa with respect to 30 hPa, of the significant anomaly areas over the Western Hemisphere.

At 300 hPa the areas with significant anomalies are much smaller, limited to a few minor areas. Though not statistically significant at least at the 95% level, the double-jet structure reported in Hio and Hirota (2002) can be seen in the composite of the low index values, in particular over the western South Pacific.

To evaluate whether the distinct maximum zonal wind anomalies for high and low index values, and hence the location of the mid- to polar latitude ozone trough can be linked with synoptic-scale perturbations associated with active tropospheric jets as suggested by Hio and Hirota (2002), the daily probability p of jet stream occurrence is calculated for high- and low-index years. The location of a daily jet stream at 300 hPa is defined by local daily zonal winds equal to or in excess of 25 m s⁻¹ and will be referred to as “jet.” Similarly, at 30 hPa the jet is defined for zonal wind values at or above 40 m s⁻¹. The aim for such a probability distribution construction is to verify whether the tropospheric double-jet structure observed in the monthly-mean zonal winds for low-index composites can be ascribed to daily double-jet occurrences, one at mid- to subtropical latitudes and the other at subpolar to lower latitudes. Similarly, the probability maps at 30 hPa can provide insight into the occurrence of peak wind values in the polar vortex jet and possible changes between high and low index values.

For high index values, the daily probability of jet occurrence at 300 hPa (Fig. 7c) is highest (p > 0.70) over and along the equatorward sector of the subtropical branch of the climatological jet stream over the Indian and western Pacific Oceans. Another region of peak probability values can be found extending from the eastern South Atlantic toward the Indian Ocean midlatitudes. Finally, another small region of high probability can be found over the Rio de la Plata basin (~34°S, 60°W).

Daily probabilities, greater than 0.8, are observed over larger areas, both in the subtropics and at mid-latitudes (Fig. 7d), for low index values. The midlatitude
Fig. 6. Zonal wind $U$ (CI = 5 m s$^{-1}$) composite fields for (a),(c),(e) high and (b),(d),(f) low 30-PW indices at (a),(b) 30, (c),(d) 100, and (e),(f) 300 hPa. Dark/darker (lighter/light) gray-shaded areas correspond to statistically significant (90%/95%) enhanced (decreased) $U$ composites as given by an unequal variance $t$ test.
region extends from southwest of South Africa as far as the date line. There is an enhancement, both in probability and longitude range, of the subtropical jet both to the east and west of Australia and over the Río de la Plata basin.

Hence, for high index values, there is only a small longitude sector to the west of Australia where there is less than 50% probability of a double-jet structure. Conversely, for low-index-value situations, the probability of a double jet in that region is greater than 64% (~20 days) over that region and slightly more than 50% (~16 days) north of New Zealand (~30°S, 170°E–140°W). Hence, in any given October day with a low 30-PW index value, there is a reasonably high probability that the double-jet structure is observed around the Australia/Oceania sector, particularly in the sector’s eastern half.

At the 30-hPa level, for high index values (Fig. 7a), over the Atlantic and Indian Oceans, there is a high probability of a strong polar vortex jet, greater than 40 m s⁻¹ (p ~ 0.9), that spirals south from Tierra del Fuego/Malvinas (50°–55°S, 60°–70°W) toward Antarctica, over Wilkes Land (65°–70°S, 170°E–180°E). During low index values (Fig. 7b), the high-probability region, which is further enhanced over the Atlantic–Indian Ocean transition, extends as far as the central South Pacific sector, again spiraling over Antarctica and even farther into the Ross Sea area (65°–70°S, 150°–170°W).

During low-index months, the probability of a simultaneous occurrence of a double jet in the upper troposphere over the eastern Indian Ocean together with a very strong polar vortex jet over the western Indian Ocean sector is greater than 57% (~18 days), that is,
more than half the days for a given low-index October. Conversely, if we consider the probability of a double-tropospheric jet over Oceania/South Pacific together with the occurrence of a strong polar vortex jet either over the Indian Ocean or over Wilkes Land, the probability is somewhat less than 50% (~14 days).

For high index values, the probability of a single, strong tropospheric jet over the western Indian Ocean or to the east of Australia, together with a strong polar vortex jet is somewhat less than 60% (~19 days). Finally, the probability of a double-jet occurrence during high index values together with a strong polar vortex jet is less than 39% (~12 days). Hence, there is a 6–7-day difference in the occurrence of tropospheric double jets together with a strong stratospheric jet, between high- and low-index Octobers, with a preferred occurrence during low-index months.

