Dynamics of ENSO-driven stratosphere-to-troposphere transport of ozone over North America

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Abstract. The El Niño-Southern Oscillation (ENSO) is known to modulate the strength and frequency of stratosphere-to-troposphere transport (STT) of ozone over the Pacific-North American region during late winter to early summer. Dynamical processes that have been proposed to account for this variability include: variations in the amount of ozone in the lowermost stratosphere that is available for STT, and tropospheric circulation-related variations in the frequency and geographic distribution of individual STT events.

Here we use a large ensemble of Whole Atmosphere Community Climate Model (WACCM) simulations (forced by sea-surface temperature (SST) boundary conditions consistent with each phase of ENSO) to show that variability in lower stratospheric ozone and shifts in the Pacific tropospheric jet constructively contribute to the amount of STT of ozone in the North American region during both ENSO phases. In terms of stratospheric variability, ENSO drives ozone anomalies resembling the Pacific-North American teleconnection pattern that span much of the lower stratosphere below 50 hPa. These ozone anomalies, which dominate over other ENSO-driven changes in the Brewer-Dobson circulation (including changes due to both the stratospheric residual circulation and quasi-isentropic mixing), strongly modulate the amount of ozone available for STT transport. As a result, during late winter (February-March), the stratospheric ozone response to the teleconnections constructively reinforces anomalous ENSO-jet-driven STT of ozone. However, as ENSO forcing weakens as spring progresses into summer (April-June), the direct effects of the ENSO-jet-driven STT transport weaken. Nevertheless, the residual impacts of the teleconnections on the amount of ozone in the lower stratosphere persist, and these anomalies in turn continue to cause anomalous STT of ozone. These results should prove helpful for interpreting the utility of ENSO as a subseasonal predictor of both free-tropospheric ozone and the probability of stratospheric ozone intrusion events that may cause exceedances in surface air quality standards.

1 Introduction

Ozone transported from the stratosphere contributes to the North American background (NAB) ozone concentration in the free troposphere (Fiore et al., 2014; Cooper et al., 2015; Young et al., 2018) and to surface ozone exceedance events that affect human health (Fiore et al., 2014; Cooper et al., 2015; Young et al., 2018; Zhang et al. 2020; Langford et al. 2022).

Unfortunately, estimating the stratospheric contribution to surface exceedances and the NAB is quite complex because atmospheric internal variability and low-frequency climate modes (e.g., the El Niño-Southern Oscillation and the quasi-biennial oscillation) combine to drive significant subseasonal-to-seasonal variations in stratosphere-to-troposphere transport.

On subseasonal-to-seasonal timescales, variability in stratosphere-to-troposphere transport (STT) is primarily modulated via two dynamical processes (Albers et al. 2018 and references therein): variations in the amount of ozone in the lowermost stratosphere available for STT, and tropospheric circulation-related variations in the frequency, depth, and geographic distribution of individual STT events (e.g., Breeden et al. 2021). The El Niño-Southern Oscillation (ENSO) is thought to modulate both processes, but unfortunately, prior research yields somewhat conflicting results. For example, on hemispheric spatial scales, Neu et al. (2014) used Tropospheric Emission Spectrometer (TES) and Microwave Limb Sounder (MLS) data (2005-2010) and suggested that the warm phase of ENSO accelerates the Brewer-Dobson circulation (BDC), which leads to more ozone in the lowermost midlatitude stratosphere, and subsequently causes an increase in STT of ozone into the midlatitude troposphere (see also, García-Herrera et al. 2006, Calvo et al. 2010, and Simpson et al. 2011). This view is supported by Zeng and Pyle (2005) who found a positive correlation between ENSO and global STT for 1990-2001. On the other hand, Hsu and Prather (2009) find a weak correlation between hemispheric or global STT and ENSO during a slightly later period (2001-2005). If more localized spatial scales are considered, some conflicting results remain. For example, Lin et al. (2015) suggested that La Niña (1990-2012) shifts the Pacific storm track northward and increases its variability, leading to more frequent deep stratospheric intrusions transporting stratospheric ozone into the lower troposphere over western North America. However, Langford (1999) found that El Niño (1993-1998) extends the subtropical jet eastward, driving transverse circulations at the nose of the jet that can also increase ozone transport into the middle and upper troposphere over western North America.

It is difficult to discern whether the aforementioned results appear to conflict because of the relatively short data records used, or whether both phases of ENSO can potentially increase STT of ozone, but with some sensitivity to the specific geographic region. Here we address both possibilities by using a large ensemble of Whole Atmosphere Community Climate Model (WACCM) simulations to quantify how ENSO modulates subseasonal variations in STT of ozone over the Pacific-North American region. The WACCM simulations reveal that ENSO-driven changes in STT are highly dependent on the time of year and geography, with stratospheric and tropospheric processes combining to increase STT in one region while decreasing STT in a second region during the same ENSO phase. Moreover, the large number of simulated ENSO years, forced by the same sea surface temperature patterns in order to reduce the influence of ENSO diversity, allows quantification of the relative importance of ENSO-induced changes in the midlatitude (Lin et al. 2015) and subtropical (Langford 1999) jets for modulating STT of ozone. And in contrast to some previous studies suggesting that the ~~BDC is~~residual circulation and isentropic mixing aspects of the BDC are the primary mediating ~~link~~links between ENSO and stratospheric changes in STT, our results suggest that it is ozone teleconnections that are most important, a finding which is consistent with Zhang et al. 2015 and Olsen et al. 2016.

Ozone teleconnections, first recognized by Reed (1950) (see also Schoeberl and Krueger 1983, ~~Mote et al. 1991, Stephenson and Royer 1995, Wang et al. 2011~~ and Salby and Callaghan 1993), result from vertical motion and horizontal advection induced by planetary wave geopotential height perturbations that are associated with opposite signed ozone perturbations (i.e., a positive geopotential height anomaly is associated with a negative ozone anomaly). In the WACCM simulations, ENSO drives ozone anomalies resembling the Pacific-North American teleconnection pattern that extend from the lowermost stratosphere to at least 50 hPa in height. Here we provide a detailed analysis using a stratospheric ozone tracer (O$_3$S from Tilmes et al. 2016), to demonstrate how the jet shifts and ozone teleconnections patterns caused by ENSO constructively contribute to drive changes in STT of ozone over North America.

Section 2 outlines the WACCM simulations used, Section 3 details how stratospheric and tropospheric processes constructively reinforce anomalous STT during different times of the seasonal cycle and for different portions of Pacific-North American region. A discussion of the implications of our results for subseasonal prediction of STT of ozone is contained in Section 4.

2 Climate model simulations

2.1 WACCM simulations

Simulations were created using the National Center for Atmospheric Research Whole Atmosphere Community Climate Model (WACCM) version 4 (Mills et al. 2017). WACCM has fully interactive chemistry in the middle atmosphere, which includes a stratospheric ozone tracer (O$_3$S) that evolves via full chemistry in the stratosphere and then decays at a tropospheric chemistry rate once it crosses the tropopause (the version of WACCM we use here includes ozone removal via tropospheric dry deposition). The O$_3$S tracer should be interpreted to represent an upper-bound of the stratospheric contribution to a stratosphere-to-troposphere ozone fold, in large part because it is missing some tropospheric chemistry that would likely decrease its tropospheric chemical lifetime (Emmons et al. 2003). WACCM has a limited representation of tropospheric chemistry, but simulates background tropospheric interannual ozone variability quite well (Hess et al. 2015). The model extends to ~140 km with 70 vertical levels, and a horizontal resolution of 1.9 degrees latitude by 2.5 degrees longitude.

We conducted two sets of "time-slice" simulations, one set each for El Niño and La Niña conditions. Each simulation is 60 years long, with the first 10 years of each simulation used as model 'spin-up' time and subsequently discarded. The simulations are forced via SST composites for each ENSO phase, which were created by averaging over all El Niño and La Niña events (defined as Niño 3.4>1 standard deviation from the March long-term mean) using Hadley Centre Global Sea Ice and Sea Surface Temperature data (HadISST2, years 1950-2008; see https://climatedataguide.ucar.edu/climate-data/nino-sst-indices-nino-12-3-34-4-oni-and-tni for a description of the Niño 3.4 index). SST anomalies evolve over a two-year cycle, growing from zero SST anomaly in January of year one, to a peak anomaly in January of year two, and then back to zero anomaly by the end of December of year 2. Thus, over the course of a 60-year simulation, the cycle repeats itself thirty times. The SSTs anomalies are tapered in space via a half-cosine weighting function so that there is no SST anomaly poleward of +/-

25° latitude (see Supplementary Fig. S1 for complete cycle). The ENSO SST composite was constructed to reproduce the largest March SST anomalies in the observational record, thus allowing us to establish an upper bound on the potential effects of ENSO-related control of spring season STT. However, even when constructed in this way, the SST anomalies still peak in January of the second year of the two-year ENSO cycle. All composites are computed for year two of the two-year ENSO cycle, which corresponds to the mature and decaying portion of the ENSO cycle. In all cases, radiative forcings, tropospheric emission precursors, and volcanic sulphates were set to preindustrial (year 1850) levels, to avoid major influence from greenhouse gases or ozone-depleting substances. The simulations contain no tropical quasi-biennial oscillation, and for solar and geomagnetic parameters, a solar cycle average was used. WACCM anomalies are created by subtracting a climatology that is created from the average over all ensembles (El Niño and La Niña simulations, i.e., 100-years of data).

