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The Teleconnection of El Niño Southern Oscillation to the Stratosphere

Daniela I.V. Domeisen[1], Chaim I. Garfinkel[2], and Amy H. Butler[3,4]

[1]Institute for Atmospheric and Climate Science, ETH Zürich, Zürich, Switzerland, [2]Fredy and Nadine Herrmann Institute of Earth ScienceHebrew University, Jerusalem, Israel, [3]Cooperative Institute for Research in Environmental Sciences, University of Colorado Boulder, Boulder, CO, USA, [4]NOAA Chemical Sciences Division, Boulder, CO, USA

Key Points:

- ENSO has a detectable impact on the composition and circulation of the stratosphere in the tropics and extratropics
- The changes in stratospheric variability due to ENSO have implications for improving surface prediction
- Recent advances in modeling have helped to put the response to the small sample of observed ENSO events in context

Correspondence to:
D. I. V. Domeisen,
daniela.domeisen@env.ethz.ch

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Abstract El Niño and La Niña events in the tropical Pacific have significant and disrupting impacts on the global atmospheric and oceanic circulation. El Niño Southern Oscillation (ENSO) impacts also extend above the troposphere, affecting the strength and variability of the stratospheric polar vortex in the high latitudes of both hemispheres, as well as the composition and circulation of the tropical stratosphere. El Niño events are associated with a warming and weakening of the polar vortex in the polar stratosphere of both hemispheres, while a cooling can be observed in the tropical lower stratosphere. These impacts are linked by a strengthened Brewer-Dobson circulation. Anomalous upward wave propagation is observed in the extratropics of both hemispheres. For La Niña, these anomalies are often opposite. The stratosphere in turn affects surface weather and climate over large areas of the globe. Since these surface impacts are long-lived, the changes in the stratosphere can lead to improved surface predictions on time scales of weeks to months. Over the past decade, our understanding of the mechanisms through which ENSO can drive impacts remote from the tropical Pacific has improved. This study reviews the possible mechanisms connecting ENSO to the stratosphere in the tropics and the extratropics of both hemispheres while also considering open questions, including nonlinearities in the teleconnections, the role of ENSO diversity, and the impacts of climate change and variability.

Plain Language Summary El Niño and La Niña events, the irregular warming and cooling of the tropical Pacific that occurs every couple of years, have disrupting impacts spanning the entire world. These remote impacts, so-called "teleconnections", also reach the stratosphere, the layer of the atmosphere starting at around 10 km above the Earth's surface. El Niño leads to a warming of the stratosphere in both hemispheres, while the lower tropical stratosphere cools. These signatures are linked by a strengthened stratospheric circulation from the tropics to the polar regions. El Niño also leads to more frequent breakdowns of the stratospheric polar vortex, a band of strong eastward winds in the polar stratosphere. For La Niña, these effects tend to be opposite, though they are not always robust, suggesting nonlinear or nonstationary effects, long-term variability, and trends in the teleconnections. The observational data record is not yet long enough to make conclusions with certainty, and models that try to reproduce the teleconnections indicate that teleconnections might be more linear than the limited number of observations indicate. Further research will be needed to separate the El Niño and La Niña teleconnections from other effects and to determine to what extent nonlinearity and nonstationarity are indeed present.

1. Introduction

El Niño Southern Oscillation (ENSO), that is, the warm (El Niño) and cold (La Niña) phases of the equatorial Pacific atmosphere-ocean coupled phenomenon (reviewed in section 2.1), has far-reaching and disrupting impacts on the global atmospheric and oceanic circulation (Halpert & Ropelewski, 1992; Horel & Wallace, 1981; Ropelewski & Halpert, 1987; Trenberth et al., 1998). Various pathways have been identified that connect ENSO to regions around the globe through so-called *teleconnections* (Bjerknes, 1969; Z. Liu & Alexander, 2007). During El Niño years, the teleconnections include drought in Australia and the maritime continent (e.g., Chiew et al., 1998), anomalously low rainfall in the Sahel region (e.g., Anyamba & Eastman, 1996) and East Africa (e.g., Gleixner et al., 2016), a weaker and/or later onset of the Indian monsoon (e.g., Dai & Wigley, 2012), and reduced tropical cyclone activity over the Atlantic Ocean (e.g., Klotzbach, 2011).

Remote effects also reach regions outside of the tropics (reviewed in sections 2.2.2 and 2.2.3): Anomalies in tropical heating and convection associated with ENSO can drive planetary-scale waves that propagate poleward (e.g., Hoskins & Karoly, 1981). During El Niño, these *Rossby wave trains* can have impacts ranging from a deepening of the Aleutian low pressure system over the North Pacific (Mo & Livezey, 1986), regional climate impacts over North America including above-average Californian rainfall (e.g., Cayan et al., 1999; Piechota et al., 1997), and a shift toward the negative phase of the North Atlantic Oscillation (NAO; Brönnimann, 2007; Fraedrich & Müller, 1992), which is embedded into the Northern Annular Mode (NAM; Baldwin & Dunkerton, 1999, 2001), and a shift in the Southern Annular Mode (SAM; L'Heureux & Thompson, 2006). The NAM and the SAM constitute the dominant patterns of atmospheric pressure characterizing the variability of the storm tracks in both hemispheres (Baldwin, 2001; Thompson & Wallace, 2000). During La Niña years, these effects are frequently—but not always—opposite.

Over the past decades, it has been discovered through both observations and model simulations that even areas as remote from the equatorial Pacific as the stratosphere (reviewed in section 3) are influenced by ENSO (e.g., Calvo et al., 2008; Hamilton, 1993a, 1993b; Manzini, 2009; Reid et al., 1989; Van Loon & Labitzke, 1987; Yulaeva & Wallace, 1994). El Niño modulates temperatures and winds in the tropical stratosphere (Figure 1 and section 4.1) and hence the radiative budget and chemical composition of the atmosphere. El Niño events also lead to a warming and weakening of the winter stratospheric polar vortex in both hemispheres (Figure 1 and sections 4.2 and 4.3). A weakening of the wintertime polar vortex associated with a complete reversal of the climatological westerly winds is termed a major Sudden Stratospheric Warming (SSW). These events have been shown to influence Northern Hemisphere (NH) wintertime surface weather and climate for days to weeks following the event (section 3.4) and may be an important component for forecasting NH wintertime surface weather and climate on subseasonal to seasonal time scales. Both models and observations tend to agree that at least El Niño—the relationship is less conclusive for La Niña—is associated with a small but significant increase in SSWs (section 4.2), but the relationship remains ambiguous due to the sensitivity of both ENSO and SSW to their respective classification, large sampling variability in the observations, and possible model biases affecting the relationship in models (Polvani et al., 2017). Because ENSO and SSW events each have important influences on wintertime weather, understanding their interaction (or lack thereof) is important for improving prediction.

There has been considerable progress on the understanding of ENSO teleconnections to the stratosphere over the past 10 years since the review by Brönnimann (2007) on the stratospheric pathway of the ENSO teleconnection to Europe. The observational record has since been extended by a number of ENSO events, including the extreme 2015/2016 El Niño event, as well as several SSW events (see Table 1). However, since ENSO events only occur every couple of years, the observational record that can be investigated for the ENSO-stratosphere teleconnection remains short, especially for the satellite era, when more reliable observations of the stratosphere became available. Over the past 10 years, progress has also been made on the understanding of the influence of the stratosphere on the troposphere and the potential for predictability that arises from this connection. The stratosphere has been recognized as an important modulator for remote impacts from a range of sources including ENSO and as a source of subseasonal to seasonal predictability (reviewed in section 3.4), and a greater range of models now allocates more resources to representing stratospheric variability. While the simulation of teleconnections in models is crucial, especially due to the short observational record, validating the representation of teleconnections in models remains challenging: Models often show teleconnections that are more linear that those observed, and it is not yet clear if this discrepancy between models and observations is related to the short observational record, a bias in the models, or both.

This review will cover to what extent the ENSO impacts onto the stratosphere are understood at present and how they have proven important for our understanding of stratospheric variability. It thereby provides an update on our knowledge for a broad diversity of ENSO events and for both hemispheres, including the influence of other phenomena that may lead to nonstationary or nonlinear behavior in the teleconnections (section 5), with an outlook addressing open research questions (section 6).

2. An Overview of ENSO and its Tropospheric Teleconnections

2.1. A Brief Description of ENSO and its Diversity

ENSO is associated with irregular variations in the strength of the easterly *trade winds* and the oceanic thermocline in the tropical Pacific, which separates the warm surface waters from the colder and nutrient-rich deep waters. ENSO variability is associated with different sea surface temperature (SST) patterns and the related

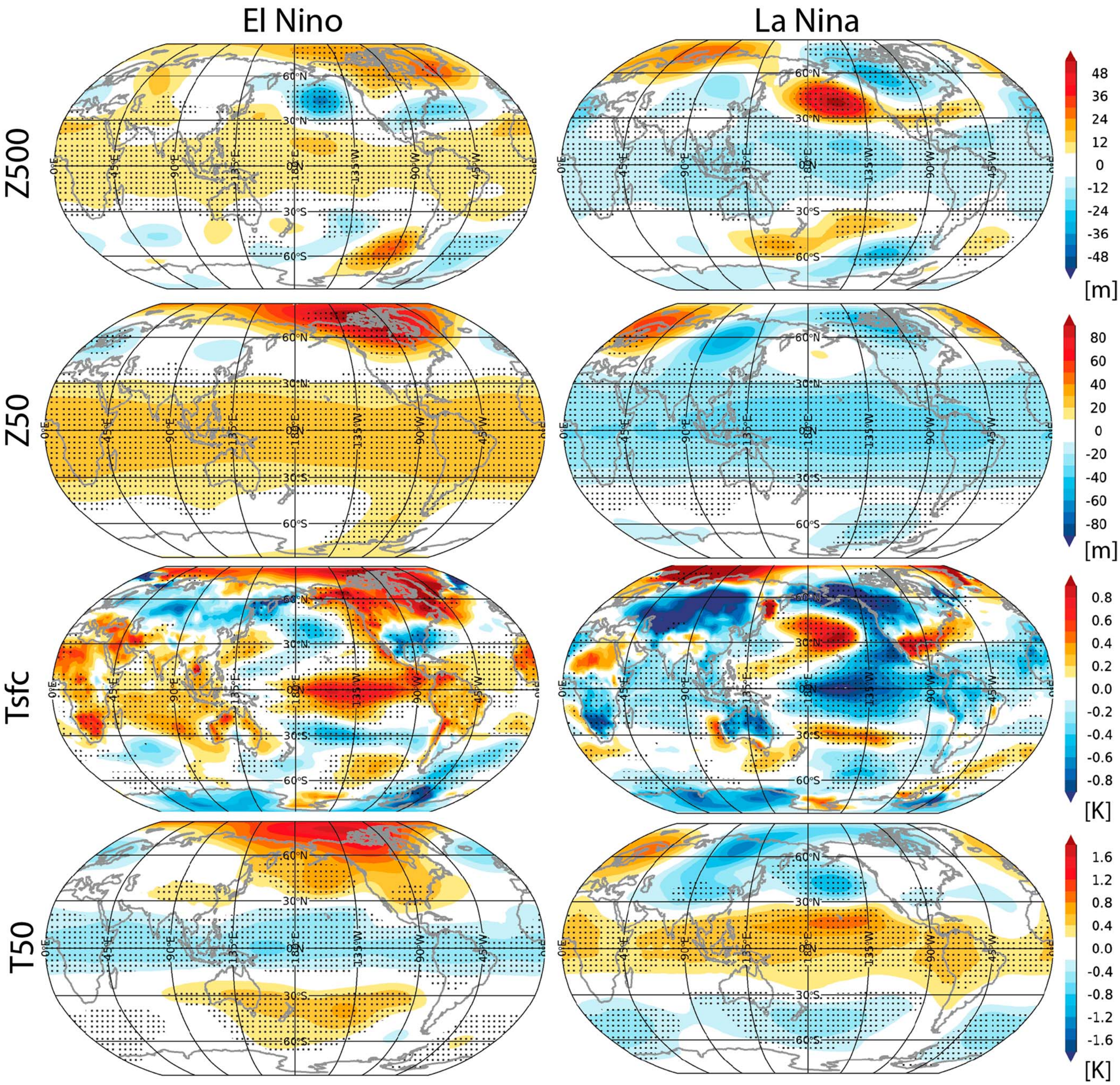

Figure 1. January-February-March composites of the El Niño (left) and La Niña (right) winters listed in Table 1 for geopotential height anomalies (shading, m) at 500 and 50 mb (top two rows) and temperature anomalies (shading, K) at the surface and at 50 mb (bottom two rows). The stippling indicates regions that are significantly different from the 1980–2010 climatology at the 95% level using a two-sided *t* test of the difference of means between El Niño and La Niña. Data is from JRA-55 reanalysis (Kobayashi et al., 2015), 1958–2016, and has been linearly detrended.

convective activity. El Niño and La Niña occur on irregular cycles of around 2–7 years (cp. Figure 2) and tend to peak in December. During the warm (El Niño) phase, the oceanic thermocline is flattened as the tropical trade winds relax, shifting the region of warm SSTs (Figure 3) as well as the regions of convective activity eastward in the equatorial Pacific. The tropical atmosphere is coupled to changes in SST through shifts in the Walker circulation (Walker, 1928), a set of zonal circulation cells in the tropics driven by atmospheric convection, giving rise to pressure changes across the tropical Pacific, the so-called *Southern Oscillation*. During La Niña, the easterly trade winds are strengthened and pile up warm water in the western part of the Pacific basin, while the thermocline moves closer to the surface in the eastern Pacific, allowing colder waters to reach the sur-

Table 1
All Years in the Historical Reanalysis, their ENSO State, and the Associated State of the Stratosphere and NAO Index

ENSO Year (Jan)	ENSO type (per CPC)	CP/EP for N3 v. N4	CP/EP Modoki	SSW	DJF Niño 3.4 (ONI)	DJFM Zonal wind anom (60°–80° N)	DJFM NAO index
1958	**El Niño**	EP	—	30 January 1958	**1.8**	−6.3	−0.9
1959	**El Niño**	CP	—	—	0.6	2.8	0.4
1960	Neutral	—	—	17 January 1960	−0.1	0.5	−0.6
1961	Neutral	—	—	—	0.0	−4.1	**1.2**
1962	Neutral	—	—	—	−0.2	**11.3**	−0.6
1963	Neutral	EP	—	30 January 1963	−0.4	0.7	**−1.1**
1964	**El Niño**	EP	—	—	**1.1**	**10.7**	**−1.5**
1965	**La Niña**	—	—	—	−0.6	**8.5**	−0.6
1966	**El Niño**	EP	—	18 December 1965, 23 February 1966	**1.4**	**−8.2**	−0.1
1967	Neutral	EP	**—**	—	−0.4	**16.2**	**1.0**
1968	Neutral	**EP**	**EP**	7 January 1968	−0.6	2.2	0.1
1969	**El Niño**	**CP**	**CP**	29 November 1968, 13 March 1969	**1.1**	−5.6	**−1.5**
1970	**El Niño**	EP	—	2 January 1970	0.5	**−10.2**	−0.3
1971	**La Niña**	EP	CP	18 January 1971, 20 March 1971	**−1.4**	−5.5	−0.7
1972	**La Niña**	EP	—	—	−0.7	**9.7**	**1.0**
1973	**El Niño**	EP	—	31 January 1973	**1.8**	−3.3	0.7
1974	**La Niña**	**CP**	**CP**	—	**−1.8**	5.8	0.8
1975	**La Niña**	—	—	—	−0.5	−3.9	0.7
1976	**La Niña**	—	CP	—	**−1.6**	**16.4**	0.9
1977	**El Niño**	EP	—	9 January 1977	0.7	**−8.3**	**−1.5**
1978	**El Niño**	**CP**	**CP**	—	0.7	−1.3	−0.4
1979	Neutral	—	—	22 February 1979	0.0	−4.2	**−1.3**
1980	**El Niño**	EP	—	29 February 1980	0.6	−1.5	0.0
1981	Neutral	—	**—**	4 March 1981	−0.3	2.2	0.4
1982	Neutral	—	**—**	4 December 1981	0.0	2.6	0.3
1983	**El Niño**	EP	**—**	—	**2.2**	5.2	**1.6**
1984	**La Niña**	CP	—	24 February 1984	−0.6	1.1	0.8
1985	**La Niña**	EP	—	1 January 1985	**−1**	**−8.6**	−0.7
1986	Neutral	EP	—	—	−0.5	5.6	0.7
1987	**El Niño**	EP	—	23 January 1987	**1.2**	**−10.5**	−0.3
1988	**El Niño**	—	—	8 December 1987, 14 March 1988	0.8	−0.1	0.4
1989	**La Niña**	**CP**	**CP**	21 February 1989	**−1.7**	4.7	**1.8**

face. A more detailed discussion of the ENSO phenomenon can be found in, for example, McPhaden (2015), Trenberth, (2001, 2017), and Timmermann et al. (2018).

A range of indices is currently in use to classify ENSO events: the Southern Oscillation Index (SOI) and the Oceanic Niño Index (ONI) (shown in Figure 2) by the NOAA Climate Prediction Center (CPC), the Multivariate ENSO Index (MEI; Wolter & Timlin, 1998), the Trans-Niño Index (TNI; Trenberth & Stepaniak, 2001), and the cold tongue index (CTI; Deser & Wallace, 1990), in addition to SST averages over a range of different regions in the tropical Pacific, denoted Niño1+2 (0–10° S, 90–80° W), Niño3 (5° N to 5° S, 150–90° W), Niño3.4 (5° N to 5° S, 170–120° W), and Niño4 (5° N to 5° S, 160° E to 150° W).