From these comparisons, it can be conclude that during low 30-PW values, there is an overall higher frequency of occurrence of double daily jets in the troposphere. Furthermore, the polar stratospheric jet, in particular, is clearly active at daily scales between 50° and 55°S and 10° and 90°E over the Indian Ocean and can be associated with the double-jet occurrence in that region. These results are in agreement with Hio and Hirota (2002), and thus they can explain why when these authors carried out an Eastern Hemisphere zonal average analysis, they were unable to fully account for the observed relationships between the stratosphere and troposphere—that is, since they were unable to reproduce the relationship when extending the Eastern Hemisphere mean tropospheric EOF results to the stratosphere. The present results show that the sector where there is a comparatively higher probability of tropospheric double-jet occurrence together with a strong stratospheric polar vortex jet is limited to less than a quadrant over the western Indian Ocean. Because of the differences in the occurrence probabilities between the two quadrants, the use of zonal averages over the Eastern Hemisphere, as in Hio and Hirota (2002), would tend to misrepresent the results when the resulting tropospheric EOFs are used as indices. Note that a similar analysis using VARMAX enhances these results with greater differences between the upper and lower VARMAX quartiles (Agosta and Canziani 2010).

The synoptic-scale $Z^2$ analysis at 300 hPa—that is, for perturbations with periods 2–8 days—also shows differences between low and high index values. The anomaly fields with respect to the 1979–2007 $Z^2$ climatology yield changes that are in good agreement with the behavior of the zonal wind composites (Fig. 6) and daily jet probability fields (Fig. 7), as described above. For high index values, wave activity enhancements are observed at midlatitudes over the South Atlantic, east of southern South America, extending into the western Indian Ocean (Fig. 8a). Wave activity decreases are found over South Pacific midlatitudes to the west of southern South America and at some locations around Antarctica. The latter decrease could be associated with a contracted tropospheric polar jet. During low index values, almost the inverse anomaly pattern is observed: over the South Atlantic and the Indian Ocean, transient activity appears to be poleward displaced, while over the South Pacific, an equatorward movement of positive wave activity anomalies is present (Fig. 8c).

The Eady growth rate (EGR; Hoskins and Valdes 1990) is used as a measure of the baroclinicity of the mean flow during high and low index values, defined as

$$EGR = 0.31 \left( \frac{f}{N} \right) \left[ \frac{\partial U}{\partial z} \right]$$

where $f$ and $N$ are the same as in Eq. (3) and $\partial U/\partial z$ is the vertical shear of the zonal wind.

It is evident that the areas of enhanced/decreased synoptic-scale activity correspond fairly well with the EGR anomalies (Figs. 8b and 8d). This means that those processes are distinct between high and low index values: changes in the vertical tropospheric shear, changes in the meridional midlatitude sea surface temperature gradient directly generating baroclinicity because of the thermal wind relation, and/or tropical-heating-inducing upper-tropospheric Rossby waves that modify the location of stationary waves and, therefore, upper-tropospheric jets (Inatsu and Hoskins 2004). Eventually, a downward control from lower stratosphere (Haynes et al. 1991) may be involved, although it is not easy to evaluate using localized diagnostic tools.

Following Takaya and Nakamura (2001) and Nishii and Nakamura (2004), $W$ flux was calculated for the anomalous quasi-stationary waves in the UT–LS region, for high- and low-index composite time-mean anomalies. The corresponding horizontal $W$ flux vectors and their vertical components are shown in Fig. 9 for 30, 100, and 300 hPa.

During high-index-value situations (Figs. 9a, 9c, and 9e), there is a region of strong vertical $W$ flux upward propagation from the troposphere into the stratosphere in the vicinity of the Drake Strait (55°–60°S, 50°–60°W), between southern South America and the Antarctic Peninsula. This vertical wave activity injection appears near the beginning of the region of stronger polar vortex jet at 100 hPa and extends almost vertically, up to 30 hPa at least. Furthermore, at each stratospheric level, $W$ flux vectors show that the wave activity propagates eastward with height along the polar vortex jet.
30 and 100 hPa, there appears to be a region of strong downward wave activity flux descending toward the tropopause near 90°E, which, in the upper troposphere, is limited in magnitude and extent. Conversely, in the upper troposphere (Fig. 9e), the W flux appears to propagate toward the Drake Strait region from lower latitudes in the central South Pacific, near 10°–20°S, 90°–110°W.