Eliassen-Palm flux vectors and the stratospheric residual circulation ($\bar{v}^*, \bar{w}^*$) are calculated on daily timescales using the Transformed Eulerian Mean formulation in spherical coordinates as defined in Andrews et al. (1987). Eddy kinetic energy ($1/2[u'^2 + v'^2]$) is calculated from daily data that is band-passed filtered. For eddy kinetic energy (EKE) near the tropopause, a 2-10 day filter is applied to highlight synoptic timescale variability, while for EKE in the upper stratosphere, a ~~slightly~~ broader 2-120 day filter is applied to capture synoptic and more slowly evolving planetary scale eddy variability (e.g., Albers et al. 2016).

In the results, most figures are shown for both ENSO phases (or the difference of El Niño minus La Niña) or are included in the Supplement. In all cases, the La Niña anomalies are essentially identical to the El Niño anomalies, but opposite in sign, which is due to the symmetry in the prescribed Niño 3.4-based SST forcing for both ENSO phases.

3 Results

The climatological seasonal cycle of Northern Hemisphere (and North American) extratropical and high latitude STT of ozone in the WACCM time slice simulations is consistent with observations (e.g., Lefohn et al. 2001, Albers et al. 2018), with notable stratospheric ozone (O_3S) present in the lower troposphere beginning in December, O_3S peaking in March and April, and then largely decreasing so that negligible O_3S ~~anomalies are~~is present by mid-summer (Supplement Fig. S2). Interannual variability in high latitude total column ozone can largely be accounted for by variability in the wave-driven BDC, with stratospheric chemistry accounting for less than 20% of the interannual variance (Fusco and Salby 1999, Salby and Callaghan 2002, Weber et al. 2011). The vertical distribution of ozone within the column is controlled by a combination of the BDC (including both advective and eddy transport) in the stratosphere during winter and stratospheric intrusions along the tropopause during spring and early summer. Within the stratosphere, extratropical stratospheric planetary wave driving peaks in Northern Hemisphere winter (Charney ~~&~~and Drazin, 1961; Randel et al., 2002), which increases ozone transport along the deep branch of the BDC leading to a seasonal build-up of ozone in the high latitude lowermost stratosphere (Konopka et al. 2015, Ploeger and Birner, 2016, Ray et al. 1999, Bönisch et al. 2009, Butchart 2014, Hegglin and Shepherd 2007). As spring proceeds into early summer, eddy activity (and hence stratospheric intrusion frequency) along the extratropical tropopause increases in vigor (Breeden et al. 2021 and references therein), which leads to a "flushing" of ozone from the stratosphere into the troposphere that accounts

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130 for the seasonal peak in STT of ozone over North America (James et al. 2003, Lefohn et al. 2001, Škerlak et al. 2014, Terao et al. 2008, Albers et al. 2018).

3.1 Stratospheric circulation

135 During late winter, El Niño accelerates the BDC along both the shallow and deep branches (Fig. 1a; the qualitatively, but opposite signed La Niña anomalies are shown Fig. 1c). Along the shallow branch (below roughly 70 hPa), there is strong upwelling between 0°-10° N and strong downwelling between 10°-30° N, which is driven by a combination of planetary and gravity waves in the subtropics as documented in previous studies (Garcia-Herrera et al. 2006, Calvo et al. 2010, Simpson et al. 2011, Diallo et al. 2019). Along the deep branch, there is enhanced poleward flow above 10 hPa (not shown) and enhanced

140 downwelling poleward of 30° N (primarily above 40 hPa), which is consistent with the observed relationship between the Niño 3.4 index and the BDC (e.g., Rao et al. 2019 and references therein). By mid-spring the enhanced residual circulation along the shallow branch of the BDC persists (Figs. 1b and 1d), though the vertical extent is reduced, while the ENSO-induced changes along the deep branch of the BDC become negligible.

The accelerated residual circulation along the deep branch of the BDC reflects, in part, an increase in planetary wave

145 driving in the extratropical upper stratosphere (not shown), which is dominated by planetary wavenumber one (Fig. 2a, see also Li and Lau 2013 and Rao et al. 2019 for similar results). In addition to accelerating the residual circulation, the anomalous planetary waves are also associated with enhanced wave breaking (McIntyre and Palmer 1983) as evidenced by anomalously large EKE ~~spanning higher latitudes~~extending from the North Atlantic eastward to the Kamchatka Peninsula (Fig. ~~2b). Because~~2b). The time tendency of ozone is proportional to the negative residual circulation meridional wind times the

150 meridional gradient of ozone, and proportional to the eddy flux divergence (see Sect. 9.4 and Appendix 9A of Andrews et al. 1987 for details); thus, because the latitudinal gradient of O_3S is negative (O_3S decreases poleward, Fig. 3 right columns), the enhanced residual circulation and planetary wave breaking (and associated quasi-isentropic mixing) increases O_3S at high latitudes and decreases O_3S in the extratropics (Fig. 2c). At the beginning of February, this O_3S pattern is seen as the north-south dipole structure between 5-10 hPa in Fig. 3a. As spring proceeds, that O_3S anomaly pattern is advected poleward and

155 downward by the enhanced residual circulation so that the positive anomaly that is initially located between 80°-90° N and 5-10 hPa during February (Figs. 3a-b) is advected downward to 20-30 hPa by April (Figs. 3e-f), and to 40-50 hPa by May and June (Figs. 3g-j). However, these anomalies never reach ~~the lowermost stratosphere~~below 70-100 hPa, and thus appear to be of negligible importance to spring season STT of O_3S. This raises the question, if it is not the deep branch of the BDC (i.e., the combined effects of quasi-isentropic mixing and the residual circulation) that govern the ENSO-induced O_3S anomalies in

160 the lowermost stratosphere~~,~~ (100-300 hPa), then what is responsible for the broad increase in extratropical O_3S anomalies in the lowermost stratosphere from winter to spring during El Nino (Fig. 3, left column)?

One possible explanation is hinted at in NASA Microwave Limb Sounder and GEOS-chemistry climate model simulations shown in Oman et al. 2013 and Olsen et al. 2016, where total column ozone was regressed onto the Niño 3.4 index to reveal a

faint pattern of anomalies extending outwards from the tropics towards North America (see their Figs. 6 and 4, respectively).

165   Similarly, Zhang et al. (2015) use National Institute of Water and Atmospheric Research column ozone (an observational product that assimilates satellite measurements) to show a similar pattern during January to March, though their column ozone patterns are a bit more difficult to interpret. Nevertheless, Zhang et al.'s suggestion that lower stratospheric ozone anomalies arise due to teleconnections forced by ENSO convection is supported by our WACCM simulations. However, the WACCM results considered here suggest that the impact of the teleconnections extends far into the interior of the stratosphere and are

170   thus not due solely to anomalies in tropopause height as suggested by Zhang et al. In particular, the O_3S dipole between 20-200 hPa and poleward of 40° N (Figs. 3a-d) is almost completely explained by vertically deep teleconnections, where geopotential height and O_3S are almost perfectly anticorrelated (cf. Figs. 4a and 4b and Figs. 4c and 4d, both on the 200 hPa surface). Similar geopotential height-O_3S patterns are observed in our WACCM simulations all the way up to 20 hPa (not shown), where the geopotential height and O_3S patterns transition to the wavenumber one structure shown in Fig. 2.