These indices are used to account for the significant variability among ENSO events: Every ENSO event differs in both its magnitude and in its location (also referred to as *ENSO flavor*) in the tropical Pacific where maximum SST anomalies are observed. The amplitude distribution is skewed toward stronger El Niño events as compared to La Niña events, which is often referred to as "asymmetry" or "nonlinearity" of ENSO (An, 2004; An & Jin, 2004). Three events in particular (1982/1983, 1997/1998, and 2015/2016, which is shown in Figure 4)

Table 1 (*continued*)

ENSO Year (Jan)	ENSO type (per CPC)	CP/EP for N3 v. N4	CP/EP Modoki	SSW	DJF Niño 3.4 (ONI)	DJFM Zonal wind anom (60°–80° N)	DJFM NAO index
1990	Neutral	—	—	—	0.1	**9.9**	0.8
1991	Neutral	CP	—	—	0.4	−0.1	0.6
1992	**El Niño**	EP	CP	—	**1.7**	0.1	0.8
1993	Neutral	—	**—**	—	0.1	**9.9**	**1.0**
1994	Neutral	—	**—**	—	0.1	3.7	**1.3**
1995	**El Niño**	**CP**	**CP**	—	**1**	4.4	**1.7**
1996	**La Niña**	EP	—	—	−0.9	**8.9**	**−1.3**
1997	Neutral	EP	**—**	—	−0.5	**12.7**	0.0
1998	**El Niño**	**EP**	**EP**	—	**2.2**	−5.8	−0.3
1999	**La Niña**	**CP**	**CP**	15 December 1998, 26 February 1999	**−1.5**	**−12.0**	0.3
2000	**La Niña**	EP	CP	20 March 2000	**−1.7**	**10.7**	**1.2**
2001	**La Niña**	**CP**	**CP**	11 February 2001	−0.7	**−9.3**	**−1.1**
2002	Neutral	EP	—	30 December 2001	−0.1	**−9.5**	−0.2
2003	**El Niño**	—	CP	18 January 2003	0.9	−2.7	−0.5
2004	Neutral	—	**—**	5 January 2004	0.4	**−8.7**	−0.2
2005	**El Niño**	**CP**	**CP**	—	0.6	5.3	−0.2
2006	**La Niña**	EP	—	21 January 2006	−0.8	**−13.3**	**−1.1**
2007	**El Niño**	—	—	24 February 2007	0.7	**8.3**	0.3
2008	**La Niña**	EP	CP	22 February 2008	**−1.6**	−1.1	0.0
2009	**La Niña**	**CP**	**CP**	24 January 2009	−0.8	**−8.1**	−0.8
2010	**El Niño**	**CP**	**CP**	9 February 2010, 24 March 2010	**1.5**	**−9.7**	**−3.1**
2011	**La Niña**	—	CP	—	**−1.4**	**17.3**	**−1.6**
2012	**La Niña**	**CP**	**CP**	—	−0.8	−1.0	**1.2**
2013	Neutral	EP	**—**	6 January 2013	−0.4	**−11.2**	**−1.6**
2014	Neutral	—	**—**	—	−0.4	1.9	0.3
2015	**El Niño**	**CP**	**CP**	—	0.6	6.3	**1.5**
2016	**El Niño**	EP	CP	—	**2.5**	2.1	0.7
2017	Neutral	—	—	—	−0.3	−1.1	−0.1

Note. The year in column 1 refers to the January of a given winter (e.g., 1958 refers to the 1957–1958 winter). Column 2 describes whether the winter is classified as an El Niño or La Niña winter based on ERSSTv5 data (Huang et al., 2017) and the centered Oceanic Niño Index from the NOAA Climate Prediction Center. Columns 3 and 4 give two methods of classifying events into central Pacific (CP) or eastern Pacific (EP), where rows in bold have identical classification between the two methods. The first method compares the Niño3 (N3) and Niño4 (N4) indices. CP events are defined where N4 is greater than 0.5 std dev, and N4-N3 is greater than 0.1 K. The second method uses the Modoki index (Ashok et al., 2007). CP (EP) events are defined where the Modoki index is greater (less) than 1 (−1). Column 5 gives the date of SSWs during a particular winter (Butler et al., 2017). Only SSWs detected in at least two reanalyses (out of JRA-55 [Kobayashi et al., 2015], ERA-40 [Uppala et al., 2006], and NCEP/NCAR [Kalnay et al., 1998] reanalysis prior to 1979; JRA-55, ERA-interim [Dee et al., 2011], and MERRA2 [Molod et al., 2015] after 1979) are included. Column 6 provides the DJF Niño3.4 sea surface temperature anomaly in degrees Celcius, where bold values are greater than +/− 1 °C. Column 7 is the DJFM-mean of the zonal wind anomaly from 60° to 80 ° N using JRA-55 reanalysis. Bold values are greater than +/− 1 standard deviation. Column 8 is the DJFM-mean NAO index, which has been detrended and standardized. Bold values are greater than +/− 1 standard deviation. ENSO = El Niño Southern Oscillation; NAO = North Atlantic Oscillation; CPC = Climate Prediction Center; SSW = Sudden Stratospheric Warming; DJF = December-January-February; ONI = Oceanic Niño Index.

were exceptionally strong. The asymmetry in the location is skewed toward a more eastward location of the SST anomalies for El Niño events (Capotondi et al., 2015; Dommenget et al., 2012), see also Figure 3. El Niño events tend to exhibit a wider range in the longitudinal location of the maximum SST anomaly in the equatorial Pacific: While La Niña events tend to peak further west in the equatorial Pacific, strong El Niño events can exhibit their maximum SST anomaly in the longitudinal range from the central Pacific to the very eastern equatorial Pacific off the coast of Ecuador (see Figure 3). Specifically, for Central Pacific (CP) El Niño events, SST anomalies peak in the central equatorial Pacific, also called dateline or Modoki El Niño events (Ashok et al., 2007; Larkin & Harrison, 2005; middle panel of Figure 3), while for East Pacific (EP or *canonical*) El Niño events, SST anomalies peak in the eastern equatorial Pacific (top panel of Figure 3), though the distinction is not

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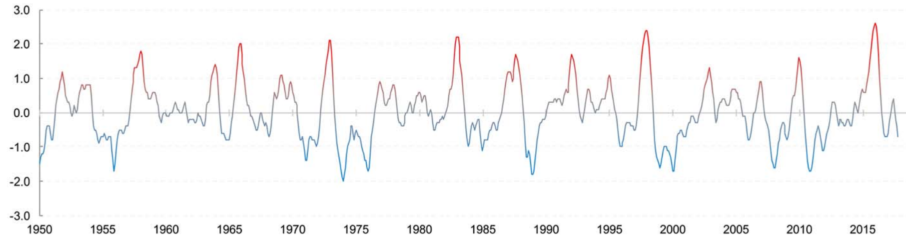

Figure 2. Time series of the Oceanic Niño Index (°C; 3-month averages from ERSST.v5 sea surface temperature [Huang et al., 2017] anomalies in the Niño3.4 region), based on centered 30-year base periods updated every 5 years for December-February 1950 to September-November 2017.

always unique and the distribution is nonbinary (Table 1). For distinguishing EP and CP ENSO events, common methods are based on the comparison of the Niño3 and Niño4 indices or the Modoki index (Ashok et al., 2007; see Table 1). CP events tend to be somewhat weaker than EP events (see Figure 3 and ; Johnson, 2013). For La Niña, the anomalous cooling tends to be weaker and located more centrally in the equatorial Pacific (bottom panel of Figure 3) as compared to the warming associated with EP El Niño. A possible cause for this behavior may be the (possibly nonlinear) modulation of the anomalies by atmosphere-ocean feedbacks (Bayr et al., 2017; Frauen & Dommenget, 2010)—hence the term nonlinearity. In this review, we will adopt the term asymmetry for differences between El Niño and La Niña and nonlinearity for processes deemed to show nonlinear behavior. For a more detailed review of ENSO diversity, see Capotondi et al. (2015) and Timmermann et al. (2018).

ENSO tends to be predictable at several months lead time, and skill increases rapidly after the so-called *spring predictability barrier* in April (Barnston et al., 2012; Duan & Wei, 2012; Luo et al., 2008; McPhaden, 2003). The first successful numerical ENSO forecasts were made in the 1980s (Cane et al., 1986; Zebiak & Cane, 1987), though predicting the occurrence and strength of an ENSO event remains a challenge, as, for example, exemplified by the poor prediction of the decay to a weak El Niño in 2014 and the extreme El Niño in 2015 by some modeling centers (S. Hu & Fedorov, 2016; McPhaden, 2015). For a review of ENSO prediction, see Barnston et al. (2012), Clarke (2014), Fedorov et al. (2003), Latif et al. (1998), and Tippett et al. (2012). The difficulty in predicting ENSO location and strength implies challenges in predicting the teleconnections arising from an ENSO event, since these are likely to depend on ENSO characteristics, as further discussed in section 2.2 for the troposphere and section 4 for the stratosphere.

2.2. Tropospheric Teleconnections of ENSO

This section briefly reviews the main ENSO impacts in the troposphere, with a focus on the impacts that are of relevance to the stratosphere. For a more general review of tropical to polar teleconnections, we refer the reader to X. Yuan et al. (2018) and the special collection “Connecting the Tropics to the Polar Regions” in Journal of Climate, published in 2014.

We first note some difficulties inherent to the understanding of ENSO teleconnections. Large internal atmospheric variability, particularly in the extratropics, can make it difficult to isolate the climate impacts of ENSO teleconnections. Deser et al. (2017) found that even ensemble model runs nudged to the same tropical SST anomalies (and thus the same ENSO phase and amplitude) produced substantial diversity in ENSO teleconnections due to internal atmospheric variability. Even if teleconnections are found in observations, this does not necessarily mean that models are always able to reproduce them (X. Yang & DelSole, 2012). The teleconnections may depend nonlinearly on the amplitude of the SST anomaly associated with the ENSO event. While stronger El Niño events tend to induce stronger teleconnections, these relationships are not necessarily linear (Hoerling et al., 1997, 2001; Jong et al., 2016). Precipitation in the tropical Pacific increases nonlinearly with SST and the meridional SST gradient (Cai et al., 2014; Chung & Power, 2015), and Hoerling et al. (2001) find nonlinear shifts in the precipitation response. However, nonlinearities in the tropical Pacific do not necessarily guarantee nonlinearities in the extratropical North Pacific response (Frauen et al., 2014). ENSO diversity may also have an influence on the teleconnections: the tropospheric responses to the two flavors (CP and EP) of El Niño clearly differ over Australia (Taschetto & England, 2009), in the tropical Atlantic (Taschetto et al., 2016), and in the North Pacific (Frauen et al., 2014; Garfinkel, Weinberger, et al., 2018). The differences in the stratospheric teleconnections due to ENSO diversity are discussed in section 4. Finally, ENSO teleconnections

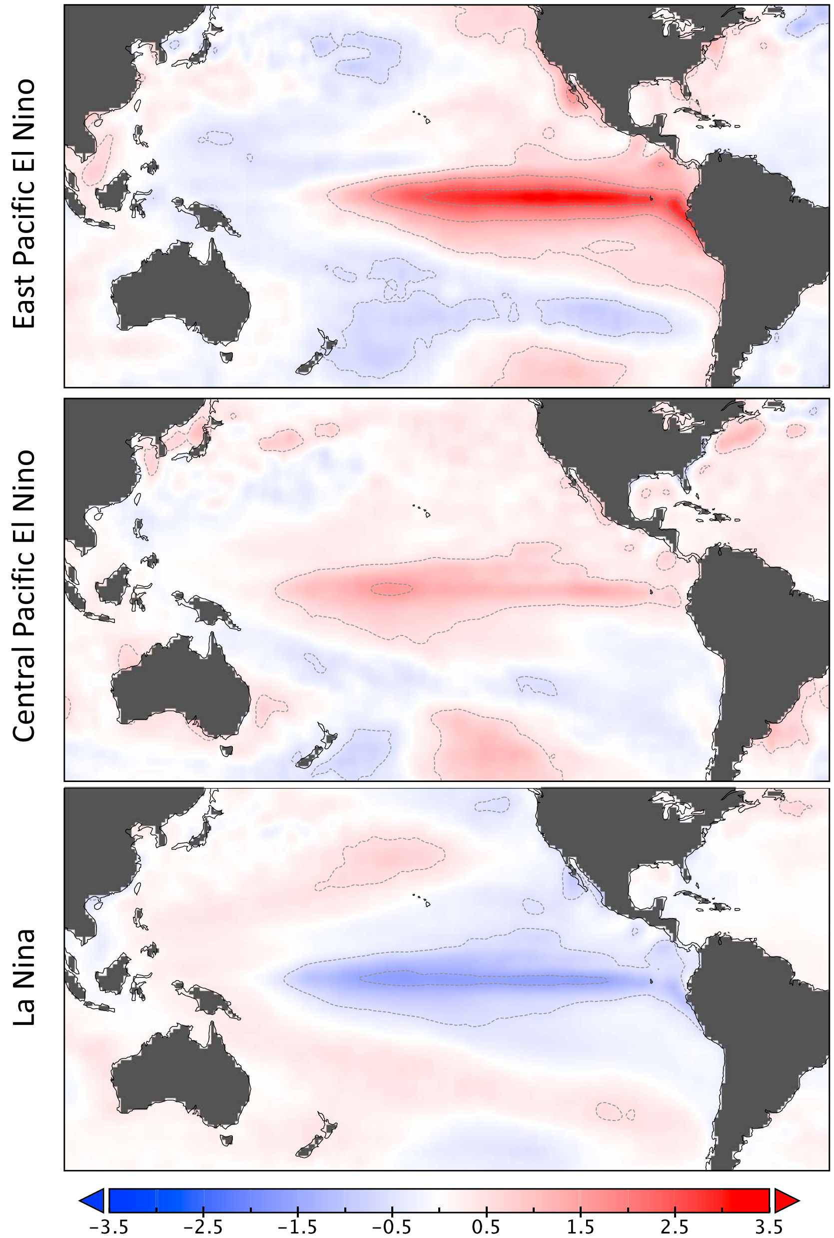

Figure 3. Sea surface temperature anomalies (°C) for November-January 1957–2017 for (top) East Pacific El Niño (12 events), (middle) Central Pacific El Niño (7 events), and (bottom) La Niña (18 events) given by the difference from climatology between 1981 and 2017 for the NOAA Optimum Interpolation sea surface temperature data set (Reynolds et al., 2002). Contours denote the location of the tickmarks on the color bar at intervals of 1 °C. The events were composited based on the Central Pacific/East Pacific classification by the Niño3/Niño4 method (third column in Table 1).

may exhibit nonstationary behavior and may be affected by long-term changes arising through internal or external forcing (see section 5).

2.2.1. Tropical Troposphere ENSO Teleconnections

ENSO has a profound impact on the tropical circulation in the Pacific, and impacts extend into the Indian (e.g., Murtugudde et al., 2000; Webster et al., 1999; Yu & Rienecker, 2000) and Atlantic Ocean basins (e.g., Giannini et al., 2001; Taschetto et al., 2016). Of particular relevance to the stratospheric response are the changes in the Walker circulation and in tropical waves, and we focus on these aspects of the response here.

The Southern Oscillation, the atmospheric component of ENSO (Bjerknes, 1969; Rasmusson & Wallace, 1983), primarily manifests as a seesaw in atmospheric pressure at sea level between the south-central Pacific and

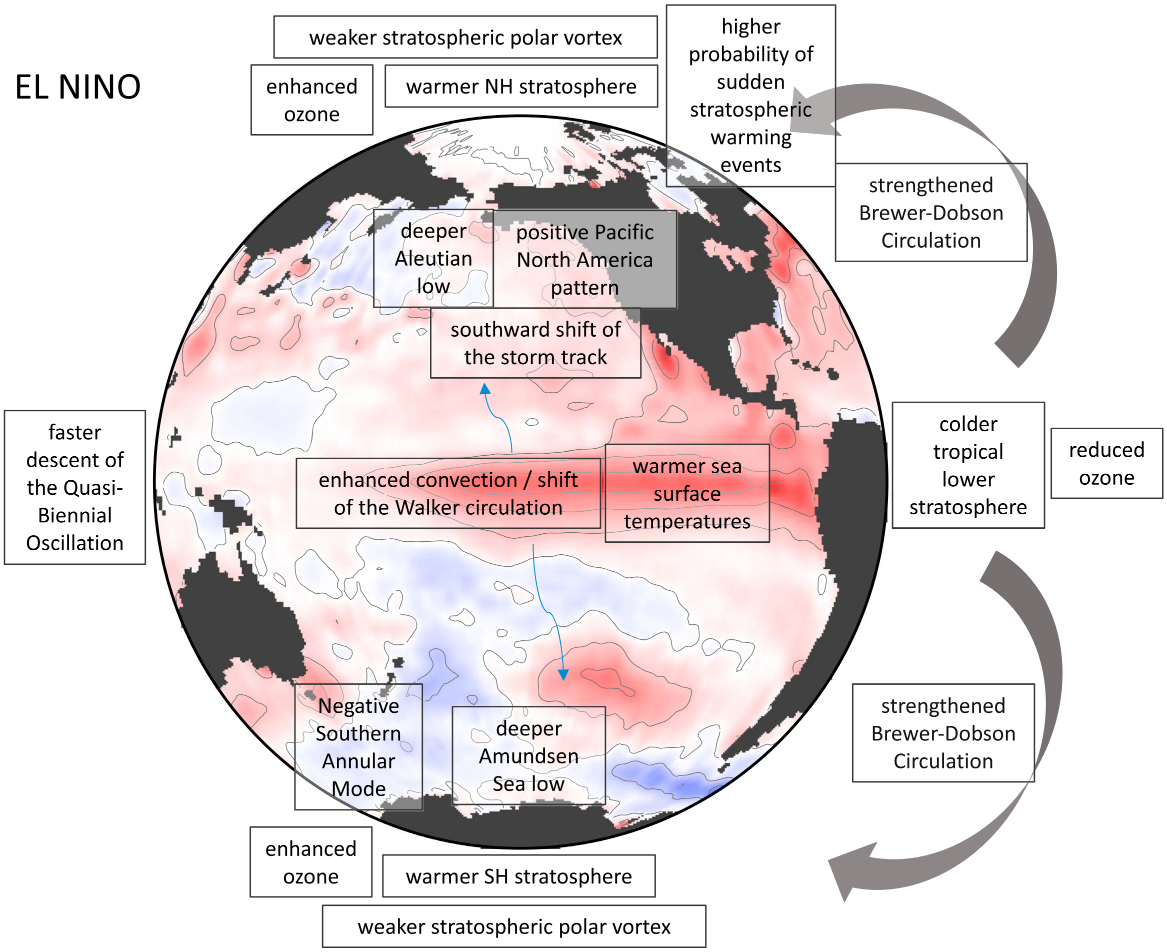

Figure 4. Schematic of the processes governing the El Niño remote response to the tropical and extratropical stratosphere. The colors indicate the sea surface temperature anomalies during November-January of the extreme 2015/2016 El Niño event, contours are drawn at 1 °C intervals. NH = Northern Hemisphere; SH = Southern Hemisphere.

the region of low pressure near northern Australia (Walker, 1928). During El Niño events, sea level pressure is anomalously high over northern Australia and anomalously low over the south-central Pacific, and vice versa during La Niña. This oscillation was originally discovered nearly 100 years ago (Walker, 1928) and is the surface pressure signature of a weakened Pacific Walker circulation during El Niño that is also evident in rainfall and winds (Rasmusson & Wallace, 1983).