For low index values (Figs. 9b, 9d, and 9f), the W flux vertical component is weaker at all levels. Furthermore, the region of localized upward propagation appears to extend from the Weddell Sea/Malvinas region (55°–60°S, 50°–60°W) beyond the Greenwich meridian at mid- to subpolar latitudes. A region of vertical propagation over southern Australia, as well as at other isolated subtropical regions, does not extend into the lower stratosphere. The displacement of the vertical propagation areas appears to coincide, in the lower stratosphere, with the origin of the eastward-displaced polar vortex jet for low index values. The regions of vertically propagating wave packets appear to be approximately similar—or even decreasing with height—at the stratospheric levels considered here, that is, with limited zonal eastward propagation. Note that there is a region of downward propagation at 100 hPa (Fig. 9d) to the northeast of the trailing edge of the polar vortex edge near 65°–70°S, 110°W that appears to extend into the upper troposphere. A horizontal W flux at 300 hPa (Fig. 9e) appears to propagate over the South Atlantic from the subtropical region near the southern Brazil coast (~25°–30°S, 50°W) to the Weddell Sea/Malvinas region. There also appears to be a meridional W flux due south of South Africa contributing to the ascent region. Overall, the W flux appears to be weaker for low-index-value situations.

For both high and low indices, there appears to be a region of vertically propagating wave packets over Wilkes Land (~70°S, 170°E–180°) at 100 hPa that does not have links with similar processes in the upper troposphere, yet this vertical QSW flux injection appears to weaken at higher levels.
Fig. 9. QSW anomalies defined by the geopotential (CI = 500 m at 30 hPa, 200 m at 100 hPa, and 150 m at 300 hPa) and associated horizontal W flux (arrows, 2 m² s⁻²) superimposed on color contours for the W flux vertical component (CI = 1 × 10⁻³ m² s⁻²). Composites for (a),(c),(e) high and (b),(d),(f) low 30-PW indices at (a),(b) 30, (c),(d) 100, and (e),(f) 300 hPa.
4. Discussion

According to NN05, a Rossby wave train can have upward/downward propagation through the tropopause at submonthly scales through the following mechanism. They argue that such a process can occur at locations where tropospheric circulation anomalies amplify, associated with incoming quasi-stationary Rossby wave trains together with or due to the presence of synoptic-scale transient eddies. The wave activity can then propagate upward into the lower stratosphere along a waveguide, if present in the vicinity of the polar vortex jet. Once in the lower stratosphere, the propagating wave activity can result in a circulation anomaly of opposite sign in the stratosphere with respect to the upper troposphere. If the waveguide extends along the polar vortex jet, then the wave activity extends downstream along the waveguide, to the east of the ascent region. When the polar vortex jet weakens, or vanishes, the waveguide in the jet exit region weakens, a downward vertical propagation can occur, and part of the wave activity propagates downward through the tropopause, thus contributing to the downstream formation of tropospheric circulation anomalies of opposite sign. In principle, this can be viewed as a refraction process through which the lower wavenumber waves 1 and possibly 2 can propagate into the stratosphere, while the higher wavenumber waves are finally refracted by a substantial westerly vertical shear back into the troposphere, as if the polar vortex jet acted as a “prism.”

The above results (Figs. 4–9), considered together, can provide insights into the QSW1 high- and low-index situations, its relation with wave activity propagation and UT–LS coupling. In this analysis, the concepts provided by NN05, which they used for submonthly quasi-stationary anomalies, are applied to composite (climatic) monthly anomalies. The upward/downward QSW propagation mechanism is apparent for high index values. In this situation, the main vertical propagation region over the Drake Strait is located in the transition between a large anticyclonic anomaly, spanning the South Pacific, and a large cyclonic anomaly straddling the Weddell Sea and western South Atlantic. This region is coincident with the entrance region of the 100-hPa polar vortex jet with zonal wind values above 35 m s$^{-1}$. The quasi-stationary wave packets propagate vertically and horizontally (eastward) in the lower stratosphere until the jet begins to weaken in the jet exit region north of Prydz Bay, Antarctica, near 75°S, 90°E. A negative GPH anomaly can be observed downstream of the downward propagation region, extending between Antarctica and Australia (Figs. 9c and 9e), of opposite sign to the one in the lower stratosphere. This GPH anomaly agrees with the single tropospheric jet situation described above (Figs. 6e and 7c).

Likewise, for low index values, the main, albeit weaker, upward W flux located eastward of Drake Strait at 300 hPa maximizes in the lower stratosphere, at 100 hPa, probably generating an opposite sign anomaly there. However, in this case, the polar jet entrance is found some distance away, farther east, thus limiting the eastward propagation of a strong horizontal W flux. Similarly, another source region with HF can be observed to the northeast of the Rio de la Plata estuary (30°S, 50°W). A sizeable, downward W flux region to the north of Wilkes Land, Antarctica (75°S, 180°), is observed at 100 hPa, associated with the equatorward edge of the western anticyclonic anomaly with the positive lobe of QSW1. The downward flux injection reaches 300 hPa, generating a circulation anomaly of opposite sign downstream.