175   The anticorrelated geopotential height-O_3S patterns (Fig. 4) responsible for the high latitude O_3S dipole between 3020-200 hPa (Figs. 3a-d) are consistent with the vertical motion and horizontal advection explanation first described and modeled by Reed (1950) using single column vertical profiles of observed ozone and atmospheric circulation. The underlying dynamics of the synoptic scale wave-ozone relationship suggested by Reed were later confirmed by Schoeberl and Krueger (1983) using Nimbus 7 Total Ozone Mapping Spectrometer ozone data and First Global GARP Experiment (FGGE) temperature

180   (geopotential height) data (Salby 1982; see also Salby 1982and Callaghan 1993). Schoeberl and Krueger conclude with several findings that are relevant to the current WACCM results, namely: (1) for medium-scale waves, geopotential height and ozone anomalies should be anticorrelated; (2) vertical motion and horizontal advection are of equal importance to generating the ozone anomalies; and (3) evanescent waves in the lower stratosphere should produce the maximum ozone signal because they will have minimal phase tilt with height and thus vertical motion and horizontal advection will cause additive anomalies in

185   ozone. The signature of the medium-scale geopotential height waves shown in Figs. 4b and 4d are part of wave structures that exhibit very little phase tilt with height (Fig. 5), which is consistent with waves that are largely evanescent (Charney and Drazin 1961). Thus, the anticorrelated (180° out-of-phase) geopotential height-O_3S patterns shown in Fig. 4 are consistent with the findings of Schoeberl and Kreuger, suggesting that the O_3S anomalies are caused by vertical motion and horizontal advection associated with medium-scale evanescent waves (i.e., planetary wavenumber >2). This is in contrast to what would

190   be expected from longer scale vertically propagating planetary waves (planetary wavenumber <3), which are typically associated with significant poleward and downward eddy-ozone flux transport in the upper stratospheric photochemical transition region where ozone and geopotential height are 180° out-of-phase, but cause minimal transport in the lower stratosphere where ozone and geopotential height tend to be close to in-phase (e.g., Fig. 8 of Hartmann and Garcia 1979; see also Garcia and Hartmann 1980, Gille et al. 1980, Hartmann 1981, Albers and Nathan 2012). While the stratospheric ozone

195   anomalies due to medium-scale waves may be largely reversible in isolation (Salby and Callaghan 1993, Fusco and Salby 1999), nonconservative process typically cause irreversible transport of ozone to the troposphere. Thus, the primary role of the ozone teleconnections may be to simply temporarily modulate the amount of ozone in the lowermost stratosphere that is

available for subsequent transport into the middle to lower troposphere via tropopause folds, potential vorticity streamers and cutoffs (Reed and Danielson, 1958; Hoerling et al., 1993; Langford and Reid, 1998; Shapiro, 1980; Sprenger et al., 2007; Škerlak et al., 2015), and transverse circulations in jet exit regions (Langford et al., 1998; Langford, 1999).

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The geopotential height teleconnections, and associated O_3S anomalies, peak in February and March (cf. Figs. 3a-d and Fig. 4 for February-March) and quickly decay thereafter (cf. Figs. 3e-h and Fig. 6 for April-May). As the teleconnections dissipate, at least a portion of the large positive O_3S anomaly that was once located over the North Pacific (Fig. 4a) is mixed northward, leading to weak positive O_3S anomalies over most of the Northern Hemisphere poleward of 50° N (Figs. 3e-j and

205 6a). This poleward mixing reflects the seasonal cycle of the stratospheric polar vortex and the build-up and breakdown of the polar transport barrier, which proceeds as follows (e.g., Manney et al. 1994). During midwinter, the polar night jet is typically located in the mid- to upper stratosphere, which, because of the associated strong vortex edge potential vorticity gradient, creates a barrier to latitudinal transport. As spring onsets, the upper portions of the vortex weaken and the polar night jet descends and establishes a transport barrier in the lowermost stratosphere. However, once the stratospheric final warming

210 occurs – typically sometime in March or April (Butler and Domeisen 2021) – the transport barrier is erased and mixing between polar and midlatitude air rapidly ensues (Manney et al. 1994, Salby and Callaghan 2007a,b), hence the anomalies shown in Figs. 3e-j and 6a. In the next section, we outline how the ENSO-induced O_3S anomalies just described constructively reinforce ENSO-induced changes in synoptic wave activity (and hence stratospheric intrusions) that modulate STT.

215 **3.2 Upper tropospheric-lower stratospheric circulation and STT**

Lin et al. (2015) suggested that deep STT of ozone over western North America should increase when La Niña (measured by the Niño 3.4 index) perturbs the polar front jet northward and invigorates it so that the frequency of deep tropopause folds increases. Consistent with this hypothesis, Breeden et al. (2021) find that in reanalysis, EKE and deep mass transport (though

220 not necessarily ozone transport) over the western US increase during time periods when the Niño 3.4 index is at least moderately negative ($\leq$-0.5° C). However, Breeden et al. further show that most of the ENSO jet-related changes occur in late winter and early spring before the Pacific jet structure transitions from its winter to summer-like state (i.e., the jet transitions from being strong and zonally contiguous to being weak, with a discontinuity over the Pacific basin, e.g., Breeden et al. Fig 2). As a result, tropopause fold depth and frequency are increased primarily in February to mid-April during La Niña,

225 and there are only smaller jet related changes in transport thereafter.

In agreement with the observed relationship between ENSO and EKE (e.g., Breeden et al. 2021, see their Supplemental Fig. S6), the WACCM El Niño simulations show decreased EKE over the North Pacific-western US and increased EKE over Baja California-southern US during February and March (Fig. 7a), and vice versa for La Niña. The lower EKE to the north and higher to the south reflects the tendency for the time mean Pacific jet to shift southwards during El Niño and northward

230 during La Niña (Shapiro et al 2001). Here, as in Shapiro et al. (2001), the time mean tends to convolve aspects of the subtropical and polar front jets (see also, Koch et al. 2006); however, what is important here, is that synoptic timescale anomalies in EKE

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are ~~well-correlated~~associated with shifts in the location of either jet, which in turn are associated with anomalous STT via midlatitude tropopause folds and mixing (Shapiro ~~1980,~~et al. 2001, Breeden et al. 2021) or transverse circulations in the subtropical jet exit region (Langford 1999). Thus, comparing regions with anomalously high or low EKE (Fig. 7a) and high or

235 low O_3S in the lowermost stratosphere (Figs. 4a and 4c) allows us to assess whether changes in tropopause fold frequency and/or transverse circulations near the nose of the jet, and stratospheric changes in the amount ozone available for downward transport, respectively, act constructively or in opposition to generate the observed anomalies of O_3S in the lowermost troposphere (Fig. 8 shows the case for El Niño; see Supplement Fig. S3 for the complementary La Niña composites).

During February and March, there is anomalously high STT of O_3S to the lower troposphere over the North Pacific,

240 maximizing between roughly 30°-45° N (Figs. 8a and 8c). In the extratropics, isentropic surfaces slope downwards from pole-to-equator (e.g., Fig. 2 of Gettelman et al. 2011), thus lower tropospheric O_3S anomalies associated with stratospheric intrusions should be located somewhat equatorward of the corresponding lower stratospheric O_3S and EKE anomalies (i.e., the EKE and ozone anomalies in the North Pacific in Figs. 4a and 7a, respectively). Because EKE is reduced over ~~this region~~the North Pacific (Fig. 7a), which in isolation should correspond to a reduction in tropopause fold frequency~~,~~ (Breeden et al. 2021),

245 the anomalously high stratospheric O_3S availability in the North Pacific (Fig. 4a) ~~must be~~is likely the controlling factor governing the enhanced O_3S transport ~~into this region.~~to the lower troposphere over the central Pacific. That said, transverse circulations (Langford 1999) associated with the El Niño-induced southward shift in the subtropical jet (note the region of enhanced EKE running east-west just north of Hawaii, Fig. 7a) may also contribute to the lower tropospheric O_3S anomalies in the central Pacific. On the other hand, any isentropic mixing associated with the enhanced EKE north of Hawaii is unlikely

250 to be contributing to the lower tropospheric O_3S anomalies because at these lower latitudes (near 30° N), isentropic surfaces in the upper troposphere-lower stratosphere are relatively flat (in the height plane), so that any isentropic mixing would result in STT of ozone into the tropical upper troposphere (e.g., Waugh and Polvani 2000, Albers et al. 2016). In contrast to the situation over the Pacific Ocean, over the western US, reduced EKE and anomalously low O_3S operate constructively to reduce STT of O_3S. Similarly, though opposite in sign, the El Niño-induced increases in EKE and lower stratospheric O_3S

255 constructively contribute to enhanced deep STT of O_3S over Baja California and the southeastern US. By April-May, the lower stratospheric O_3S anomalies and EKE and over the western US have weakened considerably (Figs. 6a and 7b, respectively), which is reflected in anomalous O_3S only reaching the mid-troposphere (Figs. 8e-h). However, enhanced EKE and anomalously high lower stratospheric O_3S continue to contribute to robust deep STT of O_3S over Baja California and the southeastern US. By June, the peak in deep STT of O_3S has receded westward and shifted northward so that the maximum

260 transport is located over Baja California and the southwestern US (Figs. 8i,j). These results make clear that the sign of anomalous STT of O_3S due to ENSO is highly dependent on the region and time of year. For example, for the western US between 30°- 45° N, February-March El Niño conditions suppress the transport of stratospheric O_3S, but as spring progresses into summer (April-June), El Niño instead enhances O_3S transport.