Enhanced convection in the central and eastern Pacific associated with ENSO launches a Kelvin wave response to the east and a Rossby wave straddling the equator to the west following the classical theory of Gill (1980). These waves are manifested in the temperature field (Figures 5b and 5d); specifically, warm tropospheric temperatures near 140° W, 15° N, and 15° S are related via the *hypsometric equation* to upper-level anticyclones straddling the equator and dynamically are associated with descending motion in the free troposphere and associated adiabatic warming (Highwood & Hoskins, 1998; Yulaeva & Wallace, 1994). These upper-level anticyclones are centered near the longitude of the peak convective response and not to the west due in part to the presence of a mean flow (Jin & Hoskins, 1995). Temperature perturbations radiate away into the midlatitude troposphere, and these are also related via the hypsometric equation to the North and South Pacific geopotential height anomalies that will be discussed in the next subsections.

Additionally, ENSO has significant teleconnections to the tropical and subtropical Atlantic (García Serrano et al., 2017). The large-scale variability of the tropical Atlantic is governed by so-called Atlantic Niños, which are connected at a lag of several months to the tropical Pacific ENSO (Latif & Groetzner, 2014; Lübbecke, 2013; Lübbecke & McPhaden, 2013; Lübbecke et al., 2018) and which are suggested to have a feedback effect onto the tropical Pacific (Rodríguez-Fonseca et al., 2009). These ENSO-related impacts on the tropical Atlantic may

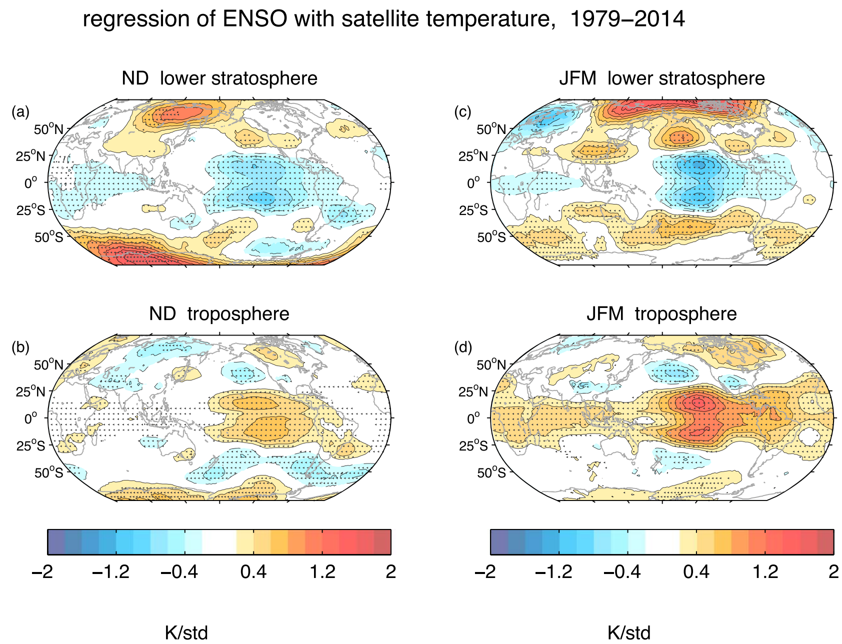

Figure 5. ENSO temperature signal from a multiple linear regression of the satellite temperatures over the period 1979 to 2014 for the (a and b) November-December and (c and d) January-March average. Areas where the signal is statistically significant at the 95% level are stippled. Data are sourced from the Advanced Microwave Sounding Unit (before 1998) and Microwave Sounding Unit (after 1988) homogenized by Remote Sensing Systems. (b) and (d) are for the total troposphere channel (updated from Fu & Johanson, 2005) and (a) and (c) for the lower stratospheric channel (updated from ; Mears & Wentz, 2009). Compare to Figure 11 of Randel and Cobb (1994). The contour interval is 0.2 K (standard deviation)$^{-1}$. The standardized Niño3.4 index is used as the ENSO predictor. The other predictors in the multiple linear regression are CO_2 concentrations, equivalent effective stratospheric chlorine, total solar irradiance, aerosol optical depth as in Aquila et al., 2016), and the two leading empirical orthogonal functions of zonal mean zonal wind from 100 to 7 hPa (i.e., the Quasi-Biennial Oscillation) in MERRA reanalysis. ENSO = El Niño Southern Oscillation.

influence European climate through a range of mechanisms including a modulation of the Atlantic Hadley Cell, providing another tropospheric pathway of ENSO to Europe (Rodríguez-Fonseca, 2016). It has been suggested that these tropospheric teleconnections could explain why the Euro-Atlantic response to ENSO appears nonlinear, that is, why the strongest El Niño events have not resulted in a strongly negative NAO but rather a different pattern (Toniazzo & Scaife, 2006; Wu & Hsieh, 2004).

2.2.2. NH Tropospheric ENSO Teleconnections

In the extratropics, ENSO teleconnections are mainly brought about by Rossby wave trains. These teleconnections are characterized by quasi-stationary high and low pressure anomaly patterns (see Figure 1) and can be maintained through midlatitude wave-mean flow interactions (Franzke & Feldstein, 2005). For more details on the mechanisms of global teleconnections, see Z. Liu and Alexander (2007), Nigam and Baxter (2015), and Stan et al. (2017). Theoretical studies (e.g., Hoskins & Karoly, 1981) have shown that the atmospheric response to a steady thermal forcing near the equator is a Rossby wave train emanating poleward from the tropics and arching eastward (cf. Figure 1). Longer wavelengths (i.e., planetary waves) propagate further poleward, while shorter wavelengths are trapped equatorward of the subtropical jet. These simple models of tropical thermal forcing reproduce remarkably well the general characteristics of observed ENSO teleconnections to the extratropics (Bjerknes, 1969; Wallace & Gutzler, 1981).

One of the dominant teleconnection patterns in the NH extratropics is the Pacific-North American Pattern (PNA; Leathers et al., 1991; Leathers & Palecki, 1992), which in its positive phase is characterized by a strengthened Aleutian low south of the Aleutian islands in the North Pacific, a positive pressure anomaly over western Canada and over the region around Hawaii, and a negative anomaly over the southeastern United States. El Niño (La Niña) is generally associated with the positive (negative) phase of the PNA and a southward (northward) shift of the Pacific storm track (Seager et al., 2010). Tropically forced Pacific teleconnections appear on a continuum of time scales; on intraseasonal time scales, the PNA is likely tied to the Madden-Julian Oscillation (MJO), while PNA-like patterns on interannual time scales are more closely related to ENSO (Johnson & Feldstein, 2010). Other teleconnections linked to ENSO include the North Pacific Oscillation (NPO), which has been related to central Pacific El Niño (Di Lorenzo et al., 2010). As shown in Figure 1, these teleconnections have direct downstream surface impacts, providing most of the predictive skill related to ENSO for, for example, North America (Ropelewski & Halpert, 1987). They also influence remote air-sea interactions, which feedback onto the atmospheric teleconnection (Alexander et al., 2002).

In the presence of a strong zonal subtropical jet, poleward propagating waves can become trapped and propagate eastward along the subtropical waveguide (Ambrizzi & Hoskins, 1997; Branstator, 2002). During central Pacific El Niño events, this *subtropical bridge* is suggested to drive a zonally extended wave train from the Pacific to the Atlantic, associated with a negative phase of the NAO (Graf & Zanchettin, 2012) and enhanced cyclogenesis in the Gulf Stream region (Schemm et al., 2016). The related increased (Y. Li & Lau, 2012) or more zonal (Drouard et al., 2013) transient eddy propagation from the North Pacific to the North Atlantic is also suggested to force a negative NAO phase (Jiménez-Esteve & Domeisen, 2018).

The tropospheric wave trains associated with ENSO convective heating also lead to changes in wave driving at both the subtropical and midlatitude tropopause. During an El Niño event, anomalous upward propagation and dissipation of planetary waves at middle and high latitudes and gravity waves in the subtropics (Calvo et al., 2010; Garfinkel & Hartmann, 2008) leads to a strengthening of the Brewer-Dobson circulation, resulting in a cooler tropical lower stratosphere and a warmer polar stratosphere. The details of these global stratospheric impacts are the focus of section 4.

2.2.3. SH Tropospheric Teleconnections

The Southern Hemisphere (SH) has been less studied as compared to the NH, although some of the most prominent and robust tropospheric teleconnections of ENSO impact the SH, in particular the South Pacific, South America, and Australia (Brands, 2017; see Figure 1).

The variability in the extratropical SH is dominated by the SAM, which has over the past decades experienced a positive trend (Thompson et al., 2000) associated with a strengthening of the westerlies at the poleward flank of the jet stream, that is, a poleward shift of the jet. This trend is likely due to a combination of increasing greenhouse gases and stratospheric ozone depletion (Dameris et al., 2014; Thompson et al., 2000). While the SAM exhibits a considerably more annular structure as compared to its NH counterpart, variability of the SAM is also dominated by semipermanent low pressure anomalies. These pressure centers consist of three large-scale low pressure anomalies around Antarctica and are dominated by the Amundsen Sea low in the subpolar South Pacific. This structure is closely related to the Pacific South America (PSA) pattern, the SH analog to the PNA (H. Ding et al., 2016; Mo & Paegle, 2001). Externally forced variability, that is, the anomalies arising from teleconnections, tends to project dominantly onto the PSA (Cavalcanti & Shimizu, 2012), that is, the anomalies manifest as one phase of the PSA.

El Niño projects onto a pressure dipole in the South Pacific (Karoly, 1989; van Loon & Madden, 1981; Figure 4), which is related to the PSA and the negative phase of the SAM (Lu et al., 2008; T. Zhou, 2004), that is, an equatorward expansion of the westerly winds and an equatorward shift of low pressure systems. Comparing to the NH tropospheric response reveals hemispheric symmetries in response to tropical forcing (Figure 1), as further documented in Seager et al., (2003, 2005), and Scaife et al. (2017). ENSO explains roughly 25% of SAM variability in austral summer but less in austral winter (L'Heureux & Thompson, 2006), though the exact seasonality and impact depends on the origin of the tropical Rossby wave train (Q. Ding et al., 2012; X. Li et al., 2015). El Niño also modifies the SH subtropical jets and transient wave activity (Simpson et al., 2011), and these changes have been shown to be important for the stratospheric impacts of El Niño as discussed in section 4.

The SH variability forced by ENSO is superimposed on long-term tropospheric variability and trends (Q. Ding et al., 2011, H. Ding et al., 2015; Fogt & Bromwich, 2006; Waugh et al., 2015), which may also be affected by tropical forcing (Allen & Kovilakam, 2017; Q. Ding & Steig, 2013; Grassi et al., 2005) as well as by stratospheric forcing, such as the surface response to stratospheric ozone depletion (Polvani et al., 2011).

3. A Brief Overview of the Stratospheric Circulation

The stratosphere is the layer of the atmosphere between about 10 and 50 km above the Earth's surface, located between the tropopause and the stratopause. It is considerably more stable as compared to the troposphere since it exhibits increasing temperature with height due to the presence of the ozone layer. Thus, it was once thought that understanding the stratosphere was not crucial for predicting tropospheric and surface dynamical variability. However, in recent decades, research has shown a distinct downward influence of stratosphere variability to the troposphere, with significant surface impacts (discussed in section 3.4). In addition, the chemical composition of the stratosphere influences the atmospheric radiative budget and hence surface climate (Ramaswamy et al., 1992; Solomon et al., 2010). In order to understand how the stratosphere responds to ENSO, it is important to understand the factors that influence the stratosphere as well as the mechanisms behind that influence. The following sections will give a brief overview of the stratospheric circulation of the tropics and the extratropics of both hemispheres in order to be able to evaluate ENSO impacts in the stratosphere. For further details on the dynamics of the stratosphere and stratosphere-troposphere coupling, please refer to Geller, (1979, 2010), Haynes (2005), Kidston et al. (2015), Plumb (2010), and Waugh and Polvani (2010).

3.1. The Tropical Tropopause Layer and the Tropical Stratosphere

The tropical tropopause layer and the tropical stratosphere are governed by large-scale rising motion, radiation, time-varying zonal winds, and mixing. We refer the reader to relevant Reviews of Geophysics articles that discuss these features extensively (Baldwin et al., 2001; Butchart, 2014; Fueglistaler et al., 2009; Holton et al., 1995; Waugh, 2002).

Briefly, the rising motion of air masses associated with the residual meridional stratospheric circulation, the upward branch of the so-called *Brewer-Dobson Circulation* (BDC), occurs within the tropics and is associated with cold temperatures near the tropical tropopause. The coldest temperatures occur over the West Pacific and Maritime continent in boreal winter (the *cold point*) and are comparable to those observed over the South Pole in winter (Butchart, 2014). The upwelling and cold temperatures in this region govern concentrations of water vapor and ozone in the lower stratosphere (Fueglistaler et al., 2009) and—via mixing with extratropical air masses—affect the lower stratospheric composition in the extratropics as well (Holton et al., 1995; Waugh, 2002). The tropical lower stratosphere exhibits pronounced zonal asymmetries that are driven by the convection beneath. Colder temperatures overlie regions of enhanced convection, while warmer temperatures overlie regions with reduced convective activity (Fueglistaler et al., 2009; Highwood & Hoskins, 1998) (Figures 5a and 5c). These zonal asymmetries in temperature are associated with zonal asymmetries in upwelling.

Finally, the waves that drive the zonally asymmetric patterns in temperature, along with smaller-scale gravity waves, drive the *Quasi-Biennial Oscillation* (QBO; Baldwin et al., 2001; Gray, 2010), the dominant mode of tropical stratospheric variability and predictable more than a year in advance (Scaife et al., 2014). The QBO manifests as downward propagating easterly and westerly wind regimes between about 20° S and 20° N with its maximum amplitude at the equator (Baldwin et al., 2001). Individual QBO wind cycles last anywhere from 22 to 34 months and average approximately 28 months (Baldwin et al., 2001). The QBO also affects tropical stratospheric temperatures and trace gas concentrations (Baldwin et al., 2001) and the stratospheric polar vortices through the so-called Holton-Tan effect (Anstey & Shepherd, 2014; Holton & Tan, 1980), which can lead to increased predictability of the extratropical stratosphere on subseasonal time scales (Garfinkel, Schwartz, et al., 2018).

3.2. The NH Stratosphere

The NH stratosphere is dominated by a polar vortex that forms in boreal autumn and decays in spring. The term *polar vortex* in this review always applies to the stratospheric polar vortex; for a disambiguation of the tropospheric and stratospheric polar vortices, see Waugh et al. (2017). In the midstratosphere at 10 hPa and 60° N, the vortex reaches westerly wind speeds of around 35 m/s in the zonal average for the January mean. Locally, significantly higher wind speeds can be reached. The winds in the NH polar stratosphere are highly variable from November to March, largely due to the perturbations and the mixing induced by large-scale planetary waves, predominantly zonal wavenumbers 1–3 (McIntyre & Palmer, 1984). These waves transport momentum from their source region to the region where they break, slowing down the westerly winds in the region of wave breaking. Stationary waves, that is, waves that do not move in the zonal direction, can only propagate vertically into the stratosphere when the zonal winds are westerly and below some critical threshold (Charney & Drazin, 1961), which is why they are inhibited in their propagation during summer, when easterly winds prevail in the extratropical stratosphere (Plumb, 1989). This propagation window is shifted for zonal phase speeds different from zero, which can lead to enhanced wave propagation ahead of SSW events

(Domeisen, Martius, et al., 2018). Wave breaking in addition contributes to the driving of the Brewer-Dobson circulation, which is associated with downwelling over the polar cap (Butchart, 2014).

As breaking planetary waves lead to a weakening of the stratospheric zonal winds around the polar cap, the wintertime westerly zonal-mean zonal winds at 10 hPa and 60° N can fully reverse to easterly, which is termed a major SSW event. These events occur roughly 6 times per decade in the NH (Charlton & Polvani, 2007), though there is considerable decadal variability (as noted in, e.g., van Loon & Labitzke, 1993), with decades of only a few SSW events (e.g., the 1990s exhibited only two SSW events) and decades of high SSW frequency (e.g., there were nine SSW events during the 2000s; see Table 1 and Butler et al., 2017). Note that the 1990s were also an exceptional period for the tropical Pacific, with very weak ENSO amplitude in the first part of the decade (Latif et al., 1997), followed by the record 1997/1998 event, though it is unclear if there is a connection to the observed lull in SSW events during the 1990s.