Figures 10a and 10b show QSW anomalies as observed in streamfunctions for high and low 30-PW index values, respectively, together with their associated W flux and their 3D divergence. Inspection of the wave flux, for high index values, shows that the quasi-stationary wave packets emanate from a source region close to Easter Island, near 10°–30°S, 100°W (Fig. 11a), where the 3D divergence calculation of the W vector highlights a source region in the upper troposphere, probably associated with an extratropical expansion of the tropical–extratropical divergence zones. The anomalous potential velocity and divergent wind anomalies over the Pacific support the hypothesis (Fig. 11a), showing a remarkable negative potential velocity anomaly center at 20°S, 20°W. This negative potential velocity anomaly appears to be associated with an extended region of negative anomalies projecting itself into the region, at least from the equatorial central Pacific. This region is downstream of positive SSTs anomalies extending from the equator to the east of Indonesia (figure not shown). The vertical motion anomalies (Fig. 11b) show a pattern of ascent–descent in correspondence with potential velocity and streamfunction anomalies at tropical to mid-latitudes over the Pacific near South America’s western coast. Thus, in association with the Coriolis parameter increase from the tropics toward the subtropical latitudes, these anomalies can result in significant vorticity anomaly sources in the subtropical branches of the regionally observed anomalous Hadley circulation. Under the presence of mean divergent wind anomalies, the region becomes highly effective as Rossby wave sources (RWSs), which are able to excite extratropical wave teleconnections (Sardeshmukh and Hoskins 1988; Qin and Robinson 1993). Figure 11c shows that there is a positive anomalous Rossby wave generation centered at...
Fig. 10. 300-hPa QSW anomalies defined by the streamfunction (CI = $10 \times 10^{-5}$ m$^2$ s$^{-1}$) and associated horizontal W flux (arrow, 2 m$^2$ s$^{-1}$) superimposed on color contours for the divergence of W (CI = 0.510$^{-6}$ m s$^{-2}$). Composites for (a) high and (b) low 30-PW indices.

25°S, 100°W because of the planetary vorticity divergence stretching term of RWS [term S1 in Eq. (1) by Qin and Robinson (1993), expanded for seasonal mean anomalies by Mo and Rasmusson (1993)], which is located slightly south of the negative potential velocity anomaly, where the anomalous divergent wind is stronger. Other W flux source regions appear along the southern tip of South America and the southeastern region of the Antarctic Peninsula (Fig. 10), probably as a result of the interaction between the quasi-stationary wave packets and the HF transient activity observed there (Fig. 8a), with a net increase in HF activity on the lee side, over the South Atlantic midlatitudes to the east of the Malvinas Islands (54°–55°S, 20°–50°W). In other words, the associated anticyclonic and cyclonic circulation associated with the QSW packets propagating from the southeastern Pacific toward the South Atlantic appear to inhibit transient wave activity to the southwest of South America and to enhance it farther east.

For low index values, weaker horizontal W fluxes seem to emanate from source regions over subtropical South America and midlatitude South Pacific and to propagate toward high-latitude South Atlantic, where another positive W flux divergence appears near 60°, 30°W (Fig. 10b). The anomalous QSW sources in these regions seem to be due to synoptic-scale activity that is significantly high, during the low-index-value situations, over the southeastern Pacific off the coast of southern South America (Fig. 8c). Note that this region of enhanced HF activity coincides with the cyclonic anomaly associated with the downward propagation of the W flux from the lower stratosphere.

The use of diagnostic tools such as the W flux does not allow the determination of the mechanisms triggering the onset of the observed QSW variability for high and low 30-PW index values. For low values, it would appear that the quasi-stationary anomalies, the resulting W flux spatial distribution, and the HF activity of the stratospheric–tropospheric coupled system result in a self-sustaining mode, for the given mean flow at mid- to high latitudes. Conversely, for high values, there appears to be wave energy propagation from the tropical central Pacific toward higher latitudes (probably associated with anomalous deep convection there) that generates tropospheric
anomalies in the vicinity of the Drake Strait, propagating upward, and generating or sustaining the stratospheric QSW1 anomaly. Both the subsequent eastward propagation along the polar vortex jet and the downward propagation over the Indian Ocean result in a prominent upper-tropospheric anomaly modulating the mean flow.

At this stage, it is not clear to what extent the winter circulation preconditioning contributes to the establishment of either low- or high-index-value situations.