4 Conclusions

265 STT of ozone is modified by both stratospheric ozone variability and tropopause level jet dynamics (e.g., Lin et al. 2015, Albers et al. 2018, Langford et al. 2022 and references therein). Understanding the relative importance of these two processes is critical for both subseasonal prediction (Lin et al. 2015, Albers et al. 2018, 2021), as well as for assessing the interannual variability of tropospheric ozone concentrations relevant to the NAB (Fiore et al. 2003, 2014).

Our results suggest that there is no conflict between the results of Lin et al. (2015) versus Langford (1999) who hypothesize
270 that La Niña versus El Niño, respectively, enhance western US STT of ozone during spring. Indeed, the two hypotheses can be reconciled by carefully accounting for the way that tropical-to-extratropical teleconnections modulate both lower stratospheric ozone availability and jet-related transport variability, which are both strongly dependent on the geographic region and month of the year under consideration. For example, during February and March, our results confirm the hypothesis of Lin et al. (2015) suggesting that when La Niña conditions are present (as measured by the Niño 3.4 index), there will be
275 enhanced STT over the western US (Supplemental Fig. S3). However, in contrast to Lin et al., our results here suggest that both tropospheric jet dynamics and stratospheric ozone availability contribute to the anomalous transport. At the same time, our results also confirm the hypothesis of Langford (1999), as we find that El Niño conditions lead to enhanced STT of ozone over Baja California and the southern US during February and March (Figs. 7a8a-d), and enhanced ozone transport over the western US during May and June (Figs. 7g8g-j). Again, our results suggest that in the later spring period, both tropospheric jet
280 processes and stratospheric ozone availability constructively reinforce each other.

~~In addition, the relatively large ensemble of WACCM time slice simulations cleanly demonstrate how tropical-extratropical teleconnections cause zonally asymmetric anomalies in lower stratospheric ozone, which subsequently strongly modulate STT of ozone. In particular, as suggested by Reed (1950) and Schoeberl and Krueger (1983), medium-scale evanescent waves, which are largely barotropic, drive vertical motions and horizontal advection that lead to ozone anomaly~~
285 ~~patterns that are nearly perfectly anticorrelated with geopotential height teleconnection patterns~~ (The relatively large ensemble of WACCM time slice simulations cleanly demonstrates which aspects of stratospheric transport are most important to ENSO-driven STT of ozone. Previous studies have highlighted that ENSO modulates the deep branch of the BDC on seasonal timescales via changes in the residual circulation and quasi-horizontal mixing (e.g., Calvo et al. 2010, Neu et al. 2014, Diallo et al. 2019, Benito-Barca et al. 2022), a result that is supported here (Figs. 1-3). However, owing to the slowness of the deep
290 branch of the residual circulation, ENSO-driven ozone anomalies originating in the upper stratosphere do not reach the lowermost stratosphere by spring (Fig. 3) when STT of ozone maximizes over North America (James et al. 2003, Lefohn et al. 2001, Škerlak et al. 2014, Terao et al. 2008, Albers et al. 2018). In contrast to the vertically propagating planetary waves (zonal wave numbers <3) that dominate the seasonal changes in the deep branch of the BDC in the upper stratosphere, significant extratropical lower stratospheric ozone anomalies evolving on *sub-monthly* timescales can arise due to medium-
295 scale planetary waves (Reed 1950, Schoeberl and Krueger 1983, Salby and Callaghan 1993). Owing to their higher vertical wavenumber (zonal wavenumbers >3), these waves are largely evanescent (Charney and Drazin 1961), yielding barotropic

geopotential height and ozone anomalies that resemble the Pacific-North American teleconnection pattern (Fig. 4). As noted by previous studies (e.g., Salby and Callaghan 1993, Fusco and Salby 1999), the ozone transport associated with such medium-scale waves often involves only a conservative redistribution of ozone that is largely reversible. Indeed, our results suggest that, at least to some extent, the large ozone anomalies associated with the ENSO-teleconnections are transient and reversible; for example, the large negative ozone anomalies seen at high latitudes over Northern Canada in February and March (Figs. 3a and 3c and Fig. 4a) largely disappear by April-March (Figs. 3e and 3g and Fig. 6a), though as mentioned in Sect. 3.1, some amount of hemispheric-scale mixing may occur during the spring breakdown of the polar transport barrier. Nevertheless, even if the ENSO-induced lower stratospheric ozone transport is, in isolation, largely reversible, processes including tropopause folds, potential vorticity streamers and cutoffs (Reed and Danielson, 1958; Hoerling et al., 1993; Langford and Reid, 1998; Shapiro, 1980; Sprenger et al., 2007; Škerlak et al., 2015) and transverse circulations in jet exit regions (Langford et al., 1998; Langford, 1999) can lead to significant irreversible STT of ozone extending downwards to the middle to lower troposphere (Fig. 8). Thus, our findings suggest that ENSO primarily modulates extratropical lower stratospheric ozone abundances, and hence STT of ozone, via transient ozone teleconnections rather than via changes in the residual circulation and mixing, in agreement with the findings of Zhang et al. (2015). While the residual circulation and isentropic mixing tend to be the focus of most BDC studies, if the BDC is more broadly defined to include any process that contributes to mass transport from the tropical tropopause to its eventual exit back into the troposphere at high latitudes (see Butchart 2014 for a historical discussion of the definition of the BDC), then reversible transport due to medium-scale waves coupled with irreversible STT represents an important process whereby subseasonal variability may contribute to BDC variability.

Fig. 4). As a result, our findings here confirm that ENSO driven ozone teleconnections are the dominate mechanism controlling lower stratospheric ozone changes relative to ENSO driven changes in the Brewer-Dobson circulation, in agreement with the findings of Zhang et al. (2013).

The sensitivity of ozone transport to the trajectory of tropical-extratropical teleconnections highlights a problematic aspect of attempting to use ENSO indices (e.g., Niño 3.4) to make subseasonal-to-seasonal predictions of STT of ozone. For example, a small northward shift in the location of the wave train depicted in Fig. 4 may change the sign of the western US ozone transport anomaly altogether (i.e., the positive ozone anomaly over Baja California could displace the negative ozone anomaly of the western US in Figs. 7a8a-d). Indeed, even during times when the Niño 3.4 index is strongly loaded, ENSO diversity (Capotondi et al. 2015) can yield distinctly different teleconnection patterns (e.g., Garfinkel et al. 2013). Moreover, when ENSO teleconnections are convolved with internal variability, the resulting anomaly patterns over North America can vary significantly (e.g., Deser et al. 2017, 2018), which further complicates using ENSO to predict STT of ozone. Thus, instead of using a Niño-based index as a predictor, a perhaps more reliably method is to use a measure of signal or signal-to-noise ratio to identify time periods when the teleconnections themselves, and therefore STT, may be more predictable (e.g., Albers and Newman 2019, Albers et al. 2021).

Code and Data Availability

The code used to perform this analysis can be accessed by personal communication with the corresponding author. The WACCM simulation data used to create figures can be accessed here: https://csl.noaa.gov/groups/csl8/modeldata/data/Albers_etal_2022/ .

Author Contributions

JRA wrote climate model analysis code, created the figures, and wrote the manuscript. AHB conducted the climate model experiments. AHB, DE, MLB, and AOL provided comments and edited the manuscript.

Competing Interests

The authors declare no conflict of interest.