During a SSW event, the stratospheric polar vortex is either displaced off the pole (*displacement* or *wave-1* event) or split into two daughter vortices (*split* or *wave-2* event; Mitchell et al., 2011). An event during which the vortex is perturbed and the meridional temperature gradient over the polar cap reverses but the zonal mean circulation at 10 hPa and 60° N does not reverse to easterlies is often termed a *minor warming*. When sunlight returns to the NH polar regions in spring, it radiatively reduces the stratospheric meridional temperature gradient, leading to a breakdown of the vortex in a *final warming* (e.g., Black et al., 2006), which generally occurs between early March to May. The extratropical stratospheric winds then remain easterly throughout boreal summer.

3.3. The SH Stratosphere

The SH winter stratosphere is considerably colder and relatively undisturbed as compared to the NH (Plumb, 1989) and exhibits massive ozone loss during spring, known as the *ozone hole*. For more details on the theory and history of ozone depletion, see P. A. Newman (2010) and Solomon (1999). While ozone depletion also occurs in the NH (e.g., Dameris et al., 2014; Manney et al., 2011; Weber et al., 2018), the SH is much more strongly affected due to the weaker tropospheric wave forcing, which permits a stronger SH polar vortex and a colder polar cap. The colder temperatures in the SH allow for polar stratospheric cloud (PSC) formation, so that the returning sunlight in spring initiates the chain reaction of heterogeneous ozone destruction induced by chlorofluorocarbons (CFC; Molina & Rowland, 1974). The isolation of the southern polar cap throughout winter and into spring by the strong polar vortex prevents an ozone influx from lower latitudes that could replenish the ozone that is chemically depleted in spring. Stratospheric ozone levels are expected to recover in the latter half of this century due to the *Montreal Protocol* that was adopted in the 1980s in order to protect the ozone layer by phasing out the production of CFCs. While the first possible signs of recovery of the Antarctic ozone hole have recently been suggested (Solomon et al., 2016; Strahan & Douglass, 2018), detection of recovery is difficult to attribute to reduced CFCs and is still debated (Ball et al., 2018; Chipperfield et al., 2017). Overall, ozone depletion has led to a cooling of the SH stratosphere in austral spring and summer (Randel & Wu, 1999), while a warming trend with significant zonal asymmetries has been observed since 1979 in the polar SH stratosphere in austral winter, likely linked to SST warming and increased wave propagation from the troposphere (Y. Hu & Fu, 2009; Wang et al., 2013).

The seasonal evolution of the SH stratosphere is not exactly opposite to the NH due to the different wave dynamics and ozone effects. In particular, the polar vortex at the 10-hPa level in the SH stratosphere on average reaches its maximum speed of about 80 m/s (in the zonal average at 60° N) in August and returns back to easterlies in November; thus, the SH polar vortex is significantly stronger than its NH counterpart. The rather undisturbed nature of the SH midwinter stratosphere translates into the frequency of occurrence of major SSWs: While minor warming events are observed in the SH (e.g., Labitzke, 1981; Quiroz, 1966), only one major SSW event (according to the NH classification) has so far been observed, in September 2002 (Butler et al., 2017). This event was characterized as a split SSW. The anomalous winter of 2002 has been widely investigated (see, e.g., the special issue by the Journal of Atmospheric Sciences, Charlton et al., 2005; Kushner & Polvani, 2005; P. A. Newman & Nash, 2005). For reviews of the SH stratosphere before the 2002 SSW event, see also the following reviews: Labitzke and van Loon (1972) and Randel and Newman (1998). Due to the relatively quiescent nature of the SH stratospheric winter, the timing of the final warming is not as variable in the SH compared to the NH (Black et al., 2006; Waugh & Rong, 2002). However, a trend toward a later breakup of the polar vortex due to the stratospheric cooling trend has been observed (Waugh et al., 1999), with potentially important implications for the SH troposphere (Byrne et al., 2017; Sheshadri et al., 2014; L. Sun et al., 2014).

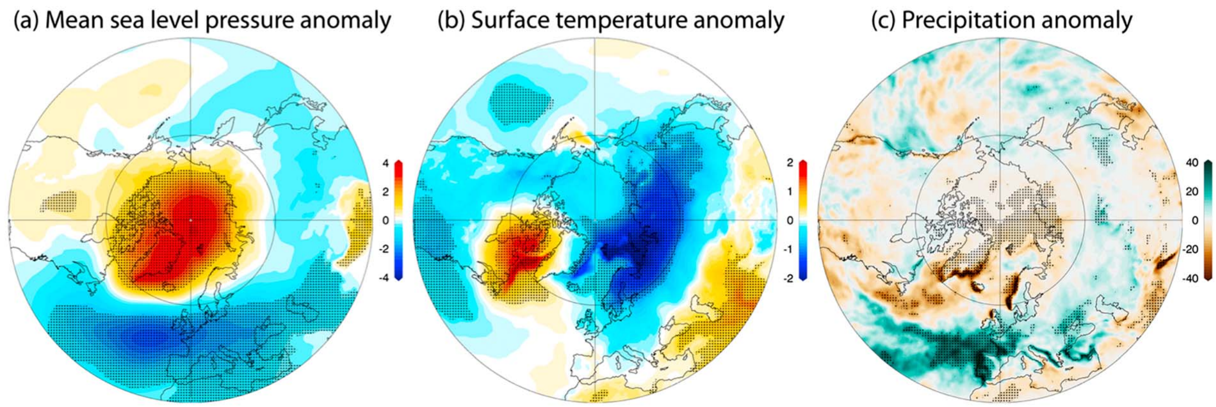

Figure 6. Composites of the 60 days after historical sudden stratospheric warming events in the JRA-55 reanalysis (Kobayashi et al., 2015) for (a) mean sea level pressure anomalies (hPa), (b) surface temperature anomalies (K), and (c) precipitation anomalies (mm). The stippling indicates regions that are significantly different from the climatology at the 95% level. Figure from Butler et al. (2017). Reprinted with permission.

3.4. The Impact of the Stratosphere on Surface Climate and Predictability

Research on the stratosphere has been motivated not only by ozone depletion but recently also by its coupling to surface climate. Specifically, it has been found that the winter and spring stratosphere can have a strong impact on surface weather and climate (e.g., Baldwin & Dunkerton, 1999, 2001; Black et al., 2006; Black & McDaniel, 2007; Thompson & Wallace, 1998). The stratosphere is suggested to be a crucial source of predictability on a variety of time scales from subseasonal and seasonal (e.g., Scaife et al., 2016; Tripathi, Charlton-Perez, et al., 2015; Tripathi, Baldwin, et al., 2015) to decadal (e.g., Garfinkel, Son, et al., 2017; Ivy et al., 2014; Kidston et al., 2015; Kretschmer et al., 2018; Reichler et al., 2012) and longer time scales (Omrani et al., 2014; Scaife et al., 2005; Waugh et al., 2015). For a more detailed review of the stratospheric impact on tropospheric variability and predictability, see Gerber et al. (2012).

In the NH, surface impacts arise when the polar vortex is dynamically perturbed, as, for example, for a SSW event. While SSW events tend to be predictable on weather time scales (e.g., Gerber et al., 2009; Karpechko, 2018; Tripathi, Baldwin, et al., 2015), the stratospheric anomalies then tend to impact weather at the surface for several weeks to months by increasing the likelihood of a negative phase of the NAO and the NAM (see Figure 6a), although not all SSW events have a surface impact (Karpechko et al., 2017; Kodera et al., 2016). A negative phase of the NAO is associated with colder weather over the eastern United States, northern Europe, and Siberia, warmer weather over Greenland, and wet anomalies over western Europe (Thompson et al., 2000; see Figures 6b and 6c). An anomalously strong polar vortex can affect the troposphere in the opposite sense, by increasing the chances of a positive NAO (e.g., Tripathi, Charlton-Perez, et al., 2015). This stratospheric influence has been shown to improve predictions of the extratropical NH in boreal winter for both statistical models (e.g., Christiansen, 2005; Karpechko, 2015; Wang et al., 2017) and dynamical models (Scaife et al., 2016) and for combinations thereof (Dobrynin et al., 2018), in particular over Europe (Butler et al., 2014). While deterministic predictability of the NAO in weather models is limited to a few days—though with a theoretical limit of about 3 weeks (Buizza & Leutbecher, 2015; Domeisen, Badin, et al., 2018)—on longer time scales, remote effects can improve statistical predictability. Medium-term predictability on time scales of weeks to months is however often limited to winters with strong stratospheric variability such as SSW events, in particular when considering remote effects from ENSO (Butler et al., 2016; Domeisen et al., 2015). At the end of winter, the stratospheric final warming is found to project onto the NAO more strongly in cases where the final warming occurs first in the midstratosphere as opposed to the upper stratosphere, which can increase surface predictability (Hardiman et al., 2011).

The mechanism of downward impact on the troposphere is not yet fully resolved. Within the stratosphere, the downward influence is mediated through wave-mean flow interaction (Plumb & Semeniuk, 2003). The impact of the lower stratospheric anomalies on the troposphere is suggested to arise and be maintained through a tropospheric synoptic eddy feedback (Domeisen et al., 2013; Kunz & Greatbatch, 2013; Plumb, 2010; Y. Song & Robinson, 2004), planetary waves (Hitchcock & Simpson, 2014, 2016; Smith & Scott, 2016), and the adjustment of the balanced circulation (Thompson et al., 2006). Tripathi, Baldwin, et al. (2015) and Kidston et al. (2015) provide overviews of the proposed mechanisms of downward impact of the extratropical stratosphere onto the troposphere.

In the SH, the downward impact is dominated by the influence of ozone. Since ozone exhibits both a strong seasonal cycle, due to depletion occurring predominantly in SH spring, and a long-term trend, the surface signature consists of both seasonal variability and multidecadal trends, respectively (Polvani et al., 2011; Seviour et al., 2014; Son et al., 2008, 2013; Thompson & Solomon, 2002, 2005; Thompson et al., 2011; Waugh et al., 2015). In fact, ozone depletion is suggested to be responsible for the dominant part of the poleward shift of the tropospheric jet stream in the SH (Gerber et al., 2012; Polvani et al., 2011). Downward coupling can also be found through interannual variability (Kuroda & Kodera, 1998) and through the impact of the final warming (Black & McDaniel, 2007; Byrne et al., 2017; L. Sun et al., 2014).

In the tropics, the QBO has recently been found to have a possible downward influence on the tropical troposphere, in particular on the MJO (Geller et al., 2017; Marshall et al., 2016; Son et al., 2017; Yoo & Son, 2016). The QBO also has a downward impact on the subtropical jet and subtropical precipitation (Garfinkel & Hartmann, 2011; Haigh et al., 2005; Hansen et al., 2016; Seo et al., 2013) as well as polar stratospheric variability (Anstey & Shepherd, 2014; Garfinkel, Schwartz, et al., 2018; Holton & Tan, 1980). Finally, wind and temperature variability in the tropical tropopause layer governs the chemical composition of the stratosphere and therefore Earth's radiative budget, as tropospheric air parcels enter the stratosphere through the tropics. Relatively small changes in the concentration of stratospheric water vapor can in turn affect global surface temperatures (Butchart, 2014; Fueglistaler et al., 2009; Solomon et al., 2010).

The increased awareness of stratospheric influence on surface weather and its predictability have led to the desire to understand remote drivers of stratospheric variability. ENSO is a prime example of an influence onto the global stratosphere that may in turn influence surface weather. The ENSO teleconnections to the stratosphere are discussed in section 4.

4. ENSO Teleconnections to the Stratosphere

4.1. ENSO Teleconnections to the Tropical Stratosphere

4.1.1. Changes in Wave-Driving and the BDC

ENSO modulates tropical stratospheric temperatures and upwelling through a variety of pathways. First, ENSO modifies the spectrum and source region of waves generated by convection, and the anomalous waves modify stratospheric temperatures directly. As discussed in section 2.2.1, El Niño leads to enhanced convection and warming in the tropical central Pacific peaking near 140° W. This warming in the troposphere is associated with lower stratospheric cooling aloft (Figures 5a and 5c; Gettelman et al., 2001; Kiladis et al., 2001; Yulaeva & Wallace, 1994; Randel & Cobb, 1994; Randel et al., 2000; Scherllin-Pirscher et al., 2012; X. L. Zhou et al., 2001). The stratospheric temperature anomalies (relative to the zonal mean) thus have the opposite sign to the tropospheric temperatures. The lower stratospheric cooling near 140° W is related to upward vertical motion concentrated in this sector (Figure 7; Highwood & Hoskins, 1998) that extends above where the diabatic heating due to moist convection ends. The extension of the upwelling anomalies above the level of peak diabatic heating is related to the upward propagation of the Rossby waves forced by El Niño (Norton, 2006; Taguchi, 2010a).

Second, ENSO modulates the zonal mean stratospheric circulation and temperature via its effect on the BDC. All studies focusing on the connection between ENSO and the BDC have found that El Niño leads to a strengthened BDC (Garfinkel, Gordon, et al., 2018; Hardiman et al., 2007, 2017; Randel et al., 2009; Taguchi, 2010a, 2008). The associated enhanced upwelling in the deep tropics during El Niño is associated with tropical stratospheric cooling (and anomalous downwelling in the polar stratosphere and a warmer pole, as discussed in section 4.2). Figure 8 shows the temperature response in boreal wintertime to El Niño relative to La Niña using radiosonde measurements (Free & Seidel, 2009). Similar changes are evident in reanalysis data (Haigh, 2003) and in models (Hardiman et al., 2007; Randel et al., 2009). These changes in the tropics are most pronounced in boreal winter. The zonally symmetric cooling during El Niño is superimposed on the regional-scale changes driven by the changes in the location of peak convection (cf. Figures 5a and 5c).

Four main classes of waves have been proposed to be important for the increase in tropical upwelling observed and modeled during El Niño.

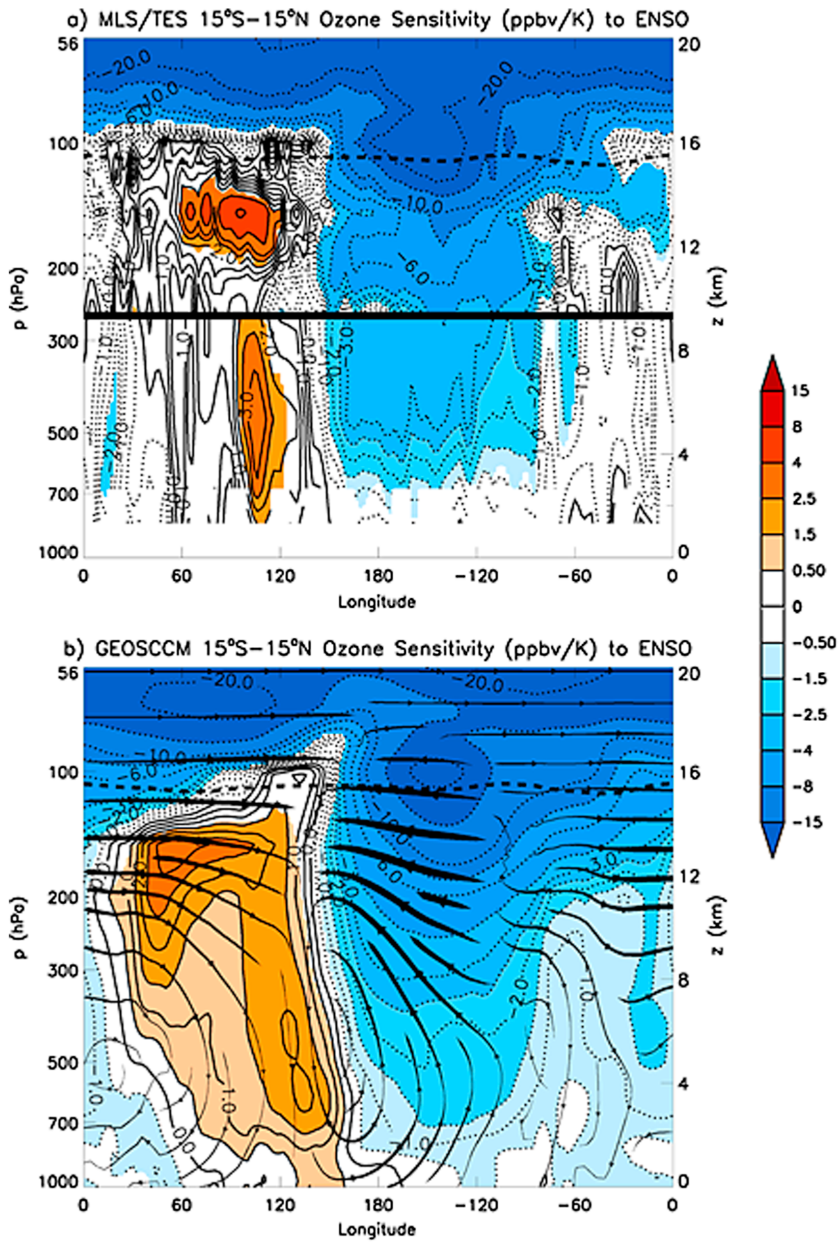

Figure 7. (a) Sensitivity coefficients (ppbv/K) to the Niño 3.4 Index from a multiple linear regression using deseasonalized tropical (15° S–15° N) average ozone from the Microwave Limb Sounder (MLS; Livesey et al., 2011) and Tropospheric Emission Spectrometer (TES; Nassar et al., 2008) onboard NASA's Aura satellite. The thick black line in (a) at 261 hPa denotes the transition from TES measurements below and MLS above. (b) GEOSCCM sensitivity coefficients resulting from the linear regression of ozone against the Niño3.4 index, over the same location. Overlaid is the anomalous circulation shown by the streamlines formed by regressing the zonal wind and vertical velocity against the Niño3.4 index. Shaded regions are significant above 2 standard deviations, and the black-dashed curve shows the mean model tropopause on both panels. Figure from Oman et al. (2013). Reprinted with permission.

