Central Pacific El Niño, the "subtropical bridge," and Eurasian climate

Hans-F. Graf[1] and Davide Zanchettin[2]

Received 6 July 2011; revised 2 November 2011; accepted 5 November 2011; published 7 January 2012.

[1] This study contributes to the discussion on possible effects of El Niño on North Atlantic/European regional climates. We use NCEP/NCAR reanalysis data to show how the two different types of El Niños (the central Pacific, or CP, and the east Pacific, or EP) result in remarkably different European winter temperature anomalies, specifically weak warming during EP and significant cooling during CP El Niños, the latter being associated with a negative phase of the winter North Atlantic Oscillation (NAO). Our results diverge from former suggestions addressing the weakened stratospheric polar vortex as the dominant factor contributing to the El Niño/NAO teleconnection. We propose a tropospheric bridge as the mechanism primarily responsible for the establishment of a negative NAO phase and of associated cold European winters. This mechanism includes the subtropical jet (STJ) waveguide being activated only during CP El Niños, when anomalous convective heating occurs near the edge of the Pacific warm pool. Under these conditions the STJ is enhanced by planetary wave flux divergence in the subtropical upper troposphere, providing favorable conditions for the propagation of a wave number 5 disturbance around the subtropical Northern Hemisphere. This wave contributes to weakening of the Azores High and, hence, to the negative NAO phase. As global warming scenarios project an increase in the frequency of CP El Niño events, the distinctive nature of this mechanism implies that the probability of cold European winters may increase as well in future decades.

Citation: Graf, H.-F., and D. Zanchettin (2012), Central Pacific El Niño, the "subtropical bridge," and Eurasian climate, *J. Geophys. Res.*, *117*, D01102, doi:10.1029/2011JD016493.

1. Introduction

[2] El Niño is known as the strongest quasi-oscillatory mode of variability in the coupled ocean–atmosphere system and has attracted immense attention, especially since the very strong event in 1982–1983 and its apparent worldwide effects on weather and climate. While early analysis [*Ropelewski and Halpert*, 1987; *Halpert and Ropelewski*, 1992] showed anomalies spreading nearly globally, Europe seemed to be unaffected by El Niños. This was challenged in several more recent publications (for a comprehensive review, see *Brönnimann* [2007]). *Merkel and Latif* [2002] discussed model results showing that during an El Niño event weather anomalies are prominent over Europe that are typically consistent with a negative phase of the North Atlantic Oscillation (NAO). This corresponds to the finding of a number of researchers [e.g., *Brönnimann*, 2007, and references therein; *Ineson and Scaife*, 2009] that during El Niño events the winter stratospheric polar vortex is weaker than normal, a situation that is indeed also favorable for the establishment of a negative NAO phase [e.g., *Perlwitz and Graf*, 1995]. So, it seems that the effects of El Niño on stratospheric circulation in boreal winter, caused by enhanced planetary wave generation in the tropics and subsequent vertical propagation into higher atmospheric layers, is responsible for the teleconnection between the tropical Pacific and the midlatitude North Atlantic.

[3] The numerical description of the "stratospheric bridge" teleconnection outlined above requires models with a well resolved stratosphere and a realistic representation of El Niño structure and variability. Comparing the performances of a climate model with and without a highly resolved stratosphere, *Cagnazzo and Manzini* [2009] found that both model configurations captured the observed weaker polar stratospheric vortex, high surface pressure over the Arctic, and low surface pressure over western and central Europe and the North Pacific. However, the model with low stratospheric resolution underestimated the amplitudes of these anomalies. Based on simulations with an intermediate climate model with high stratospheric resolution, *Bell et al.* [2009] suggested that indeed El Niño results in a weaker stratospheric polar vortex, but that there is an upper limit of El Niño intensity above which the stratospheric effects are less important and tropospheric mechanisms dominate. This is somewhat counterintuitive if only the stratospheric bridge were at work, since one would expect that a stronger El Niño, i.e., higher positive sea surface temperature anomalies

[1]Center for Atmospheric Science, University of Cambridge, Cambridge, UK.
[2]Max-Planck-Institute for Meteorology, Hamburg, Germany.

0148-0227/12/2011JD016493

(SSTAs), would lead to stronger tropical tropospheric heating and thus to more planetary wave activity disturbing the stratospheric circulation and, hence, affecting the NAO.