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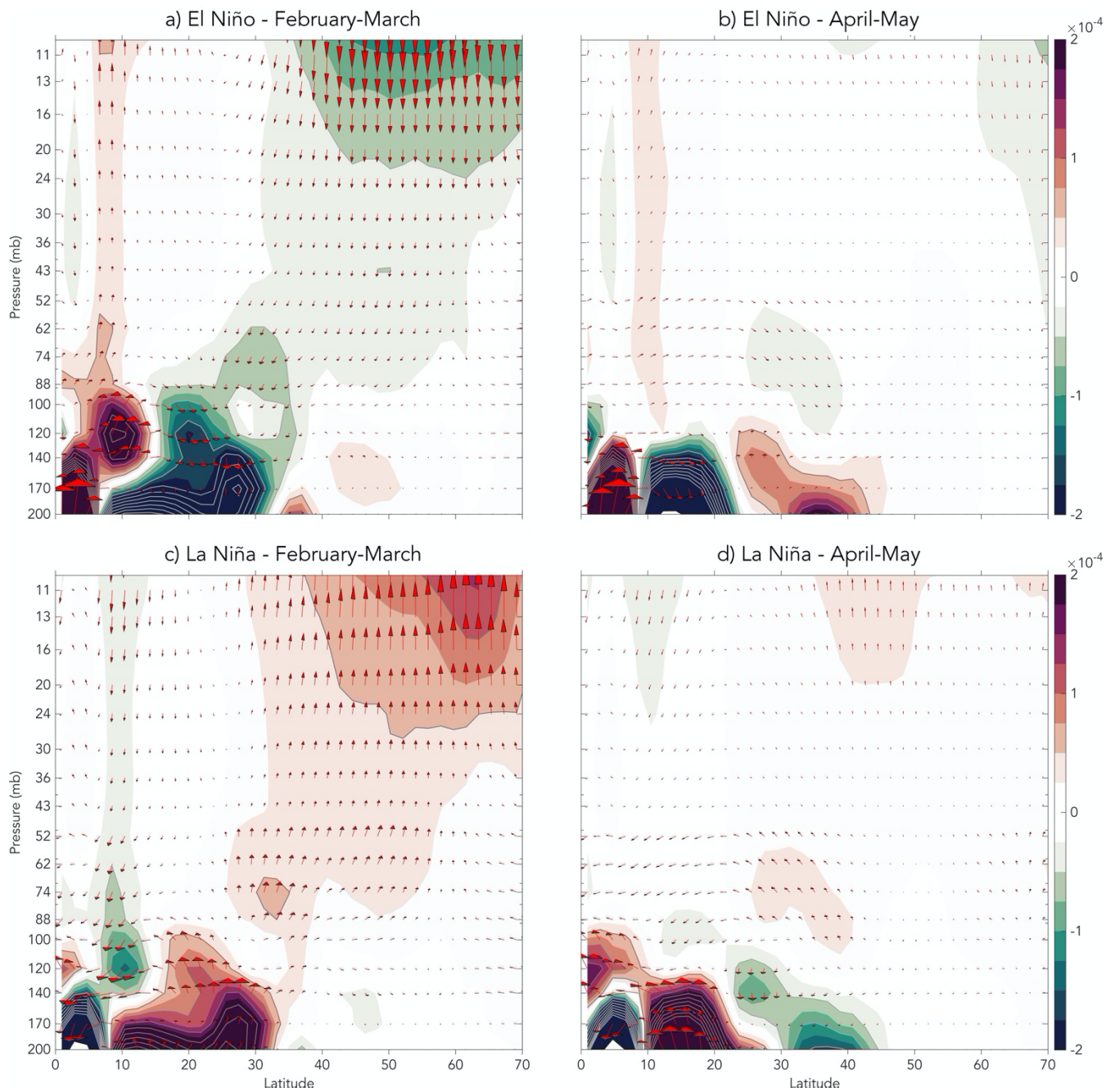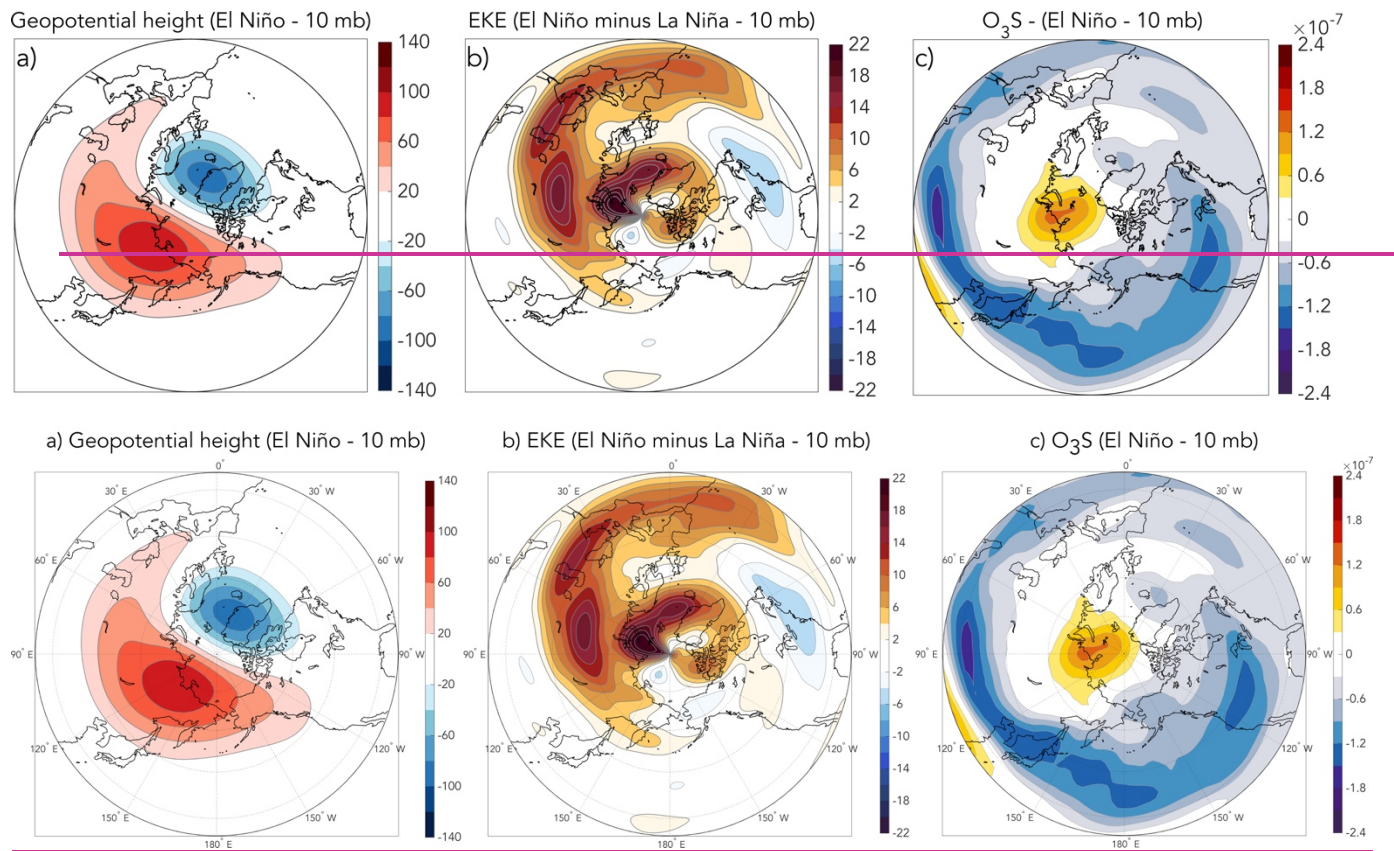
Figure 1: Transformed Eulerian mean (TEM) residual circulation anomalies (red arrows) and TEM residual vertical velocity anomalies ($\bar{w}^*$, filled color contours) for: (a) February-March and (b) April-May of the El Niño time slice simulations; and (c) February-March and (d) April-May of the La Niña time slice simulations. Units for the residual circulation are in mm s^{-1}.