1. The change in tropical/subtropical stationary waves noted in the previous paragraph (i.e., the zonal asymmetry between 140° W and 160° E) can impact the BDC (e.g., Taguchi, 2010a), though this effect was not found to be important by Simpson et al. (2011) for explaining differences in the BDC during El Niño.
2. As further discussed in section 4.2, constructive interference between the wave train generated by El Niño and the extratropical stationary waves (García-Herrera et al., 2006; Garfinkel & Hartmann, 2008; Hamilton, 1993b; Ineson & Scaife, 2009; Manzini et al., 2006) contributes to the wave driving of the tropical upwelling.

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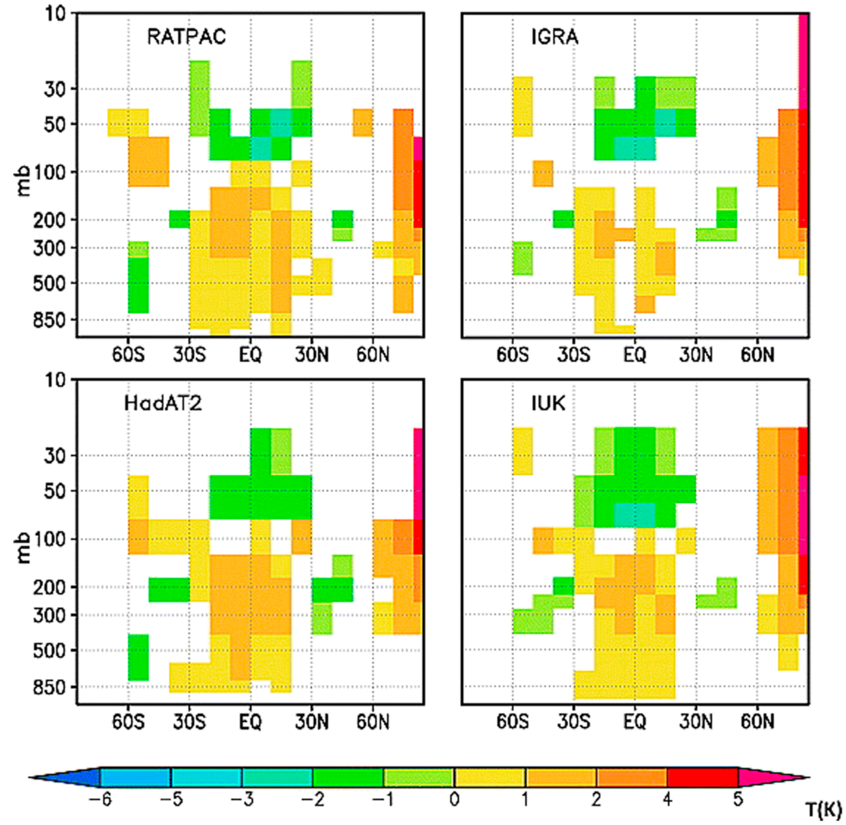

Figure 8. Temperature signal (K) from a linear regression associated with El Niño Southern Oscillation (using the Niño3.4 index) using December, January, and February in four distinct radiosonde data sets, shown for areas where the signal is statistically significant at the 95% level. From 80° to 90° S data are missing for the lower troposphere in all data sets. Figure from Free and Seidel (2009). Reprinted with permission.

3. El Niño leads to enhanced gravity wave generation from orographic features (e.g., mountains) in the NH due to the intensification and upward extension of the subtropical jet and hence allows for increased wave driving in the lower stratosphere (Calvo et al., 2010). This effect is most prominent for upwelling north of the equator (Simpson et al., 2011).
4. Finally, El Niño leads to enhanced transient synoptic-scale wave drag in the SH subtropical lower stratosphere due to an enhanced upward flux of wave activity from the troposphere into the lower stratosphere between 20° and 40° S (Simpson et al., 2011).

While there is still some uncertainty as to the precise composition of the tropical and subtropical waves that drive tropical upwelling, there is a consensus that processes local to the tropics and subtropics contribute to the upwelling, as might be expected for dynamical reasons (Grise & Thompson, 2013; Ortland & Alexander, 2014; T. Zhou et al., 2012). We note that even an idealized aquaplanet model that lacks continents, a warm pool, or topography in its default configuration simulates a strengthening of the shallow branch of the BDC and a weakened polar vortex in response to a zonally localized ENSO-like tropical SST forcing (H. Yang et al., 2014).

The tropical stratospheric response to La Niña is opposite to that of El Niño in winter in both models and observations (Calvo et al., 2010; Garfinkel, Gordon, et al., 2018); however, in spring, the modeling results of Garfinkel, Gordon, et al. (2018) suggest that El Niño and La Niña teleconnections are nonlinear with both strong El Niño and La Niña leading to tropical lower stratospheric warming and only moderate El Niño leading to cooling.

In addition to the differences between El Niño and La Niña, the tropical stratospheric wave response differs between EP and CP El Niño: The shift in the longitude of peak convection also leads to a shift in the Rossby and Kelvin wave response in both observational (D. Sun et al., 2013) and modeling studies (Garfinkel et al., 2013). There is no consensus as to which El Niño flavor impacts the zonally symmetric tropical response more strongly: the model experiments of Zubiaurre and Calvo (2012), F. Xie et al. (2014), and Garfinkel, Gordon, et al.

(2018) disagree as to which flavor more strongly modulates the BDC. The different conclusions reached by these studies likely reflects sensitivity to the precise nature of the SST anomalies imposed, and specifically it is important to consider the relative weakness in the underlying forcing for CP El Niño when interpreting the relative weakness of the stratospheric response (Garfinkel, Gordon, et al., 2018; F. Xie et al., 2012).

4.1.2. Modulation of the QBO

ENSO also impacts zonal wind in the tropical lower stratosphere: The wind anomalies associated with the QBO propagate downward more rapidly during El Niño than during La Niña (Taguchi, 2010b). As noted by Taguchi (2010b), this effect cannot be due to the effects of ENSO on the BDC, as the strengthened BDC during El Niño would, in isolation, lead to slower downward propagation. Rather, it is likely that ENSO modulates the waves generated within the troposphere that then propagate up and drive the QBO. In satellite and reanalysis data, atmospheric Kelvin wave activity increases during El Niño events, leading to a faster downward propagation of the westerly flow regime, in particular if conditions lower in the stratosphere allow for Kelvin wave propagation (Das & Pan, 2016). In the model experiments of Schirber (2015), both resolved and parameterized wave driving increases during El Niño due to enhanced convective activity, leading to faster downward propagation specifically for the westerly QBO regime.

The observed wind anomalies associated with the QBO are weaker during El Niño than during La Niña (Taguchi, 2010b), though this effect is only evident after 1990 (Geller et al., 2016). The strengthening of the QBO wind anomalies during La Niña may be associated with deeper convection that launches a broader spectrum of gravity waves, though this strengthening of the deepest convection during La Niña is again only evident after 1990 (Geller et al., 2016). The model experiments of Schirber (2015) suggest that La Niña may accelerate the easterly QBO regime while El Niño may accelerate the westerly regime. This raises the question if there is a relationship between ENSO and any specific phase, either easterly or westerly, of the QBO. Christiansen et al. (2016) find that strong El Niño events can lead to phase locking of the QBO: an ensemble of model simulations with observed SSTs simulated similar winds in the equatorial stratosphere in the 2 to 4 years after the strong ENSO event in 1997/1998 resembling those observed. In contrast, no such alignment is found in a parallel model ensemble forced with climatological SSTs. However, there is no indication that El Niño events in general lead to a preferred phasing of the QBO over the entirety of the historical record, as the connection between El Niño and specific QBO phases has not been constant in time (Christiansen et al., 2016; Garfinkel & Hartmann, 2007; Z.-Z. Hu et al., 2012; Huang et al., 2012). Finally, it has been speculated that the recent strong 2015/2016 El Niño event contributed to the recent disruption in the regularity of the QBO (Coy et al., 2017; Osprey et al., 2016) by modifying subtropical winds and thereby allowing for extratropical Rossby waves to propagate all the way to the equator (Barton & McCormack, 2017; Coy et al., 2017).

4.2. ENSO Teleconnections to the NH Stratosphere

Here we review the mechanisms driving the remote ENSO teleconnection to the NH polar stratosphere. The first part of the pathway from the tropical Pacific to the NH stratosphere concerns the North Pacific response to tropical Pacific forcing: the Aleutian low is deepened during El Niño events (e.g., Barnston & Livezey, 1987). The second part of the pathway concerns the upward propagation of Rossby waves into the stratosphere: during El Niño, the deepened Aleutian low strengthens the wave flux into the stratosphere through constructive linear interference with the climatological stationary wave pattern (Garfinkel & Hartmann, 2008; Hamilton, 1993b; Ineson & Scaife, 2009; Manzini et al., 2006; Smith & Kushner, 2012).

In contrast, during La Niña, tropical convection is shifted westward in the tropical Pacific, leading to a weakening of the Aleutian low and destructive linear interference with the climatological wave pattern in reanalysis data (Iza et al., 2016). Figure 9 shows that strong La Niña winters are associated with a stronger and colder than normal NH polar vortex, due to weakened stratospheric wave driving. The weakening of the Aleutian low for La Niña results in a more even distribution between the climatologically stronger Aleutian low and the Icelandic low, and therefore a more dominant wave-2 propagation into the stratosphere, compared to El Niño, which is associated with zonal wavenumber 1 (Barriopedro & Calvo, 2014; Garfinkel & Hartmann, 2008; Y. Li & Lau, 2013) due to the deepening of the Aleutian low. The fingerprints of this modulation in wavenumber 1 are evident in Figure 5b, as the warming over the pole is strongest from 60° to 300° E (Garfinkel & Hartmann, 2007).

The anomalous upward propagating planetary waves during El Niño amplify vertically into the stratosphere and break, which weakens the polar vortex, warms the stratospheric polar cap, and strengthens the Brewer-Dobson circulation (see Figures 1 and 5; Calvo et al., 2010; García-Herrera et al., 2006). The anomalous

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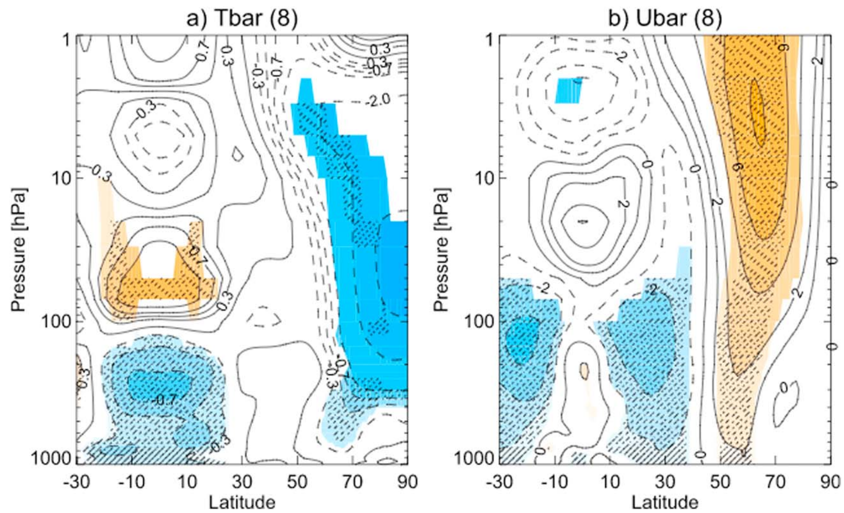

Figure 9. Composites of the December-January-February (a) zonal mean temperature and (b) zonal wind anomalies for strong La Niña events (smaller than −1 standard deviations). Contour intervals for temperature are +/− 0.3, 0.5, and 0.7 K up to +/−1 K and every 1 K thereafter. Contours for zonal wind are +/− 1 m/s up to +/− 2 m/s and every 2 m/s thereafter. Solid (dashed) contours denote positive (negative) anomalies. Numbers in brackets indicate the number of winters in each composite. Colors indicate areas significant at the 90% confidence levels, and stippling indicates significance at the 95% level. Figure from Iza et al. (2016), © American Meteorological Society. Used with permission.

downwelling at polar latitudes is associated with a warming of the NH polar stratosphere by up to 4 K in radiosonde measurements (Figure 8; Free & Seidel, 2009; Randel et al., 2009). Satellite measurements (Figure 5) and reanalysis data (Camp & Tung, 2007; Garfinkel & Hartmann, 2007), Figure 1) also show warming in the lower stratosphere at both poles during El Niño. The increased wave driving can lead to SSW events, which are preceded by extreme episodes of upward wave flux (Polvani & Waugh, 2004). The induced temperature and wind anomalies can first be observed in the upper stratosphere in early winter and can then be observed to descend to the lower stratosphere and the surface by late winter and early spring (Manzini et al., 2006 and Figure 10). After SSW events, these anomalies can contribute to a negative NAO at the surface and influence climate in the Euro-Atlantic region for weeks to months (section 3.4).

This response of the polar stratosphere to El Niño has been successfully simulated in climate models and seasonal prediction models (Figure 10; Bell et al., 2009; Domeisen et al., 2015; Ineson & Scaife, 2009; García-Herrera et al., 2006; Manzini, 2009; Richter et al., 2015). Models with higher vertical resolution or a higher model top exhibit a more realistic surface response, pointing to the importance of correctly simulating stratospheric processes (Butler et al., 2016; Cagnazzo & Manzini, 2009; Hurwitz et al., 2014).

While the understanding of ENSO's influence on the NH polar stratosphere has greatly improved over the last decade, the large variability of the extratropical atmospheric circulation and the influence of other external factors have made it difficult to quantify the role of the ENSO-stratosphere pathway in winter. Additionally, there may be nonstationary and nonlinear influences on the remote ENSO-stratosphere pathway, as discussed below and in section 5. These limitations are demonstrated in Table 1 and Figure 11. Table 1 provides the mean strength of the polar vortex and the NAO for each individual ENSO winter, along with the ENSO amplitude and flavor. There is considerable variability between individual ENSO events (Deser et al., 2017), though in general the polar vortex is weaker in El Niño winters (see also Table 2). The winters with the strongest polar vortex are usually La Niña winters, though there are some La Niña winters, particularly in years with SSWs, where the vortex is anomalously weak (Iza et al., 2016). Indeed, the recent SSW event in 2018, one of the strongest on record, occurred during La Niña, as did the very strong SSW in 2009.

Figures 11a–11c consider the interannual variability of the extratropical atmosphere over the period 1958 to 2017 stratified by the Niño3.4 SST anomalies in each winter. A linear correlation analysis of the teleconnections indicates the difficulty in identifying relationships in the presence of large internal variability: Figure 11a indicates a rather linear relationship between tropical Pacific variability (characterized by Niño3.4 SST) and the impacts in the North Pacific, though Frauen et al. (2014) suggest this might be due to the superposition of the nonlinear impacts of EP and CP El Niño. Figure 11b compares the tropical Pacific directly with winds in

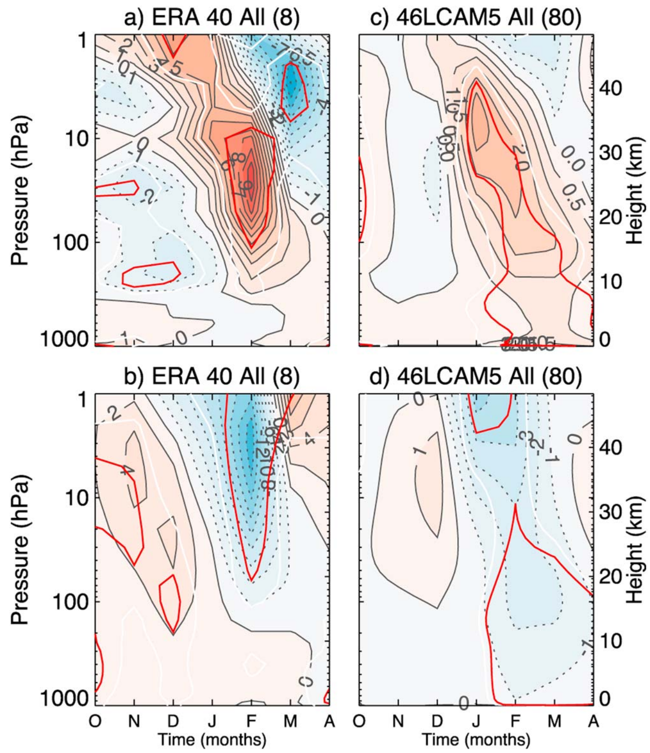

Figure 10. El Niño composites of zonal mean anomalies of temperature at 80° N (top) and zonal wind at 60° N (bottom) from October through April for ERA-40 reanalysis data (Uppala et al., 2006; left column) and the 10-member ensemble of the 46LCAM5 model (right column). The number of El Niño events included in each panel's average is depicted in parentheses in the title of each panel. The contour interval for temperature anomalies is 1.0 K for ERA-40 and 0.5 K for 46LCAM5. The contour interval for zonal wind anomalies is 2.0 ms^{-1} for ERA-40 and 1.0 ms^{-1} for 46LCAM5. The statistical significance of the signal based on a student *t* test at the 85% and 95% levels are depicted by white and red lines, respectively. Figure from Richter et al. (2015). Reprinted with permission.

the NH winter stratosphere; though the correlation is weak overall ($r = -0.21$), in general weaker winds occur during El Niño and stronger winds occur during La Niña. The impact of El Niño on the Northern Annular Mode is weak as well ($r = -0.19$; Figure 11c).