[4] Very recently, there have been numerous reports on a "new" type of El Niño, which is characterized by predominantly positive SSTAs in the tropical central Pacific rather than in the East Pacific. Different names were given to this phenomenon: "central Pacific El Niño" [*Kao and Yu*, 2009; *Yeh et al.*, 2009], "Warm Pool El Niño" [*Kug et al.*, 2009], "Dateline El Niño" [*Larkin and Harrison*, 2005], and "El Niño Modoki" [*Ashok et al.*, 2007; *Weng et al.*, 2007] as well as, based on a more process-oriented study of El Niños in coupled ocean–atmosphere circulation models, S(urface)- and T(hermocline)- modes [*Guilyardi*, 2006]. Central Pacific (CP) and East Pacific (EP) events were described already by *Graf* [1986] (available at http://www.geog.cam.ac.uk/people/graf/graf11.pdf) as events driven solely by local surface wind anomalies not including the thermocline and events following the delayed oscillator theory including deepening of the thermocline by propagating Kelvin waves, respectively.

[5] *Di Lorenzo et al.* [2010] report that the North Pacific Gyre Oscillation (NPGO), which is defined as the second mode of SSTAs variability in the Pacific basin, very closely matches the CP El Niño SSTA pattern. They conclude that the frequency of CP El Niños drives the decadal variability of NPGO via the strength of the North Pacific Subtropical High. This is consistent with indications from *Brönnimann et al.* [2007] and *Zanchettin et al.* [2008], who noted that El Niño impacts most significantly on European wintertime rainfalls during positive (i.e., warm) phases of the Pacific Decadal Oscillation (PDO), which describes the leading mode of SSTA variability in the extratropical North Pacific. Also, indications exist from cluster analysis of observational data [*Gouirand and Moron*, 2003; *Brönnimann et al.*, 2007] that besides an El Niño signal that yields a cooling over northern Europe and a negative NAO phase, another, opposite pattern can develop. Hence, there are hints in scientific literature pointing toward different sensitivities of the climate system for different distributions of tropical SSTAs.

[6] With this study we shall outline the effects the two different types of El Niño have on weather and climate over Europe and worldwide. We will show that the "stratospheric bridge" is not the only mechanism connecting SSTAs in the tropical Pacific with the NAO. We provide evidence for a tropospheric mechanism that links the equatorial Pacific SSTAs with the atmospheric circulation over Europe and the North Atlantic quite efficiently in the case of CP El Niños, and much less efficiently in the case of EP El Niños. These two mechanisms lead to winter climate anomalies over central Europe that are opposite in sign for the two El Niño types, so that if all El Niños are considered regardless of their type, statistically less significant effects emerge.

2. Data and Methods

[7] This study is based on NCEP/NCAR reanalysis data [*Kalnay et al.*, 1996] covering the period 1948–2001 to 2010–2011 that were provided by the NOAA/OAR/ESRL PSD, Boulder, Colorado, USA. CP and EP El Niños are defined based on differences in their temporal development, underlying physical mechanisms, and overall patterns of SSTAs. We do not consider La Niña events since we concentrate on different atmospheric responses to the two types of El Niño. We focus our analysis on the Northern Hemisphere, thus avoiding problems with the reliability of this data set on the Southern Hemisphere [*Trenberth and Guillemot*, 1998]. For SSTs we also use the Kaplan SST data set [*Kaplan et al.*, 1998]. Specifically, we first define CP and EP El Niños based on the combined regression-EOF (empirical orthogonal function) analysis as proposed by *Kao and Yu* [2009]. They derived typical patterns and respective principal component time series for EP El Niño types by first subtracting the SSTAs regressed on the Niño4 index from the total SSTA field, and then performed an EOF analysis. Similarly, for CP El Niño types they first subtracted the SSTA regressed on the Niño 1.2 index. Following this analysis EP El Niños start in 1951, 1957, 1965, 1972, 1976, 1982, 1997, and 2008; CP El Niños start in 1963, 1968, 1977, 1986, over extended periods 1991–1995 (stronger events in 1991 and 1994), 2002–2005 (stronger event in 2002), and 2009. Different proposed indices for CP El Niños have shown slightly different results [*Ren and Jin*, 2011]. We therefore confirmed the selection above by visual inspection of the geographical position of the SSTA patterns. EP El Niños start in the East Pacific, have their maximum SSTA in the region of the climatological equatorial Pacific cold tongue, and do not extend to the west of the date line. CP El Niños start and have their maximum SSTA in the central Pacific straddling the date line. Our selection of the El Niño type years corresponds with the analysis by *Kug et al.* [2009], who only considered years after 1970, for the overlapping years. CP El Niños generally start during boreal summer and terminate in the following spring or early summer. Events further considered in the analysis are EP El Niños 1951, 1957, 1965, 1972, 1076, 1997, and 2008 and CP El Niños 1968, 1977, 1986, 1994, 2002, and 2009. We excluded from our analysis the El Niño years that coincided with the strong volcanic eruptions of Agung (1963), El Chichon (1982), and Pinatubo (1991), since these added additional forcing, which cannot easily be subtracted from the observations. Both El Niño types reach their mature phase during northern winter. Until the 1970s the sequence of EP El Niño started regularly after a period of cool CP SSTAs followed by warming there during decay of the EP El Niño. This behavior changed for the two big (mixed) EP El Niños in 1982–1983 and 1997–1998, which both followed extended periods of warm central Pacific and terminated by its cooling.