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Figure 2: February-March 10 hPa anomaly composites of (a) geopotential height (units: gpm) for El Niño, (b) eddy kinetic energy (2 to 120-day filtered with units: m^2 s^{-2}) for El Niño minus La Niña, and stratospheric O$_3$S (units: mass mixing ratio) for El Niño.
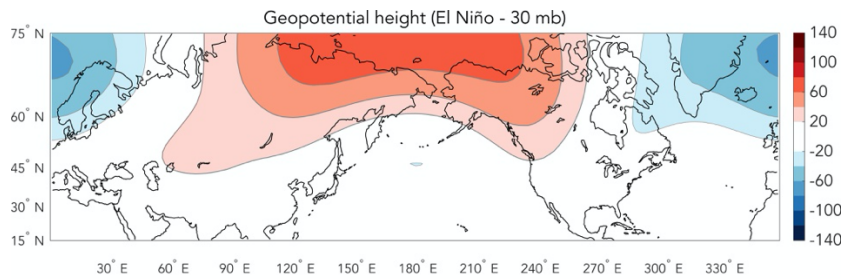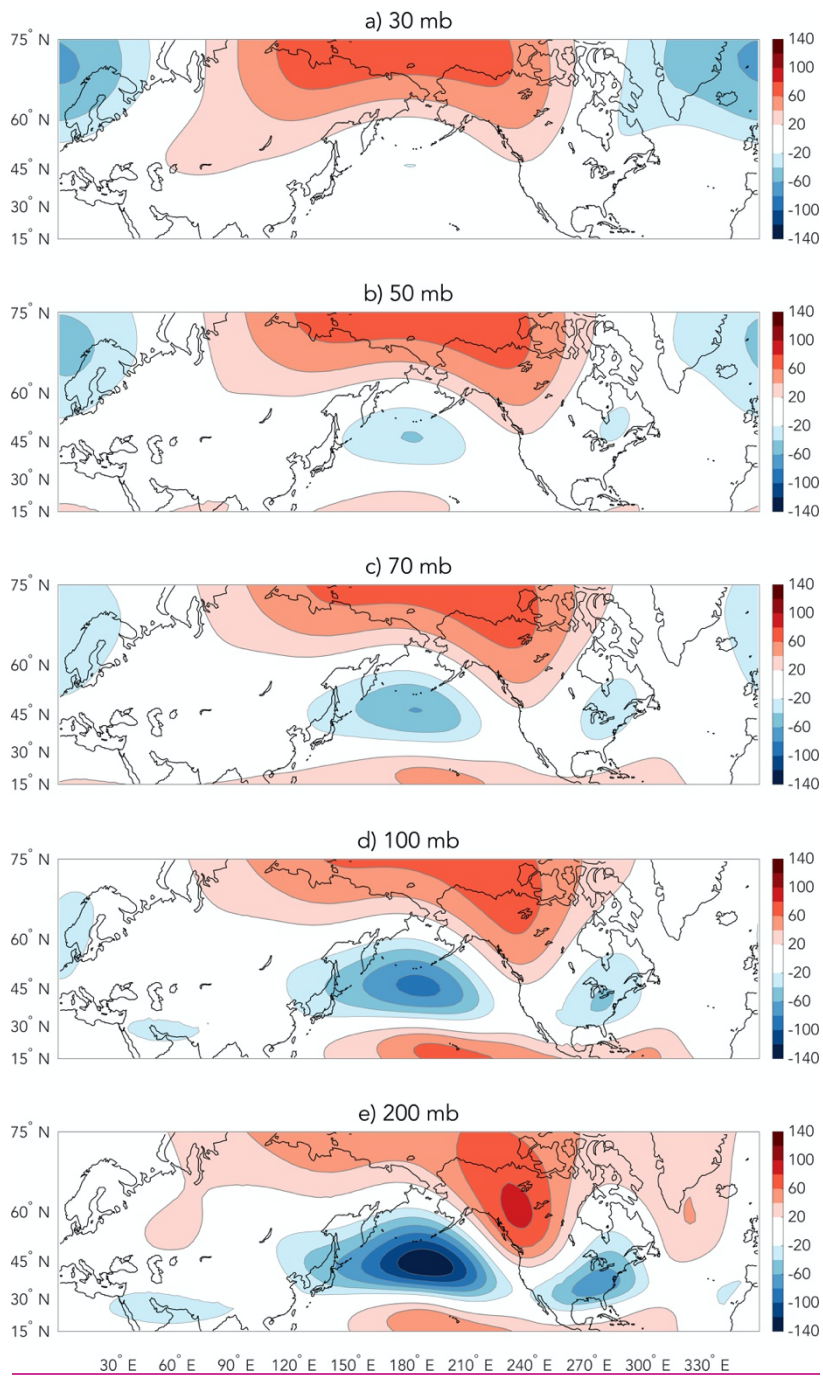
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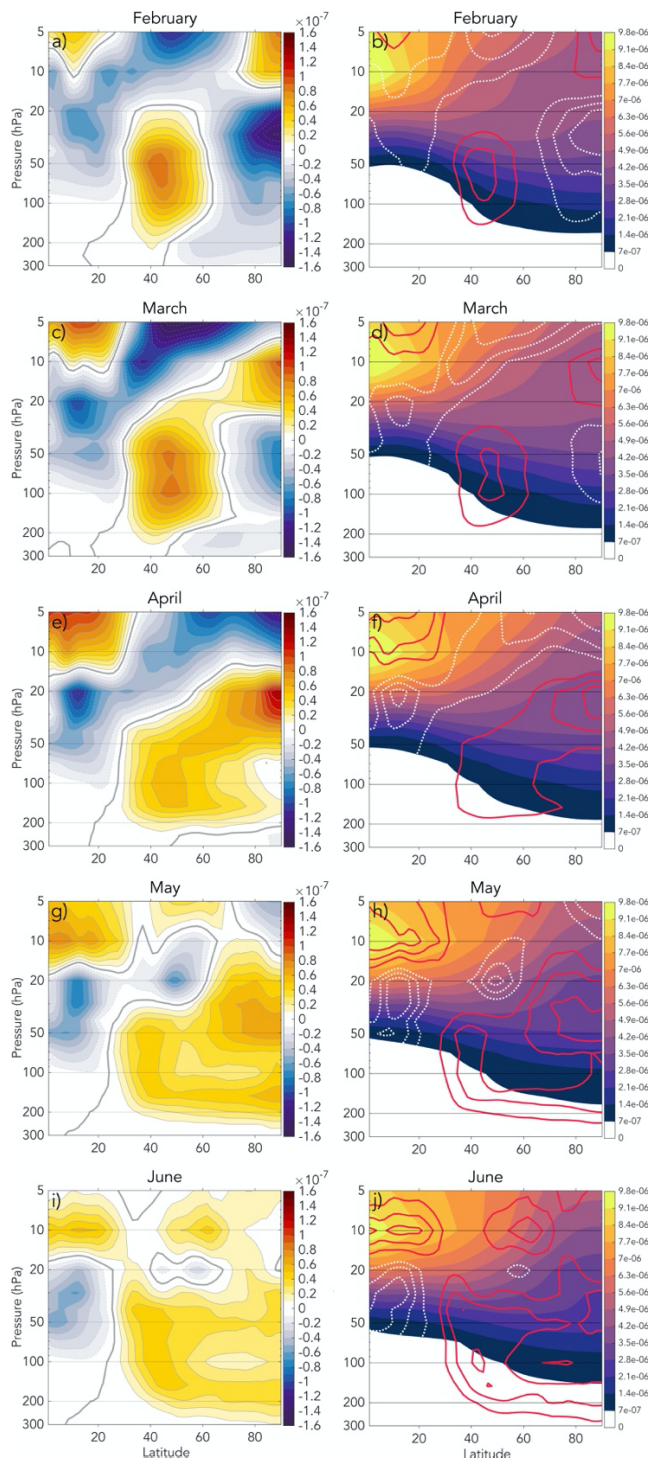
Figure 3: February-June El Niño composites of: (left column) O₃S anomalies, and (right column) climatological O₃S (filled contours) with positive (red contours) and negative (white dashed contours) O₃S anomalies from the left columns overlayed. All units are mass mixing ratio.
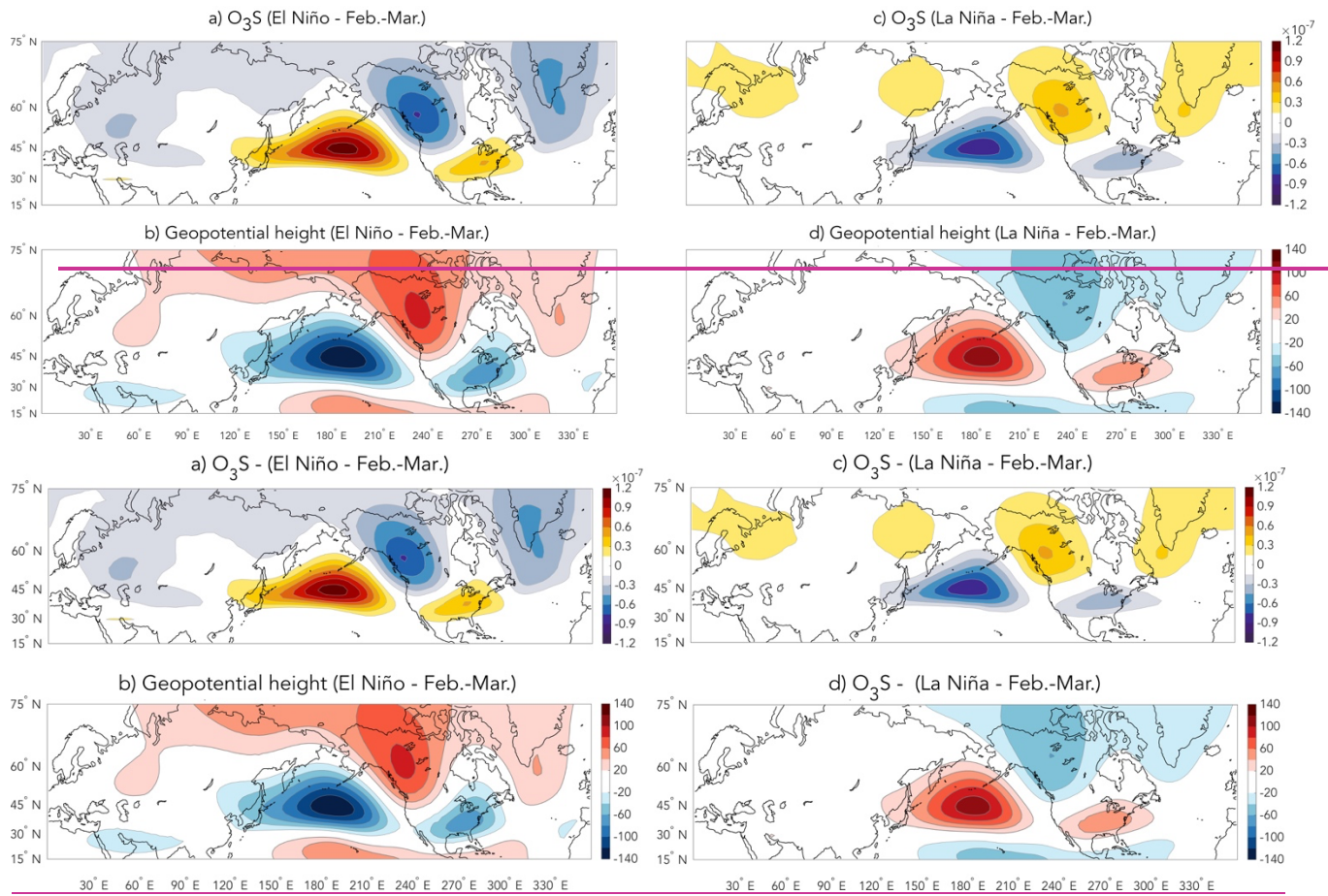
Figure 4: February-March (200 hPa pressure level) anomaly composites of: O_3S (units: mass mixing ratio) for (a) El Niño and (b) La Niña; and geopotential height (units: gpm) for (c) El Niño and (d) La Niña.
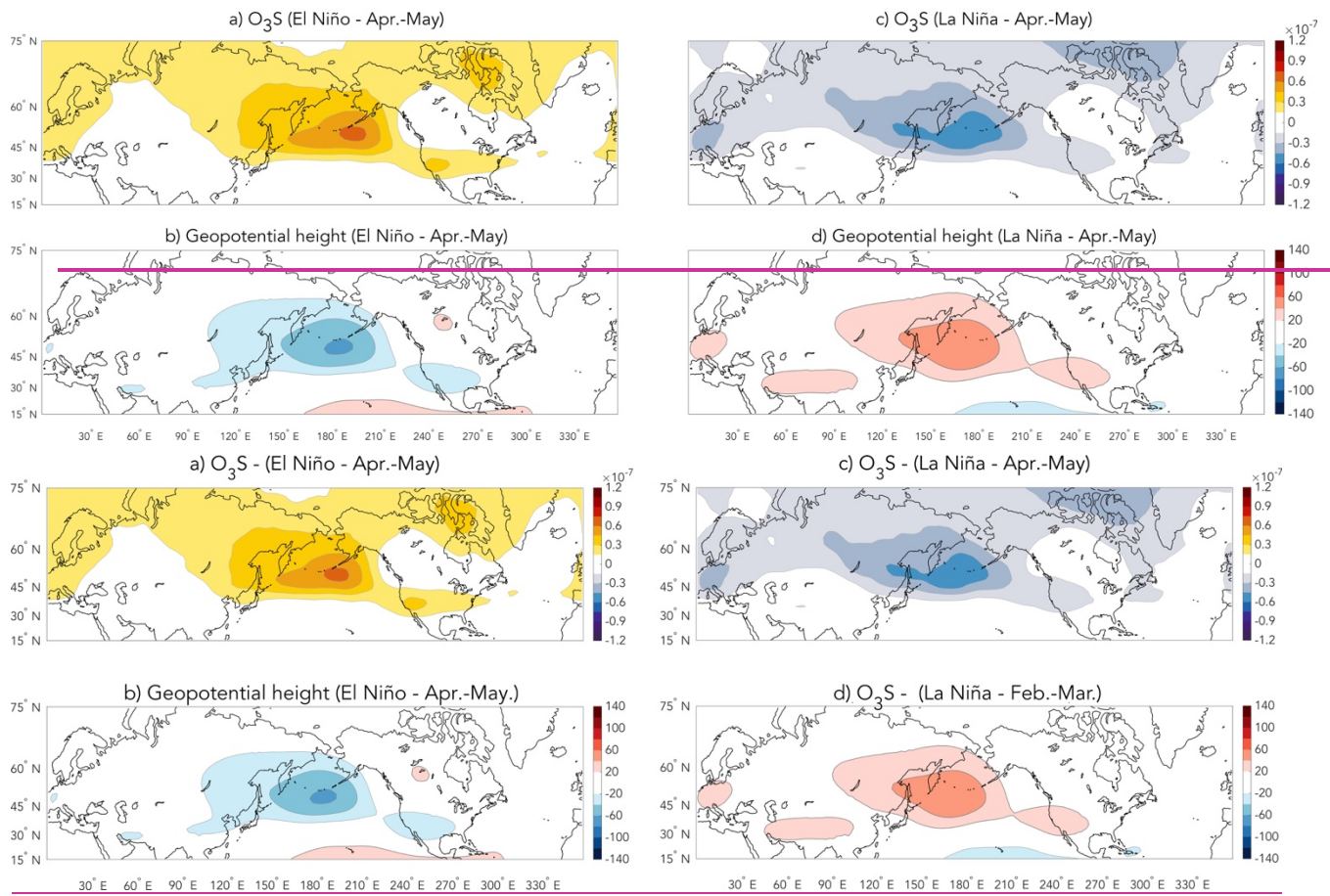
21