One explanation for why the correlations between the tropical Pacific and the polar stratosphere and extratropical surface are weak when all years are considered is that extremes of the polar vortex may not respond symmetrically to the two opposite phases of ENSO, at least in the short observational record (Table 2). In the observational reanalysis, SSW events occur during both El Niño and La Niña winters at a higher rate than during ENSO-neutral winters (Table 2; Butler & Polvani, 2011; Garfinkel, Butler, et al., 2012). This is relevant for seasonal forecasting because the occurrence of a single SSW during winter can dramatically alter seasonal climate over Europe and Siberia (Butler et al., 2014; Richter et al., 2015; Polvani et al., 2017). In fact, in the composite mean (Table 2), the sign of the stratospheric response depends on whether a SSW occurred or not, no matter the phase of ENSO; and the sign of the winter surface NAO/NAM in general matches the sign of the stratospheric zonal wind anomaly rather than ENSO phase (Figure 11d; Hansen et al., 2017). The same is true for predictability over the North Atlantic, which is increased for winters with SSW events; however, predictability is further increased for years with both a SSW and an El Niño event (Domeisen et al., 2015). The enhanced occurrence of SSWs in both El Niño and La Niña in the reanalysis might simply be related to sampling variability (i.e., the samples from reanalysis data are not sufficient to detect a signal over internal noise in the extremes of the zonal wind distribution and spurious trends can be observed in some data sets before the satellite era; Badin & Domeisen, 2014) or to the influence of other factors such as the QBO (see section 5.2; Richter et al., 2011; Taguchi, 2015). The observed La Niña-SSW relationship has also been shown to be sensitive to the classification of SSW events (K. Song & Son, 2018) and to the classification of ENSO events, which in turn depends on the SST data set (Polvani et al., 2017).

Table 2
Mean Values of the DJF Oceanic Niño Index (Column 3), the DJFM Zonal Wind Anomaly from 60° to 80° N Using JRA-55 (Kobayashi et al., 2015) Reanalysis (Column 4), and the Detrended and Standardized DJFM NAO Index, Multiplied by a Factor of 10 (Column 5)

	Number of years (SSW freq)	DJF ONI	DJFM Zonal wind Anomaly (m/s)	DJFM NAO index (×10)
All ENSO	40 (0.70)	0.2	−0.2	−0.5
All El Niño	22 (0.73)	1.2	−1.3	−1.2
All La Niña	18 (0.67)	−1.1	1.1	0.4
All Neutral	20 (0.45)	−0.2	2.0	1.0
El Niño + SSW	12	1.1	−4.8	−5.7
La Niña + SSW	10	−1.2	−4.1	−0.3
Neutral + SSW	9	−0.2	−2.8	−5.0
El Niño, no SSW	10	1.3	3.0	4.2
La Niña, no SSW	8	−1.0	7.7	1.4
Neutral, no SSW	11	−0.1	6.0	6.0
CP El Niño (N4 > N3)	8 (0.50)	0.8	0.3	-1.5
CP El Niño (Modoki)	9 (0.56)	1.2	-0.1	-1.3

Note. These composite values are based on the number of ENSO years in column 2, for different phases of ENSO and for years with and without at least one or more SSW, and for two different definitions of CP El Niño. Column 2 also includes the frequency of SSWs (the total number of SSWs divided by the number of years) for each category. ENSO = El Niño Southern Oscillation; NAO = North Atlantic Oscillation; CP = Central Pacific; SSW = Sudden Stratospheric Warming; DJFM = December-January-February-March.

Ensemble model simulations generally simulate relationships between ENSO and SSW occurrence that are more linear than those observed (Polvani et al., 2017). In general, the observed enhancement of SSWs during La Niña is not replicated in most CCMVal2 (SPARC, 2010) models (Garfinkel, Butler, et al., 2012) or high-top (Taylor et al., 2012) CMIP5 models (K. Song & Son, 2018), though SSW statistics based on absolute zonal wind values can be highly sensitive to model biases (Kim et al., 2017). It is possible that models fail to simulate a La Niña-SSW connection due to poor simulation of the diversity of ENSO and its interactions with the midlatitude circulation, resulting in tropospheric teleconnections that are too linear and too zonal compared to reality (Garfinkel, Butler, et al., 2012; L'Heureux et al., 2017). More research is needed to determine whether these processes are adequately represented.

While the observed relationship between La Niña and SSWs is not well-replicated in models, several modeling studies have simulated moderate increases in SSWs during El Niño winters (Bell et al., 2009; Domeisen et al., 2015,Garfinkel, Butler, et al., 2012; Y. Li & Lau, 2013; Polvani et al., 2017; Taguchi & Hartmann, 2006) in agreement with reanalysis. Dynamically, it makes sense that stronger wave driving in El Niño winters would at least occasionally lead to more extreme stratospheric disturbances. Mechanistically similar relationships between changes in the North Pacific and SSWs have been found independently on decadal time scales (Hurwitz et al., 2012; Woo et al., 2015) and intraseasonal time scales (Garfinkel, Feldstein, et al., 2012; Kang & Tziperman, 2017).

Nonetheless, because SSWs predominantly occur independently of ENSO (i.e., Polvani et al., 2017, estimate that more than 75% of SSWs occur irrespective of ENSO), several studies have stratified winters by whether an SSW has occurred or not, in order to better isolate the effects of El Niño on the vortex (Iza & Calvo, 2015; Polvani et al., 2017). This stratification, while not a perfect method of isolating the ENSO response, does allow for a clearer picture of the effects of ENSO on the mean state of the stratosphere, as the modulation of the wave driving induced by ENSO is relatively modest as compared to the wave driving observed during SSW events (Garfinkel, Butler, et al., 2012; Polvani & Waugh, 2004). Figure 11b separates the impact of ENSO on the NH vortex in winters with a SSW from those without a SSW and indicates that correlations are statistically significant only for winters without SSW events. Figure 11c in turn indicates that a significantly higher linear correlation can be obtained between the tropical Pacific and the NAM for winters with SSW events, consistent with the ability of models to predict the surface impact over Europe for winters that exhibit both an El Niño and a SSW event (Butler et al., 2016; Domeisen et al., 2015). That a surface connection is only evident for winters

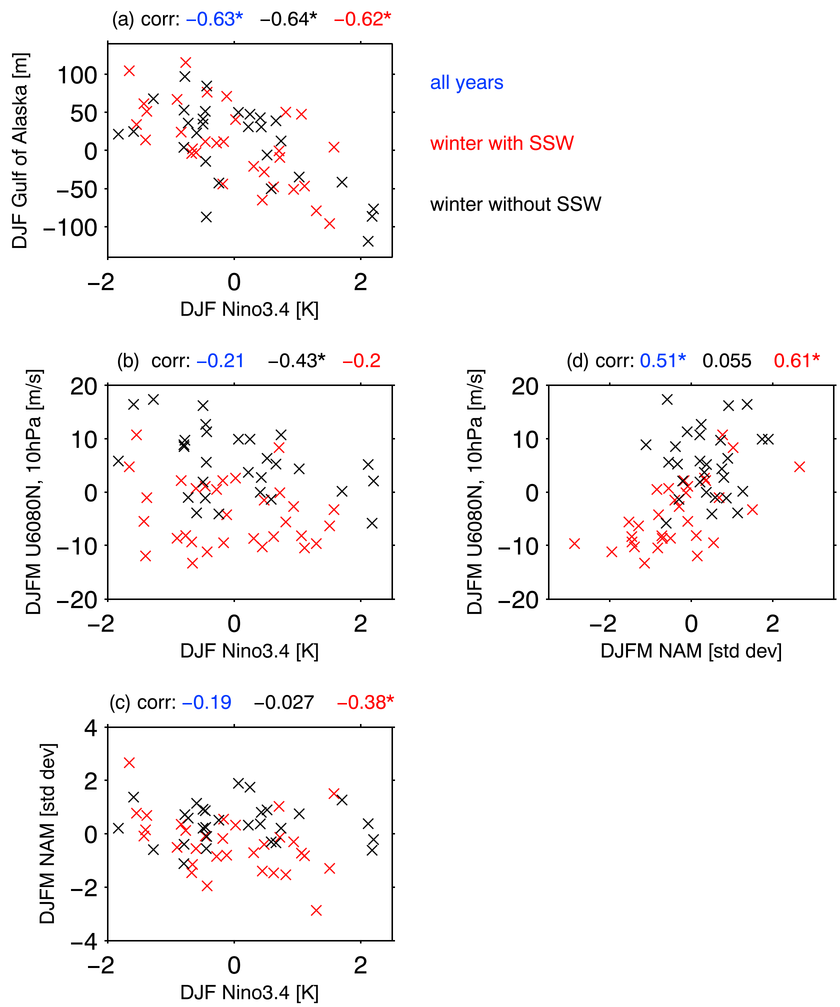

Figure 11. Interannual relationship between El Niño Southern Oscillation and variability in (a) sea level pressure over the Gulf of Alaska in DJF (as defined in Garfinkel, Hurwitz, et al., 2012); (b) zonal wind at 10 hPa, 60–80° N in DJFM; (c) the NAM (taken here as the Arctic Oscillation at 1,000 hPa from the Climate Prediction Center). (d) indicates the relationship between the NAM at 1,000 hPa and the zonal wind at 10 hPa, 60–80° N in DJFM. Stars denote correlations that are statistical significantly different from zero at the 95% level using a two-tailed Students *t* test. NAM = Northern Annular Mode; DJFM = December-January-February-March.

with a SSW is consistent with the observation that vortex variability robustly affects the NAM only when an SSW occurs (Figure 11d).

As discussed in section 2.1, ENSO exhibits significant diversity among events. It is not yet resolved to what extent ENSO diversity affects the diversity of stratospheric responses. A change in the location or strength of the tropical thermal forcing can considerably affect wave propagation into the stratosphere (Barriopedro & Calvo, 2014; Garfinkel, Butler, et al., 2012; Manzini et al., 2006). For example, thermal forcing from tropical convection over the Indian Ocean or the central Pacific can modulate the stratospheric response due to a different projection onto the extratropical climatological wave pattern (Fletcher & Cassou, 2015; Fletcher & Kushner, 2011; Rao & Ren, 2016a; X. Zhou et al., 2017). However, it has been difficult to discern a difference in the observed response of the NH polar stratospheric circulation to central versus eastern Pacific El Niño events due to the short observational record (Graf & Zanchettin, 2012), the influence of SSWs (Calvo et al., 2017; Iza & Calvo, 2015), and the sensitivity to the method for classifying central Pacific ENSO events (Garfinkel, Hurwitz, et al., 2012). The last two rows of Table 2 suggest differences in the stratospheric response depending on which CP definition is used (Garfinkel, Hurwitz, et al., 2012).

In the absence of a clear, methodologically robust difference in the observed stratospheric response to CP El Niño events as compared to EP El Niño events, many studies have forced models with CP SST patterns, but these studies have not reached a consensus. Some find little robust difference in the response between central Pacific and eastern Pacific El Niño events (Hegyi et al., 2014; Hurwitz et al., 2014), some argue that only eastern Pacific events lead to a weakening of the vortex (Calvo et al., 2017; F. Xie et al., 2012), while others argue that both lead to weakening of the polar vortex but with the vortex weakening in early winter stronger during EP El Niño (Garfinkel, Hurwitz, et al., 2012). It is not immediately clear why these different modeling studies reach different conclusions. Nevertheless, it is reasonable to expect a weaker response to CP El Niño events: The deepening of the Aleutian low occurs further south and is weaker in magnitude for CP El Niño (Di Lorenzo et al., 2010; Garfinkel, Hurwitz, et al., 2012; Garfinkel, Weinberger, et al., 2018; Sung et al., 2014; Yu & Kim, 2011), and hence, constructive interference with the climatological stationary waves is less efficient. This difference in the location of the deepened Aleutian low is likely due to differences in the zonal wavenumber and amplitude of the anomalies in tropical convection (Garfinkel, Weinberger, et al., 2018); however, internal variability can mask this effect (Deser et al., 2017). While there is a consensus that EP El Niño leads to a weakening of the NH polar vortex (Hurwitz et al., 2014), less agreement exists regarding the response to CP El Niño.

Another question that arises is whether strong El Niño events lead to a proportionately stronger Arctic stratospheric response as compared to more moderate El Niño events. While El Niño events tend to be associated with an increased frequency of SSW events in NH winter (Butler & Polvani, 2011), no SSW event was observed during the winters of the exceptionally strong El Niño events of 1982/1983, 1997/1998, and 2015/2016 (Palmeiro et al., 2017; Rao & Ren, 2017; the SST anomalies during the 2015/2016 El Niño are shown in Figure 4). On the other hand, the final warming in 2016 was one of the earliest on record, and only a very limited sample size of extreme ENSO events is available so far. Brönnimann et al. (2004) suggest that the strong extended El Niño period in 1940–1942 might have been associated with more SSW events. Rao and Ren (2016b) use observational data to argue that moderate El Niño and strong La Niña events are more efficient than strong El Niño and moderate La Niña events, respectively, in modulating the northern winter stratospheric variability in reanalysis data. However, the large internal variability in the stratosphere makes it difficult to detect such a nonlinearity (e.g., Figure 11 or Figure 11 of ; Rao & Ren, 2016b). Modeling studies disagree as to whether nonlinearities are present: while Rao and Ren (2016c) find evidence for nonlinearity, Richter et al. (2015) and X. Zhou et al. (2018) find that the modeled response to the strong El Niño events is proportionate to the response following moderate events.

Even if the Arctic stratospheric response to El Niño events is linear, nonlinearities may be important for the Atlantic sector response. Toniazzo and Scaife (2006) find that the North Atlantic sector response to moderate El Niño events resembles the negative phase of the NAO, while the response to stronger El Niño events bears the hallmark of a wave train launched from the tropical Atlantic, where very strong El Niño events can restructure the Walker circulation (and references therein ; Giannini et al., 2001). The model experiments of Bell et al. (2009) also suggest that very strong El Niño events impact the Atlantic sector via a tropospheric route while moderate El Niño events impact the Atlantic sector via the stratosphere.

Finally, there may be differences in ENSO tropospheric and stratospheric teleconnections between early and late winter. Moron and Gouirand (2003), King et al. (2018), and Ayarzagüena, Inseon, et al., (2018) suggest that there exists a significantly different teleconnection of ENSO to the Europe/North Atlantic sector in November-December versus January-March. Herceg-Bulić et al. (2017) find that the January-March ENSO response over Europe is maintained by both the stratosphere and extratropical Atlantic SSTs. Jiménez-Esteve and Domeisen (2018) find that the tropospheric teleconnection across North America that yields a negative NAO for El Niño is present during January-March, while the signal is limited to February for La Niña, modulated by decadal variability.

4.3. ENSO Teleconnections to the SH Stratosphere

Similar to the NH, the SH stratospheric response to El Niño events consists of a weakened stratospheric polar vortex associated with warmer temperatures in spring and summer (see Figures 1 and 5), along with enhanced poleward tropospheric planetary wave activity in the central South Pacific (Hurwitz, Song, et al.,2011). The stronger wave driving is associated with a poleward extension and strengthening of the South Pacific convergence zone, which can lead to an increased strength of the planetary wave flux into the stratosphere, for example, through linear interference (Smith & Kushner, 2012). This response is modulated by the QBO, with stronger planetary wave events during El Niño events with easterly QBO (Hurwitz, Song, et al., 2011). An easterly QBO is also suggested to have contributed to the only observed major SSW event that occurred in the SH in 2002 (Gray et al., 2005). T. Li et al. (2016) indeed suggest in a modeling study that the superposition of an

El Niño event and an easterly QBO may lead to an early breakdown of the SH polar vortex. In addition, Grassi et al. (2009) suggest a modulation by the Pacific Decadal Oscillation (PDO). Simpson et al. (2011) find the SH wave forcing of the BDC in response to ENSO is also modified in the subtropical lower stratosphere in austral summer/fall (i.e., December-March) due to changes in transient synoptic-scale wave drag in the subtropics.

As discussed in section 4.1.1, El Niño events are found to be associated with increased planetary wave driving in the NH, a strengthening of the Brewer-Dobson circulation and thus a cooling of the equatorial stratosphere. This response reduces the magnitude of the climatological SH stratosphere meridional temperature gradient and thus causes a slowdown and early breakdown of the polar vortex (T. Li et al., 2016), consistent with the warmer stratospheric temperatures in spring and summer found in Hurwitz, Song, et al. (2011). For the exceptional SH SSW event in 2002, tropical SSTs in the form of a moderate El Niño event are suggested to have played a role by modifying the stratospheric background state and thereby allowing for increased wave driving (Grassi et al., 2008).

As for the NH, the responses to different flavors of El Niño (EP vs. CP) and the different phases of ENSO (El Niño vs. La Niña) are more difficult to disentangle. However, several modeling studies agree on the response to CP El Niño: Hurwitz, Newman, et al. (2011) find that EP El Niño events do not exhibit a response in the SH stratosphere, while CP El Niño events are associated with the higher planetary wave activity described above. C. Yang et al. (2015) confirm the finding of weaker SH stratospheric winds for CP El Niño along with a strengthened BDC, with enhanced wave-1 propagation in August and enhanced wave-2 propagation in September. The spring warming for CP El Niño is also confirmed in the Whole Atmosphere Community Climate Model (*WACCM*; despite a SH spring cold bias; Zubiaurre & Calvo, 2012). Lin et al. (2012) find that both La Niña-like and CP El Niño-like tropical SST patterns lead to increased planetary wave activity in reanalysis data as well as chemistry-climate models. They also find zonal shifts in the stratospheric planetary wave patterns associated with the tropical forcing, though zonal mean changes are relatively muted. In contrast, Zubiaurre and Calvo (2012) find that the zonal mean response to La Niña is opposite to that of El Niño at least in a qualitative sense. Hence, the La Niña signal in the SH stratosphere is again—like in the NH—less clear than the El Niño signal.