[8] According to our selection, CP El Niños are, hence, by no means "new," but happen to occur more often recently. Actually, an analysis of the Kaplan SST data set [*Kaplan et al.*, 1998] shows that CP El Niños were rare but not absent in the first half of the 20th century. They are diagnosed for the events of 1914–1915, 1930–1931, and 1940–1941–1942. This apparent clustering of CP and EP El Niños suggests that the occurrence of different El Niño types could be related to multidecadal phases of the PDO in a similar way to that previously demonstrated for the magnitude and the frequency of ENSO (El Niño–Southern Oscillation) events [*Kiem et al.*, 2003; *Kiem and Franks*, 2004]. Because of global warming signals and PDO regime shifts potentially biasing our results due to the increased frequency of CP El Niños in the most recent decades, we used linearly detrended data throughout the analysis. We acknowledge that by our approach, potentially relevant features may be

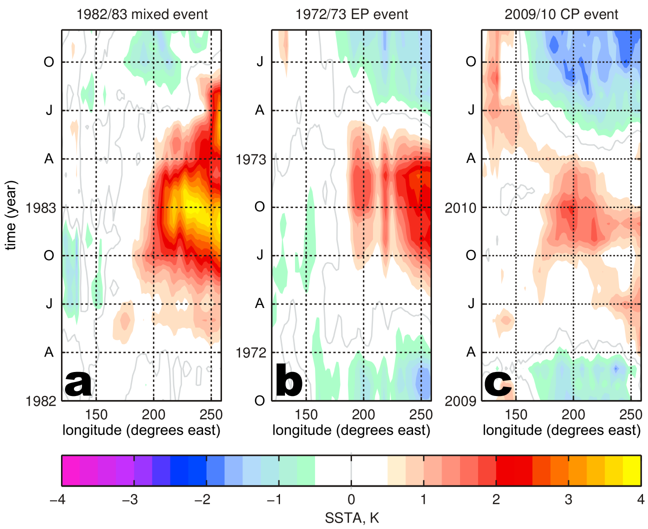

Figure 1. Hovmöller diagrams illustrating the typical evolution of tropical Pacific deseasoned monthly SSTAs (NCAR reanalysis skin temperature) for different El Niño types: (a) mixed El Niño, (b) east Pacific El Niño, and (c) central Pacific El Niño.

partly smoothed out, particularly those related to ENSO-PDO interactions. However, we also looked at nondetrended results and did not detect relevant differences that would change our conclusions.

[9] We acknowledge also that the number of events that can be studied is quite limited, but currently there is no way of extending the data set. In order to account for such small samples, we use a detection and attribution approach [e.g., *Zanchettin et al.*, 2012] and concentrate our discussion on features passing the attribution test. A nonparametric estimate of the likelihood for the attribution of the obtained anomalies to the selected El Niños is determined by analyzing the results yielded by a large number (here 500) of random sequences of winter atmospheric states over the investigated period. The anomaly distribution yielded by the random sequences is used as a basis to evaluate the confidence level for the attribution of a signal to a chance occurrence. More specifically, signals are considered to be statistically significant if they exceed the 5th–95th percentiles evaluated for the random anomaly distribution. The results obtained by following this approach are very similar to those obtained by using the Mann–Whitney U test [*Steel and Torrie*, 1960]. In addition to these tests we also performed "field significance tests" by using a Monte Carlo approach similar to that employed by *Eshel et al.* [2000]. Here, local significances assessed by the Mann–Whitney U test for a set of 500 random sequences of winter atmospheric states is used to determine the likelihood for locally significant signals covering a given areal fraction of the domain.