Geopotential height (El Niño - 30 mb)

Geopotential height (El Niño - 50 mb)

Geopotential height (El Niño - 70 mb)

Geopotential height (El Niño - 100 mb)

Geopotential height (El Niño - 200 mb)
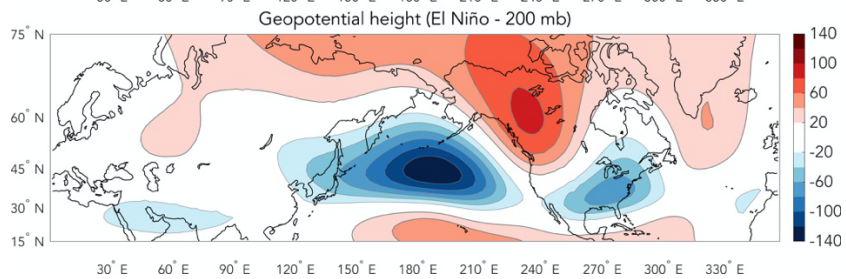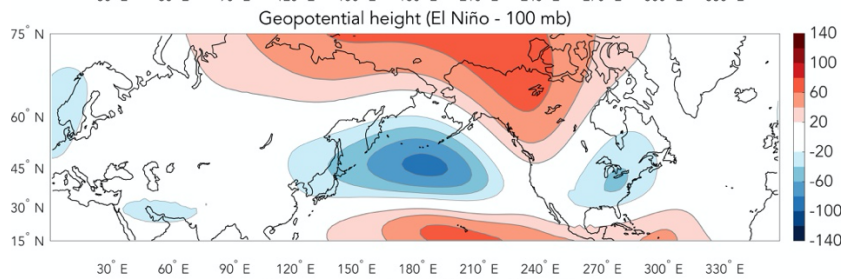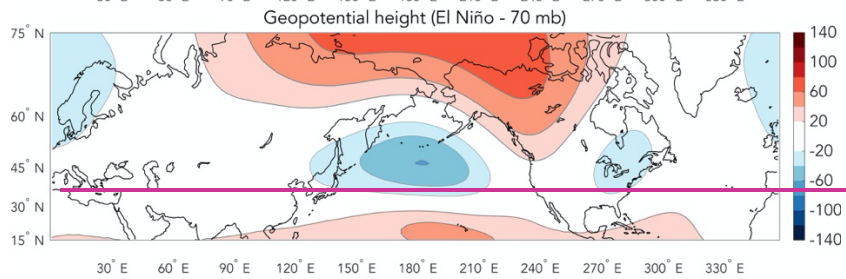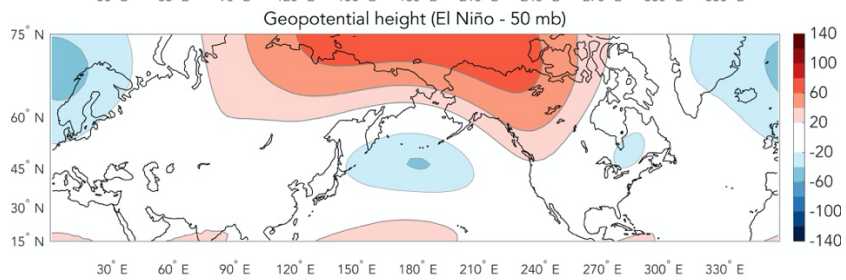
Figure 5: February-March El Niño geopotential height (units: gpm) anomaly composites between 30-200 mb (top to bottom).