In summary, sections 4.2 and 4.3 indicate that El Niño generally leads to a warmer polar stratosphere in both hemispheres. However, the two hemispheres differ in terms of which El Niño flavor (CP and EP) most strongly affects the stratosphere, with a stronger effect for CP El Niño in the SH and a stronger effect for EP El Niño in the NH. In both hemispheres, the effect of La Niña is less well understood.

4.4. Stratospheric Composition Changes

El Niño also impacts trace gas concentrations in the stratosphere, and here we focus on two of the species most important for the stratospheric radiative budget: water vapor and ozone.

In the tropical tropopause layer, water vapor increases in the region with warm anomalies and decreases in the region with cold anomalies (cf. Figures 5a and 5c). These local changes in tropical water vapor can exceed 25% below the cold point (Gettelman et al., 2001; Hatsushika & Yamazaki, 2003; Konopka et al., 2016). The pattern of warming and cooling in the tropical tropopause layer leads to an eastward shift in the location of dehydration near the top of the cold point during El Niño as compared to during La Niña (Fueglistaler & Haynes, 2005; Hatsushika & Yamazaki, 2003). The net effect of these temperature anomalies on water vapor above the tropical cold point is difficult to predict a priori, as these zonally asymmetric changes are superimposed on the larger-scale warming or cooling associated with changes in the BDC (Scaife et al., 2003).

The two largest El Niño events since satellites began measuring stratospheric water vapor (i.e., the events in 1997/1998 and 2015/2016) clearly preceded moistening of the tropical lower stratosphere (Avery et al., 2017; Fueglistaler & Haynes, 2005), though the observed impact of more moderate events is less clear (Garfinkel, Gordon, et al., 2018). The net effect of El Niño on water vapor at the cold point is the residual of the large temperature anomalies in the West Pacific and CP (Davis et al., 2013; Geller et al., 2002; Gettelman et al., 2001; Konopka et al., 2016), and zonally averaged changes in entry water vapor for the ENSO events considered by Gettelman et al. (2001) are 0.1 ppmv. In addition, the impact of El Niño on stratospheric water vapor changes between midwinter and spring: in spring, El Niño more robustly leads to enhanced water vapor and La Niña more robustly leads to dehydration (Calvo et al., 2010; Garfinkel et al., 2013; Konopka et al., 2016; Scaife et al., 2003). This seasonality may be related to changes in the climatological location of dehydration between winter and spring (Garfinkel et al., 2013). Finally, ENSO modulates the location of cirrus clouds in the tropical tropopause layer (Virts & Wallace, 2010) and may possibly modulate stratospheric water vapor via direct injection of cloud ice (Avery et al., 2017,Garfinkel, Gordon, et al., 2018).

El Niño leads to a reduction in ozone in the tropical lower stratosphere in satellite data (Hood et al., 2010; Oman et al., 2013), and a similar change is evident in models (Figure 7; Cagnazzo et al., 2009; Marsh & Garcia, 2007; Randel et al., 2009). Changes in ozone can reach 15% for a strong event (Hood et al., 2010; Randel et al., 2009). This reduction is due to the anomalously strong tropical upwelling during El Niño: ozone concentrations increase by more than an order of magnitude from the troposphere into the lower stratosphere and hence increased upwelling transports comparably ozone-poor air into the tropical lower stratosphere (Hood et al., 2010). In the tropical tropopause layer (and also in the troposphere), ozone concentrations increase over the Indian Ocean and Maritime continent but decrease over the central and eastern Pacific for El Niño (Figure 7). This zonally asymmetric pattern is due to anomalous upwelling within the Walker cell under El Niño: enhanced upwelling over the central Pacific advects ozone-poor air upward as compared to the Indian Ocean where upwelling is weakened.

ENSO also leads to regional anomalies in midlatitude ozone due to horizontal transport and mixing as well as vertical transport associated with its extratropical wave train (Hitchman & Rogal, 2010; Rieder et al., 2013; Yan et al., 2018; Zhang et al., 2015). Increases in ozone in the polar lower stratosphere during El Niño have also been noted in both observations and models (Cagnazzo et al., 2009; Randel & Cobb, 1994; Randel et al., 2009; Sassi et al., 2004). Finally, in the upper stratosphere, El Niño leads to an increase in ozone due to a combination of photochemical and dynamical processes (Hood et al., 2010).

5. Factors Influencing ENSO Teleconnections

Besides ENSO, a wide range of phenomena influence the stratosphere, and ENSO teleconnections to the stratosphere are superposed on, and potentially interact nonlinearly with, the influence from these other phenomena. This can lead to nonstationarity in the teleconnection. In addition, long-term trends such as climate change and ozone depletion/recovery can affect the background state along the pathway of the teleconnection and in the impact region, as well as ENSO itself. In this section we explore these interactions and discuss the potential for nonlinearities.

5.1. Interaction With Other Phenomena

Phenomena such as the QBO impact stratospheric variability, and hence, the effect of ENSO on the stratosphere can be influenced by the prior existence of a specific QBO phase. The temperature anomalies associated with the QBO in the tropical stratosphere modulate those associated with ENSO and therefore impact lower stratospheric water vapor concentrations (Geller et al., 2002; Liang et al., 2011; W. Yuan et al., 2014). In the extratropical NH, the observed response to El Niño as compared to La Niña is stronger during the westerly phase of the QBO (wQBO) than during the easterly phase (eQBO; Figure 12; Free & Seidel, 2009; Garfinkel & Hartmann, 2007). Two complementary explanations have been offered for this effect: Calvo et al. (2009) argue that this effect can be traced back to the modification of the stratospheric background state by the QBO (Anstey & Shepherd, 2014; Holton & Tan, 1980), while Garfinkel and Hartmann (2010) focus on the ability of the QBO to modify the tropospheric teleconnection of ENSO, as the North Pacific tropospheric teleconnections of ENSO are stronger during wQBO as well (Figure 12). In model experiments, there is no consensus as to whether the combined influence of the QBO and ENSO can simply be understood as a linear combination of the two forcings (Hansen et al., 2016) or whether nonlinearities indeed are pronounced (Calvo et al., 2009; Richter et al., 2015) with the wQBO phase delaying the onset of polar NH warming during El Niño until later in winter relative to eQBO. Finally, the NH polar response to CP El Niño may depend on the phase of the QBO (F. Xie et al., 2012).

In addition, El Niño events are characterized not only by warming in the central and east Pacific but also by warming in the Indian Ocean, which peaks in the spring following the peak SST anomalies in the Pacific Ocean (Murtugudde et al., 2000; Schott et al., 2009; Su et al., 2001; Webster et al., 1999). Teleconnections of El Niño can be driven both by the Indian Ocean warming and by any lingering SST anomalies in the Pacific. For example, impacts of El Niño in parts of East Asia are dominated by the Indian Ocean warming (S.-P. Xie et al., 2009), and variability in the strength of the connection between the South Asian monsoon and ENSO is governed by the Walker circulation (Webster & Yang, 1992) and Indian Ocean variability (Ummenhofer et al., 2011). The negative NAM response in the Arctic stratosphere to El Niño forcing is damped by the Indian Ocean warming (Fletcher & Cassou, 2015; Fletcher & Kushner, 2011; Rao & Ren, 2016a). The springtime tropical stratospheric

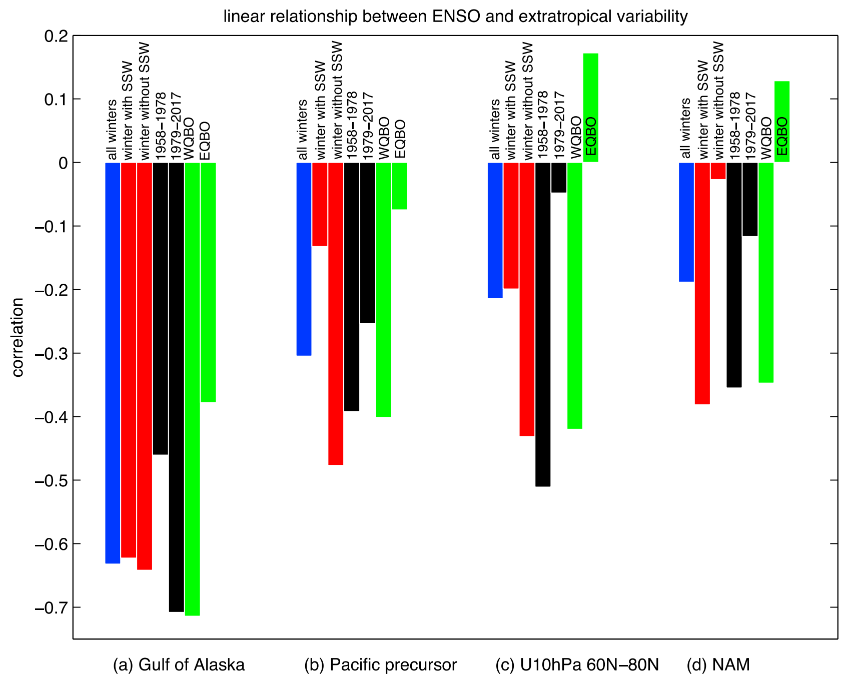

Figure 12. Correlation between Niño3.4 in DJF and (a) sea level pressure in the Gulf of Alaska (as defined in Garfinkel, Butler, et al., 2012) in DJF from the JRA-55 reanalysis (Kobayashi et al., 2015); (b) SSW precursor region (as defined in Garfinkel, Butler, et al., 2012) in DJF from the JRA-55 reanalysis; (c) zonally averaged zonal wind from 60° N to 80° N in DJFM at 10 hPa from JRA-55; (d) the NAM (taken here as the Arctic Oscillation at 1,000 hPa from Climate Prediction Center). Blue denotes the correlation upon considering all years from 1958 to 2017, red upon discriminating between winters with or without SSW, black upon separately considering the period before 1979 and after 1979, and green upon separating EQBO (zonal winds at 50 hPa in January more easterly than climatology) from WQBO (zonal winds at 50 hPa in January more westerly than climatology). DJF = December-January-February; SSW = Sudden Stratospheric Warming; WQBO = Westerly phase of the Quasi-Biennial Oscillation; EQBO = Easterly phase of the Quasi-Biennial Oscillation; NAM = Northern Annular Mode.

response to ENSO in both zonally averaged temperature and in water vapor is suggested to be governed by these Indian Ocean anomalies and not by lingering anomalies in the Pacific (Garfinkel, Gordon, et al., 2018).

5.2. Nonstationarity in ENSO Teleconnections

The observed connection between ENSO and the NH polar vortex appears to exhibit nonstationary behavior, that is, long-term changes in the teleconnection. The linear correlation between ENSO and the NH vortex has weakened in recent years (Figure 13c), and if one examines only the period since 1979, it is not statistically significant in the lower stratosphere (J. Hu et al., 2017). Consistent with this, decadal variability is also present in the Atlantic sector response to ENSO. When limiting the analysis period to 1979–2017, Figures 12d and 13d show a weakening correlation between ENSO and the NAM. This change is consistent with changes in the tropospheric anomalies during ENSO: the North Pacific low extended into Northeast Asia during the period before 1979 but was confined to the Northeast Pacific since 1979 (Figures 13a and 13b; J. Hu et al., 2017). A switch in correlation between the North Pacific and the North Atlantic can be observed at the same time (King et al., 2017). As discussed in Garfinkel et al. (2010), a low in the northeastern Pacific is less effective at weakening the vortex as compared to a low that extends into the northwestern Pacific. In addition, the North Pacific ridge associated with La Niña has shifted closer to North America since 1979 (Kumar et al., 2010; S. Yang et al., 2017) where it leads to an enhanced wave-2 signal (S. Yang et al., 2017). Decadal variability is also evident in the teleconnection of ENSO to the SH, albeit with a different timing with respect to the decadal variability in the NH (Evtushevsky et al., 2018).

It is not clear whether this decadal variability in the impact of ENSO teleconnections is forced or associated with internal variability of the atmosphere-ocean system. Decadal ocean variability in the Pacific sector may lead to decadal changes in the strength of ENSO teleconnections in the North Pacific (Gershunov & Barnett, 1998), while the North Pacific in addition exhibits multidecadal variability independent of the tropical Pacific (Latif, 2006). In turn, the North Pacific acts as a modulating factor on the decadal variability of ENSO in the

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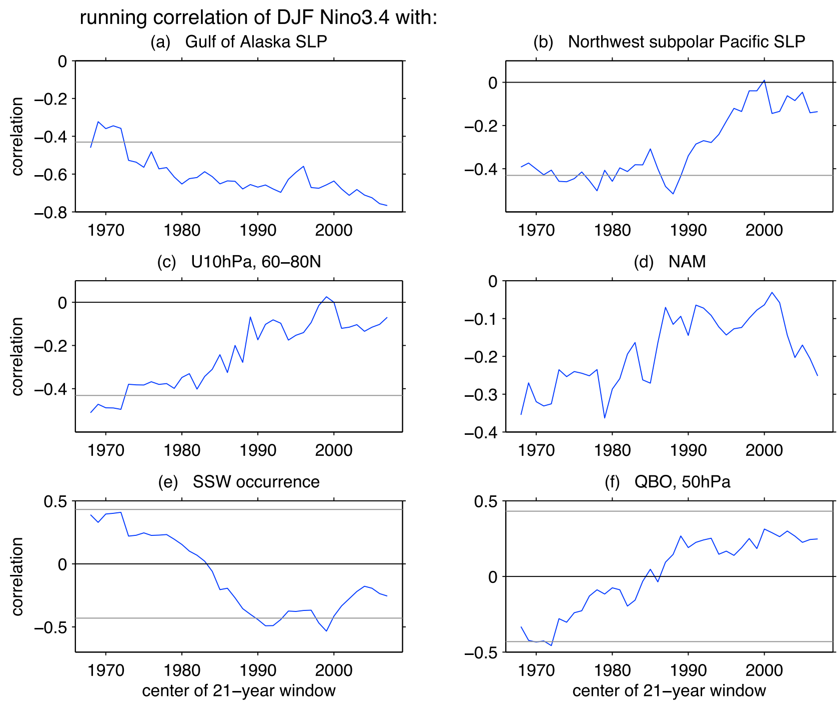

Figure 13. Running correlations over 21-year windows between Niño3.4 for DJF and extratropical variability in (a) SLP in the Gulf of Alaska region (as defined in Garfinkel, Hurwitz, et al., 2012) for DJF from the JRA-55 reanalysis (Kobayashi et al., 2015); (b) SSW precursor region (as defined in Garfinkel, Hurwitz, et al., 2012) in DJF from the JRA-55 reanalysis; (c) zonally averaged zonal wind from 60° N to 80° N in DJFM at 10 hPa from the JRA-55 reanalysis; (d) the NAM (taken here as the Arctic Oscillation at 1,000 hPa from Climate Prediction Center); (e) the occurrence of an SSW as in Table 1; (f) the QBO in January (zonal wind at 50 hPa from radiosondes, updated from ; Naujokat, 1986). Gray lines indicate statistical significance at the 95% level for a 21-year correlation. Compare (c) to Figure 11 in J. Hu et al. (2017). DJF = December-January-February; NAM = Northern Annular Mode; SSW = Sudden Stratospheric Warming; QBO = Quasi-Biennial Oscillation; SLP = sea level pressure.

tropical Pacific via atmospheric teleconnections (Barnett et al., 1999). Note, however, that the dominant mode in the Pacific sector was in the same phase from 1947 to 1977 and again from 1998 to 2014 (M. Newman et al., 2016). While the stratospheric response weakened after 1977, there is no indication of a recovery from 1998 to 2014, and hence, there is a mismatch between the phasing of North Pacific decadal variability and decadal variability in the ENSO-vortex relationship. Another possibility is that decadal variability in the Atlantic sector SSTs may modulate the Atlantic sector impacts of ENSO (Ayarzagüena, López-Parages, et al., 2018; López-Parages & Rodríguez-Fonseca, 2012; López-Parages et al., 2015; López-Parages, Rodríguez-Fonseca, Dommenget, et al., 2016; López-Parages, Rodríguez-Fonseca, Mohino, et al., 2016), though again the timing of decadal changes in the Atlantic are not consistent with the timing of decadal changes in the stratospheric response.

In addition to explanations involving the dominant modes of oceanic variability, atmospheric modulation of the teleconnection may also play a role for the nonstationarity of ENSO teleconnections to the stratosphere, possibly complementary to the oceanic modulations. El Niño (La Niña) was associated with more (fewer) SSW events in the first half of the period shown in Figure 13d, but not in the second part. This changed relationship might have impacted the relationship of ENSO with the NAM. The QBO may also have contributed to the modified ENSO teleconnections. eQBO occurred preferentially during El Niño and wQBO occurred preferentially during La Niña before around 1979 (i.e., a negative correlation in Figure 13e), but since then, eQBO has occurred preferentially with La Niña, and wQBO has occurred preferentially with El Niño (positive correlation in Figure 13e). In other words, as the modulation of the vortex by the QBO is of similar magnitude to that of ENSO (Garfinkel & Hartmann, 2007), their respective influences reinforced each other before around 1979 yet canceled each other after, leading to apparent decadal variability. Statistical techniques can be used to mitigate such a tendency (e.g., linear regression as in Figure 5), and once the linear QBO effect is removed in a statistical sense, ENSO and Arctic stratospheric temperatures are significantly correlated at least locally over the period

1979 to 2014 (Figure 5). Indeed, the regression of the Niño3.4 index with lower stratospheric Arctic temperatures when multiple linear regression is used (as in Figure 5) is approximately double of that when no effort is made to remove other processes that are known to affect Arctic stratospheric temperatures. Another possibility is that this apparent decadal variability is unforced and occurred by chance and simply reflects the noise inherent in the climate system. The shortness of the observational record does not allow us to discriminate among these possibilities, and model experiments are necessary to clarify their relative importance.