[10] Figure 1 shows three typical tropical Pacific SSTA patterns as they evolve for three different types of El Niño. Figure 1a depicts the situation as was observed around 1982–1983, which we refer to as a mixed event. This El Niño started in May 1982 clearly with positive SSTAs across the whole equatorial Pacific basin. These intensified east of 200°E during winter 1982–1983 and, after some weakening during spring 1983, a second phase of El Niño developed starting from the South American coast. This second event extended westward, but no further than 240°E, peaking in summer 1983. Figure 1b depicts the evolution of the El Niño event around 1971–1972, which exemplifies the dynamics of an EP El Niño. It very clearly was initiated at the South American coast, spreading then into the central Pacific, but not extending westward across the date line. Figure 1c depicts the pattern of the most recent El Niño, which started in June 2009 and lasted until May 2010, exemplifying the dynamics of a CP El Niño. Positive SSTAs are in this case much smaller in amplitude and are concentrated in the central Pacific, while there is a tendency toward negative or near normal SSTAs in the far eastern Pacific after November. In the tropics, deep convection and associated rainfall is observed over SSTs exceeding approximately 26°–28°C [*Zhang*, 1993]. During CP El Niños, positive SSTAs are added to the climatologically warm SSTs of the central and western equatorial Pacific and have therefore a comparatively bigger effect on convective activity and rainfall than the SSTAs during EP El Niños, which are added to the cooler temperatures along the equatorial Pacific cold tongue.

[11] In the following, the winter (DJFM) anomalies of a selection of diagnostic climate variables during the two different El Niño types, i.e., only CP El Niños and only EP El Niños, as well as during all El Niños are investigated. Since the "all El Niños" analysis essentially represents a weighted result of CP and EP effects, we refrain from showing the associated results. Nonetheless, we discuss them when deemed necessary and particularly in terms of statistical significance of the signal. Anomalies during El Niños are defined as departures from the state of an undisturbed climate, here associated to those winters (including La Niña winters) when there was no El Niño of any type and no volcanic disturbance.

[12] Eliassen-Palm flux diagnostics are evaluated based on monthly averaged NCAR reanalysis data following the transformed Eulerian-mean set of equations by *Andrews et al.* [1987, p. 128].

3. Results

[13] Figure 2 shows the mean winter near surface air temperature (SAT) anomalies found for the mature phases of EP and CP El Niños. In the tropical Pacific, SAT anomalies reflect the SSTA pattern typically associated to the two El Niño types. Significantly positive SAT anomalies are found stretching out into the climatological Pacific cold tongue region for EP El Niños, reaching from the South American coast to about 170°W. Significantly positive SAT anomalies are instead concentrated in the central Pacific region for the CP El Niños, showing a maximum around 160°W and crossing the date line to the west. Also, for CP El Niños there are no positive SAT anomalies south of 10°S in the eastern Pacific.

[14] Outside the tropics great SAT differences are found in mid and high latitudes over Eurasia mainly for CP El Niños. During EP El Niños, central and eastern Europe show positive SAT anomalies and a larger area of locally significantly colder temperatures is found in Mongolia, northern China, and East Siberia. However, a field significance test for

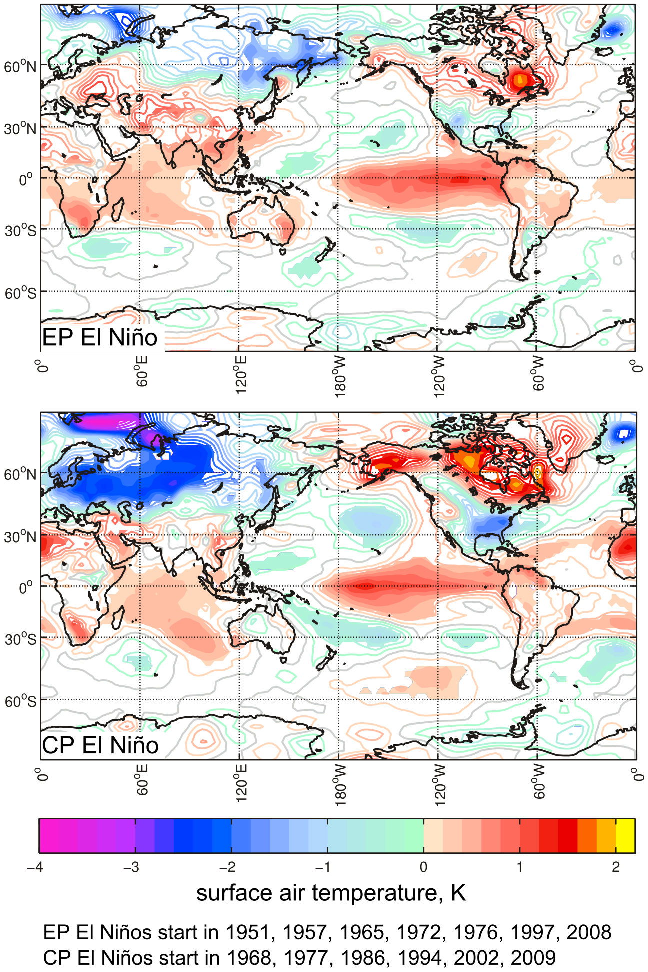

Figure 2. Global pattern of surface air temperature anomalies during the mature phases of (top) EP and (bottom) CP El Niños. Color shaded areas indicate statistically significant anomalies.