5. Conclusions

The present study, through the analysis of TOC interannual variability, has found links between the SH stratosphere and troposphere spring dynamics. After statistically filtering all the known global TOC variability predictors, it is found that the residual $\varepsilon$TOC interannual variability is dynamically coupled to the stratospheric QSW1 phase behavior. Thus, a QSW1 phase variability index, such as 30 PW, can be an additional predictor in a multiregression analysis of TOC, at least over the SH.

Inspection of both $\varepsilon$TOC and stratospheric GPH anomalies show that the interannual phase variability can be linked to statistically different mean flow patterns, as derived from the upper- and lower-quartile values of both 30 PW and VARMAX. High index values correspond to a westward rotation of the midlatitude ozone trough and of the zonally asymmetric QSW1 phase, low values highlight their eastward rotation. While at 30 hPa, the observed patterns are dominated by QSW1, whereas at 100 hPa and even more so at 300 hPa, the observed differences between high and low index values show that the anomaly changes are also linked at least to quasi-stationary waves 2 and 3. These higher-order quasi-stationary waves also share the interannual phase inversion behavior. Hence, the observed interannual phase variability in midlatitude TOC and stratospheric QSW1 appear as distinct evidence of a significant interannual variability pattern across the SH troposphere and lower stratosphere.

The composite circulation patterns can be associated with a predominantly single poleward tropospheric jet for the upper-quartile values and a predominantly double-jet structure for the lower-quartile values, in agreement with Hio and Hirota (2002). For low index values, there is both a higher daily probability of double-jet occurrence in the troposphere and a stronger stratospheric jet. These jet structures and their daily behavior arise together with significantly distinct synoptic-scale activity anomalies over the SH mid- to high latitudes for the high and low index values.

The observed tropospheric anomalies, associated with the westward and eastward stratospheric QSW1 phase rotations, show active quasi-stationary waves through the $W$ flux diagnosis (NN05). For low index values, associated with stratospheric eastward phase rotations, the quasi-stationary waves result in a self-sustaining state of the stratospheric–tropospheric coupled system. For high index values associated with westward phase rotation, the overall mid- to high-latitude state appears to be associated with wave energy propagation from the tropical central Pacific toward higher latitudes.

**FIG. 11.** (a) 200-hPa potential velocity anomalies ($CI = 0.15 \times 10^{-6}$ m$^2$ s$^{-1}$) and anomalous divergent winds (arrows, 1 m s$^{-1}$). (b) 500-hPa vertical motion anomalies (omega, $CI = 7.5 \times 10^{-3}$ Pa s$^{-1}$). (c) Anomalous RWS due to the presence of mean divergent wind anomalies on a subtropical region with relevant planetary vorticity gradient during high 30-PW values ($CI = 0.1 \times 10^{-10}$ s$^{-2}$). For the vertical motion anomalies, dark/darker (lighter/light) gray-shaded areas correspond to statistically significant (90%/95%) positive (negative) anomalies as given by an unequal variance $t$ test.
The present analysis shows that during the austral spring, there are significant interactions/coupling between the ozone layer, the troposphere, and the stratosphere that can be traced by the phase changes in TOC and QSW1 in the stratosphere. Such changes and troposphere–stratosphere interactions are linked to both the upward and downward propagation of quasi-stationary wave anomalies. These occur at well-defined locations, for example, there is a preferential upward W flux propagation into the stratosphere in the vicinity of the Drake Strait, near 60°S, 40–60°W, that is enhanced during high index values. Downward propagation through the tropopause may preferentially occur over Indian Ocean midlatitudes at those times and over the central South Pacific during low index values. Furthermore, the tropospheric quasi-stationary wave anomalies link with the synoptic-scale wave activity through complex two-way interactions.

Previous tropospheric–stratospheric interaction studies over the SH have focused on the zonal-mean relationships, within the conceptual SAM framework, during the end of spring and early summer, when downward propagation of October stratospheric anomalies impact upon tropospheric processes, for example, over the Antarctic Peninsula. The current study argues that significant interactions can occur during spring itself. Further studies need to address possible links with the state of the stratosphere and troposphere during winter, and a possible preconditioning of the spring behavior. Work is under way to understand the processes underlying the eastward migration in the TOC midlatitude structure noted here and previously by various authors, as well as in both the stratosphere and the troposphere (Huth and Canziani 2003; Grytsai et al. 2007; Malanca et al. 2005; Yuchechen et al. 2007; Barrucand et al. 2008).

The current observational results highlight the need to extend the research of the troposphere–stratospheric coupling processes to the analysis and modeling of zonally nonuniform time-varying interactions.

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REFERENCES