Nonstationarity has also been observed in SH ENSO teleconnections (Fogt & Bromwich, 2006; Yu et al., 2015), but in the SH, the observed relationship between ENSO and the SAM has strengthened after the early 1990s in contrast to the NH where it has weakened. It is hypothesized that this change in the SH is associated with the increase in CP El Niño events during the recent time period, which changed both the tropospheric and stratospheric pathways (see section 4.3).

5.3. Long-Term Trends and Climate Change

Climate change can impact ENSO teleconnections in several ways: (1) Changes in the nature of ENSO events in the tropics can lead to differences in the forcing of teleconnections in terms of their strength and origin. (2) The teleconnections may be modified even for an identical ENSO-related SST anomaly: an altered atmospheric or oceanic mean state either due to long-term changes in SSTs in both the tropics and midlatitudes or due to changes in the tropospheric or stratospheric circulation or composition can lead to differences in Rossby wave propagation. (3) The impact of the teleconnections may be different for changes in the background state of the impact region.

1. Several model studies predict that the nature of ENSO will be modified under a changing climate, with projections of increased amplitude and/or frequency of the events (e.g., Cai et al., 2014, 2015; Latif et al., 2015; Timmermann et al., 1999; Zheng et al., 2018), though not all studies agree on the sign and amplitude of a possible change in ENSO (Chen et al., 2017a; Collins et al., 2010; Hurwitz et al., 2013; Latif & Keenlyside, 2009; Stevenson, 2012). For a review of the changes in ENSO with climate change, see S. Yang et al. (2018). The presence of linear trends in SSTs complicates efforts to identify whether ENSO events have already intensified (L'Heureux et al., 2013). It is however clear that CP El Niño events have become more frequent over the past decades (Lee & McPhaden, 2010; McPhaden, 2012). This shift has been attributed to changes internal to the climate system related to the feedback processes between the ocean and the atmosphere (Guan & McPhaden, 2016; Lübbecke & McPhaden, 2014). Model projections have suggested that CP El Niño events will continue to become more prominent with respect to EP El Niño events in the future (Yeh et al., 2009). However, more recent model studies do not agree as to whether this shift will indeed persist in a warmer climate (Chen et al., 2017b; Taschetto et al., 2014). In addition, unforced coupled ocean-atmosphere model runs spontaneously generate multidecadal epochs dominated by CP or EP events (Capotondi et al., 2015). While it remains unclear if the distribution between EP and CP El Niño will change in the future, it is possible that if the nature of ENSO events indeed changes, then changes in the stratospheric response could be expected with climate change (Hurwitz et al., 2013).
2. Even if the character of ENSO events was to not appreciably change, trends in the tropospheric midlatitude mean state and the stratosphere may affect how teleconnections propagate to and impact the stratosphere. Changes are expected in the extratropical quasi-stationary sea-level pressure distribution—CMIP5 models simulate a deepening of the Aleutian low under climate change (Chang, 2014; Karpechko & Manzini, 2017), while other models predict an eastward shift of the ENSO impact in the North Pacific (Müller & Roeckner, 2008)—and hence, a given teleconnection will be superimposed on locally altered climatological stationary waves. This might lead to an altered wave propagation into the stratosphere.
3. On top of these changes, the stratosphere itself exhibits a long-term radiative cooling trend induced by the increase in greenhouse gases. These are however suggested to be offset or superseded in the NH polar regions by the long-term strengthening trend in the BDC (Butchart et al., 2006; Karpechko & Manzini, 2017; Manzini et al., 2014), leading to an overall warming of the Arctic stratosphere and cooling of the lower tropical stratosphere. In the SH, the effect of climate change may in addition be counteracted by the potential recovery of stratospheric ozone discussed in section 3.3. However, an examination of the role of ozone recovery and radiative changes in CO_2 on the stratospheric response to a given SST perturbation yielded minimal changes in either hemispheres—in contrast, it was found that the future response to ENSO is controlled by changes in SST patterns (Hurwitz et al., 2013).

Because changes in the character of ENSO events with climate change are uncertain, it is unclear how extratropical teleconnections will evolve, and there is considerable model spread in the response of ENSO teleconnections to climate change (Yeh et al., 2018). However, some features appear robust across model simulations. Specifically, warmer climatological SSTs lead to an eastward shift in tropical Pacific convection anomalies in response to ENSO (Cai et al., 2015; Z.-Q. Zhou et al., 2014), which in turn leads to weaker ENSO teleconnections in the Northwest Pacific and stronger teleconnections closer to North America (Kumar et al., 2010; Hurwitz et al., 2013; Z.-Q. Zhou et al., 2014). These tropospheric changes could lead to a weaker stratospheric response, as similar tropospheric changes have already been linked to a weakened stratospheric response over the past several decades (section 5.2; J. Hu et al., 2017; S. Yang et al., 2017). However, the specific mechanism of Z.-Q. Zhou et al. (2014) and Cai et al. (2015) that is expected to lead to an eastward shift in the tropical Pacific convective response to ENSO has not yet manifested, as the past decades have shown a weak cooling of SSTs in the eastern and central Pacific, contrary to future projections from CMIP5 models (Meng et al., 2011; Sohn et al., 2012).

5.4. Feedbacks and Impacts From the Stratosphere Onto ENSO

In addition to ENSO affecting the stratosphere, separate effects have been proposed as to how variability in the stratosphere, or processes that profoundly affect the stratosphere, can affect ENSO. While this is not the focus of this paper, we here briefly mention a number of these. It has been suggested that enhanced (reduced) Arctic ozone in March preferentially leads to a La Niña (El Niño) event, 20 months later (Garfinkel, 2017; F. Xie et al., 2016). Nowack et al. (2017) suggest based on a modeling study that neglecting possible future changes in the vertical structure of tropical ozone might be one of the reasons for the simulated higher frequency of extreme ENSO events. Enhanced solar variability may also affect ENSO, though the robustness of this connection is difficult to establish (Haam & Tung, 2012; Roy & Haigh, 2012). Explosive volcanic eruptions in the NH or the tropics lead to SST anomalies that resemble El Niño, and hence a volcanic eruption during an El Niño phase prolongs the duration of El Niño, whereas the same eruption during La Niña years tends to hasten its decline (F. Liu et al., 2017; Ohba et al., 2013; Pausata et al., 2015). That being said, coupled ocean-atmosphere models that are used to forecast and understand ENSO processes can generate realistic ENSO events without any of these stratosphere-related forcings, and some of these currently remain speculative, but an improvement of the model prediction from including these processes might be possible.

6. Summary and Outlook

In summary, the stratospheric circulation (reviewed in section 3) is profoundly affected by ENSO variability in the equatorial Pacific (reviewed in section 2.1). Figure 4 gives an overview of the impacts on the stratosphere (discussed in section 4) and the tropospheric teleconnections relevant to the stratospheric response (reviewed in section 2.2). El Niño modifies the tropical wave spectrum generated by tropical convection and hence leads to a faster downward propagation of the QBO and an increased influx of ozone-poor air in the regions of strong convection in the central and eastern Pacific (see section 4.1.2). El Niño also leads to increased wave forcing in the subtropics and midlatitudes, which is associated with a strengthening of the BDC (section 4.1.1). The change in midlatitude tropospheric wave driving comes about through Rossby wave trains originating in the tropical and subtropical Pacific that then project on and deepen the subpolar quasi-stationary pressure systems, leading to enhanced upward wave propagation. The strengthened BDC during El Niño events leads to a cooling of the lower tropical stratosphere and a warming of the polar regions of both hemispheres. The increased wave driving leads to a weakening of the polar vortices of both winter hemispheres and is associated with an increased frequency of SSW events in the NH during El Niño years (sections 4.2 and 4.3). While the response during La Niña years is generally opposite, it is less robust.

We would like to emphasize that the robustness of the connections between ENSO and the global stratosphere is not fully resolved. In particular, it is not resolved to what extent El Niño and La Niña teleconnections are opposite to one another, whether the response to a strong El Niño event is proportionately larger than the response to a moderate event, and to what extent it matters whether the peak SST anomalies for a given event are located in the central Pacific as opposed to the eastern Pacific (section 4). The involvement of other phenomena such as the QBO affecting both the teleconnections (section 5.1) and the tropical Pacific itself (section 5.4) further complicates the analysis. Decadal variability (section 5.2) and long-term trends (section 5.3) further modulate how and to what extent ENSO teleconnections impact the stratosphere.

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Hence, there is inherent difficulty in reaching robust conclusions as to the relative importance of ENSO for stratospheric variability: The large variability in the atmosphere makes it difficult to identify the relevant signals. The observational data sets are only available for limited time periods and do therefore not yet allow for a robust identification of possible nonlinearities in the remote connections. The use of models to improve the data availability in order to identify nonstationarity and nonlinearities does so far not provide a definitive answer — models often simulate ENSO events and teleconnections that are considerably more linear as compared to the available observational data (e.g., Garfinkel, Butler, et al., 2012). On the other hand, the limited duration of the observational data set limits our ability to robustly identify model biases (Deser et al., 2017).

Nevertheless, even with the current generation of models, ENSO can currently be used for probabilistic forecasts because the variability in the tropical Pacific is long-lived (Butler et al., 2016; Domeisen et al., 2015). However, the relatively short observational record, long-term trends, and nonstationary behavior limit the possibilities for training of the model and for cross validation of these forecasts.

In order to better understand the impact of ENSO on the stratosphere, progress therefore needs to be made both in the modeling of ENSO events themselves and in the understanding and representation of their teleconnections in models. It has been shown that progress can be made through a better understanding of ocean-atmosphere interaction and feedbacks (Bayr et al., 2017) in order to improve the simulation of ENSO diversity and variability in the tropical Pacific, which will ultimately lead to improvements in the understanding and prediction of ENSO itself. Phenomena with faster variability but in which the underlying physical processes are similar (e.g., regionally enhanced convection in the tropical Pacific) such as the MJO and tropical rainfall (which can also be associated with ENSO) can be used to consider possible nonlinearities of the stratospheric response to wave trains from the tropics in a far-larger sample than is currently possible with ENSO (e.g., Cassou, 2008; Garfinkel, Feldstein, et al., 2012; Kang & Tziperman, 2017; Scaife et al., 2017; Schwartz & Garfinkel, 2017). However, model biases in the mean state and the exact representation of tropical events render it difficult to compare models and observations for different types of tropical events for both ENSO and the MJO (Dawson et al., 2011; Henderson et al., 2017; Yoo et al., 2015). Hence, we should expect that an improved simulation of the mean state in the extratropics (Dawson et al., 2011; Henderson et al., 2017) and a better representation of tropical convection in models (Yoo et al., 2015) will improve teleconnection quality.

While the stratosphere cannot currently be deterministically predicted by more than about 1–2 weeks in advance (Gerber et al., 2009; Karpechko, 2018; Tripathi, Baldwin, et al., 2015), probabilistic prediction of SSW events is possible on longer time scales (Scaife et al., 2016). Predicting the likelihood of a stratospheric event — which ENSO is suggested to modulate — could significantly improve forecasts for regions affected by stratospheric impacts, which includes the extratropical regions of both hemispheres, as well as the tropics. In order to predict the state of the stratosphere, the type (or absence) of interaction between ENSO and other phenomena such as the QBO, the MJO, and solar variability also needs to be better understood. Feedback processes from the stratosphere and other regions onto the tropical Pacific and onto the ENSO teleconnections will also be an important factor. It will be interesting to explore to what extent ENSO prediction skill is sensitive, and potentially enhanced, if additional forcings and feedbacks from the tropical Atlantic (section 2.2.1, the North Pacific (section 5.2) and external sources (discussed in section 5.4) are represented.

In conclusion, significant progress has been made over the past 10 years in terms of understanding ENSO impacts onto the stratosphere. However, long-term trends and nonstationary behavior render the analysis of ENSO and its teleconnections to the stratosphere more difficult. Forecasting and understanding the connection between ENSO and the stratosphere therefore remains a challenge.

Glossary

Aleutian low Quasi-stationary low pressure system in the North Pacific near the Aleutian Islands.

Amundsen Sea low Quasi-stationary low pressure system in the South Pacific in the Amundsen Sea.

Arctic Oscillation The dominant mode of climate variability in the Northern Hemisphere extratropics, representing a seesaw in mass between the extratropics and the poles, here used synonymously with the Northern Annular Mode.

Atmospheric Kelvin wave Classically considered to be a nondispersive, equatorially confined wave with zero meridional velocity that propagates eastward at the same speed a gravity wave would propagate in the

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absence of rotation. See Garfinkel, Fouxon, et al. (2017) for a geophysical fluid dynamics definition of Kelvin waves.

Brewer-Dobson Circulation (BDC) The mean mass transport within the stratosphere, which consists of rising motion in the tropics and subsiding motion over the poles.

Canonical El Niño A "regular"/Eastern Pacific El Niño event.

Cold point The coldest region of the tropical tropopause where temperatures can drop below 190 K.

El Niño Modoki An El Niño event that peaks further west in the equatorial Pacific in comparison to the *canonical* El Niño.

ENSO flavor The type of ENSO event in terms of its peak location in the equatorial Pacific.

Final warming The final reversal of zonal mean zonal winds in the stratospheric polar vortex of both hemispheres at the end of each winter. The winds then remain easterly until the next fall.

Hadley cell A thermally driven and zonally symmetric circulation consisting of the equatorward movement of the trade winds between about latitude 30° and the equator in each hemisphere, ascent near the equator, poleward flow aloft, and, finally, descent peaking near latitude 30°.

Hypsometric equation By taking the vertical integral of the hydrostatic equation, it can be shown that vertical distance between two pressure levels is proportional to the mean temperature.

Madden-Julian Oscillation (MJO) The dominant organized fluctuation in tropical weather on weekly to monthly time scales. It can be characterized as an eastward moving "pulse" of cloud and rainfall near the equator that typically recurs every 30 to 60 days.

Montreal Protocol Officially the "Montreal Protocol on Substances that Deplete the Ozone Layer" is an international treaty agreed on in 1987 to protect the ozone layer by phasing out ozone-depleting substances

North Atlantic Oscillation (NAO) The Atlantic sector manifestation of the Northern Annular Mode; more specifically, an oscillation between the strength of the Icelandic low pressure and the Azores high pressure centers near the surface.

Ozone hole The result of human-caused large-scale extreme ozone depletion in the Southern Hemisphere in austral spring.

Quasi-Biennial Oscillation A quasi-periodic downward propagating oscillation between easterly and westerly winds in the tropical stratosphere with a period of 22 to 34 months.

Rossby wave train Alternating quasi-stationary high and low pressure anomaly patterns that can constitute teleconnections between the tropics and the extratropics.

Southern/Northern Annular Mode (SAM/NAM) The main modes of variability in the extratropical Southern/Northern Hemisphere, respectively.

Spring predictability barrier The inherent difficulty to make an accurate model prediction of ENSO before May of a given year.

Sudden stratospheric warming (SSW) Commonly defined as a reversal of the zonal mean zonal wind in the extratropical midstratosphere associated with a strong warming of the stratosphere over a few days to weeks.

Teleconnection A remote influence of global scale in the atmosphere-ocean system.

Trade winds Northeasterly (NH) or southeasterly (SH) surface winds over the tropics, given by the surface signature of the tropical Walker cells.

Walker cell/circulation Zonal overturning cells in the atmosphere above the tropical ocean basins with surface winds moving water and air westward and with eastward flow aloft, with upward motion in convective regions, and subsidence in the eastern part of the Ocean basin.

Warm pool The region in the West Pacific and Maritime Continent where sea surface temperatures approach and exceed 30 °C climatologically.

Wave-1, Wave-2 Zonal wave numbers 1 and 2

Acronyms

AO Arctic Oscillation

BDC Brewer-Dobson Circulation

CCMVal Chemistry-Climate Model Validation Activity

CFC Chlorofluorocarbons

CP Central Pacific (referring to an El Niño Modoki event)

CPC NOAA Climate Prediction Center

CMIP3/5 Coupled Model Intercomparison Project Phase 3/5
CTI Cold tongue index
DJF December-January-February
EN El Niño
ENSO El Niño Southern Oscillation
EP East Pacific (referring to an East Pacific, i.e., a canonical El Niño event)
eQBO Easterly phase of the Quasi-Biennial Oscillation
JFM January-February-March
LN La Niña
MEI Multivariate ENSO Index
MJO Madden-Julian Oscillation
NAM Northern Annular Mode
NAO North Atlantic Oscillation
NDJ November-December-January
NH Northern Hemisphere
NOAA National Oceanic and Atmospheric Administration
NPO North Pacific Oscillation
N3 Niño3 area in the equatorial Pacific
N4 Niño4 area in the equatorial Pacific
ONI Oceanic Niño Index
PDO Pacific Decadal Oscillation
PNA Pacific North America pattern
PSA Pacific South America pattern
PSC Polar Stratospheric Cloud
QBO Quasi-Biennial Oscillation
SAM Southern Annular Mode
SH Southern Hemisphere
SLP Sea level pressure
SOI Southern Oscillation Index
SON September-October-November
SST Sea surface temperature
SSW Sudden Stratospheric Warming
TNI Trans-Niño Index
WACCM Whole Atmosphere Community Climate Model
wQBO Westerly phase of the Quasi-Biennial Oscillation

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Erratum

In the originally published version of this article, Table 1 included two errors. In the first row, last column, −0.9 erroneously published as −40.9; and in the second row, last column, 0.4 incorrectly published as 0.41. Table 1 has since been corrected, and this version may be considered the authoritative version of record.