625

23
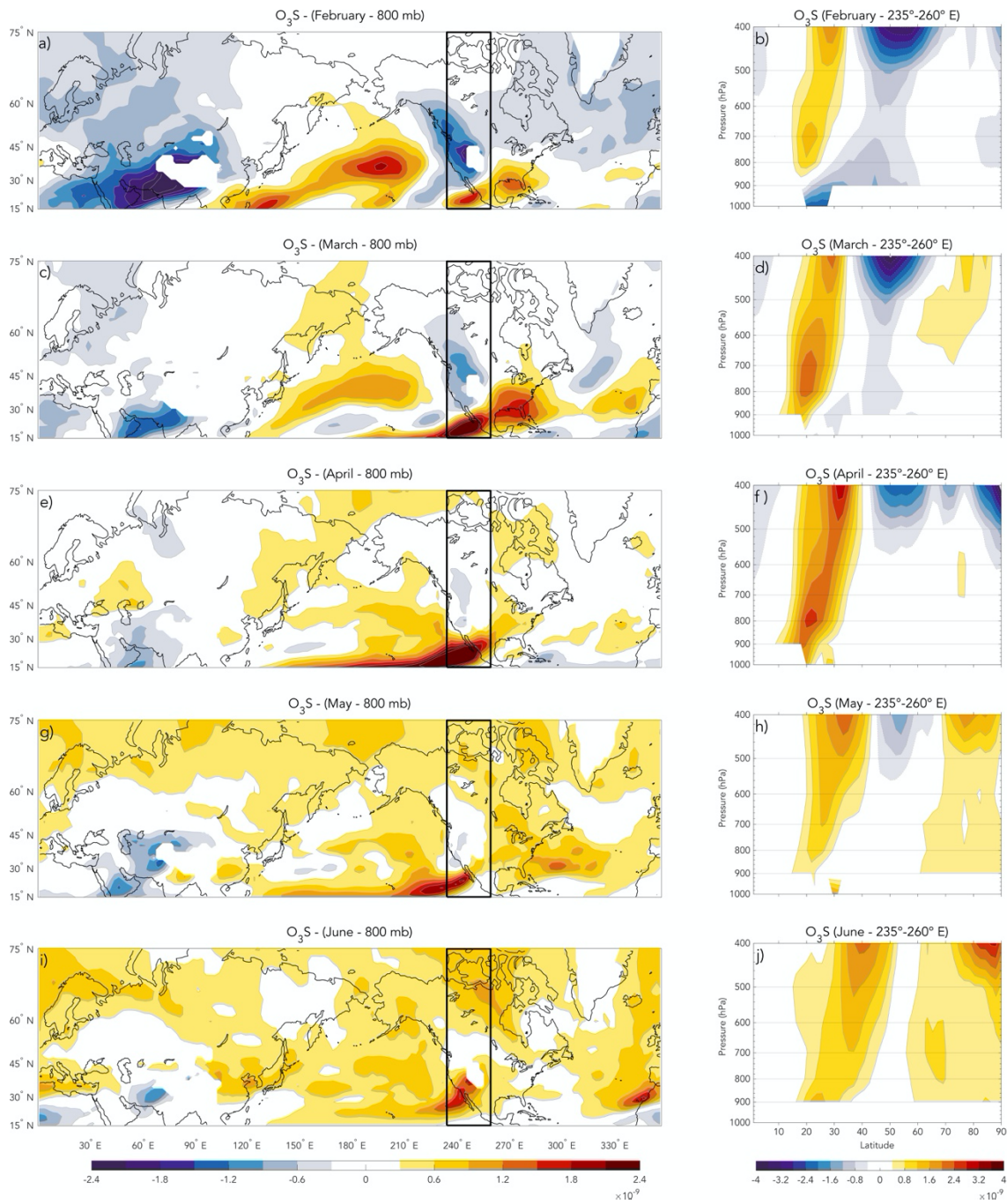
Figure 6: April-May (200 hPa pressure level) anomaly composites of: O$_3$S (units: mass mixing ratio) for (a) El Niño and (b) La Niña; and geopotential height (units: gpm) for (c) El Niño and (d) La Niña.
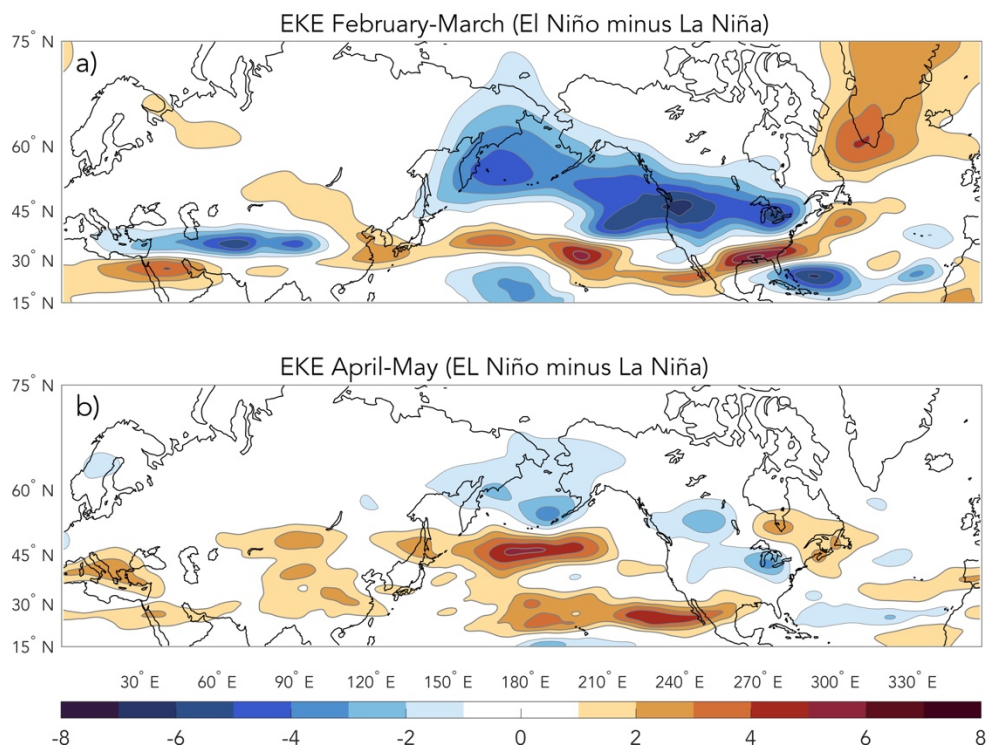
Figure 7: 200 hPa pressure level eddy kinetic energy (2 to 10-day filtered) anomaly composites for: (a) February-March (El Niño minus La Niña), and (b) April-May (El Niño minus La Niña). Units are: m^2 s^{-2}.

Figure 8: February-June El Niño composites of: (left column) 800 hPa O₃S anomalies, and (right column) zonal mean anomaly cross-sections of O₃S where the longitudinal boundaries (235°-260° E) of the zonal mean correspond to the box overlaying the left column composites. All units are mass mixing ratio. White areas reflect missing data associated with topography.