midlatitude (30°N to 60°N) SAT anomalies indicates that the null hypothesis of no EP El Niños signals on the field can only be rejected with confidence below the 50% level (not shown). The midlatitude field response to EP El Niños is thus very likely originated by chance. In contrast, for CP El Niños a field significance test for midlatitudes indicates that the null hypothesis can be rejected with confidence very close to the 95% level (not shown); hence the associated results are quite unlikely originated by chance. During CP El Niños, central and eastern Europe is up to 2 K colder than normal and these anomalies are statistically locally significant as are the cold anomalies over western Russia. For western Russia, only the stretch from Spitsbergen to Northwest Siberia is significant, according to the MWU test (not shown). East Siberia is much less affected during CP El Niños than during EP El Niños. A closer inspection of individual El Niño winters reveals that the cooling of Europe/Siberia occurred in all CP El Niño years except 1994–1995, when Europe experienced a mild winter (not shown).

[15] During EP El Niño winters, positive SAT anomalies emerge as well over North America, though they are locally significant only over Newfoundland, and the field significance test for midlatitude temperature anomalies reveals that these are probably by chance. During CP El Niño winters, much warmer anomalies are found over northeastern North America. They extend well into the Canadian Arctic. Both El Niño types are generally associated with cooling of the southern United States, but the strongest anomalies are found in the southeast (southwest and only locally significant) during CP (EP) El Niño, with stronger cooling during CP El Niños. Also, during CP El Niño winters, stronger and significantly warmer SAT anomalies than during EP El Niños extend farther east and also south of 30°S over the Indian Ocean. Southeast Asia is significantly warmer during boreal winter under EP El Niño conditions. Australian temperatures are significantly affected only by EP El Niños, specifically with warm anomalies over the eastern coast.

[16] All El Niño winters except the winter 1994–1995 (not shown) are characterized by a weak stratospheric polar vortex, very consistent with previous studies [e.g., *Brönnimann*, 2007; *Ineson and Scaife*, 2009]. As shown in Figure 3 (left), both EP and CP El Niños are associated with a weaker than normal stratospheric polar vortex at the 50 hPa level. The peculiar conditions during winter 1994–1995 reduce the area of statistically significant weakening of the stratospheric polar vortex in the CP El Niño ensemble. The average weakening of the vortex is much stronger and statistically significant over the whole polar cap during EP El Niños. In the midlatitudes we find statistically significant negative geopotential height anomalies over the North Pacific and the North Atlantic, including central and west Europe during CP El Niño winters. As shown by 1000 hPa geopotential height anomalies (Figure 3, right), the effect of El Niño on the lower-tropospheric circulation over the North Atlantic is only significant for CP El Niños, when a negative NAO-like pattern emerges, which is dominated by a negative anomaly of the Azores height. Both El Niño type winters are characterized by positive pressure anomalies in the western hemisphere Arctic. None of the El Niño composites provides a clear indication for a weakening of the Icelandic Low. For CP El Niños also a very strong and significant negative near-surface pressure anomaly is found over the North Pacific between 30°N and 60°N, while pressure anomalies found there in EP El Niño winters are only marginal.

[17] These results obtained by investigating the two different types of El Niño individually challenge former suggestions that the "stratospheric bridge" is, via vertical propagation/reflection of midlatitude and high-latitude planetary waves as they interact with the lower stratosphere westerlies, the dominant mechanism for the El Niño/NAO teleconnection. The very weak polar stratospheric vortex during EP El Niños does not reflect upon the tropospheric circulation, while the less disturbed vortex during CP El Niños clearly coincides with a negative NAO-like pattern. Thus, our current results suggest that there must be another, tropospheric mechanism at work. We suggest a "subtropical bridge" based on the waveguide provided by the subtropical jet (STJ).

[18] Poleward propagating Rossby waves generated by anomalous atmospheric heating in equatorial latitudes may be trapped by the STJ and propagate eastward within the subtropical waveguide [e.g., *Ambrizzi and Hoskins*, 1997; *Branstator*, 2002]. The stationary wave number in β-plane approximation $\boldsymbol{K_s} = (\beta^* / \mathrm{u})^{1/2}$, with $\beta^* = \beta - \delta^2 \mathrm{u} / \delta y^2$ the meridional gradient of absolute vorticity, exists for u > 0 and

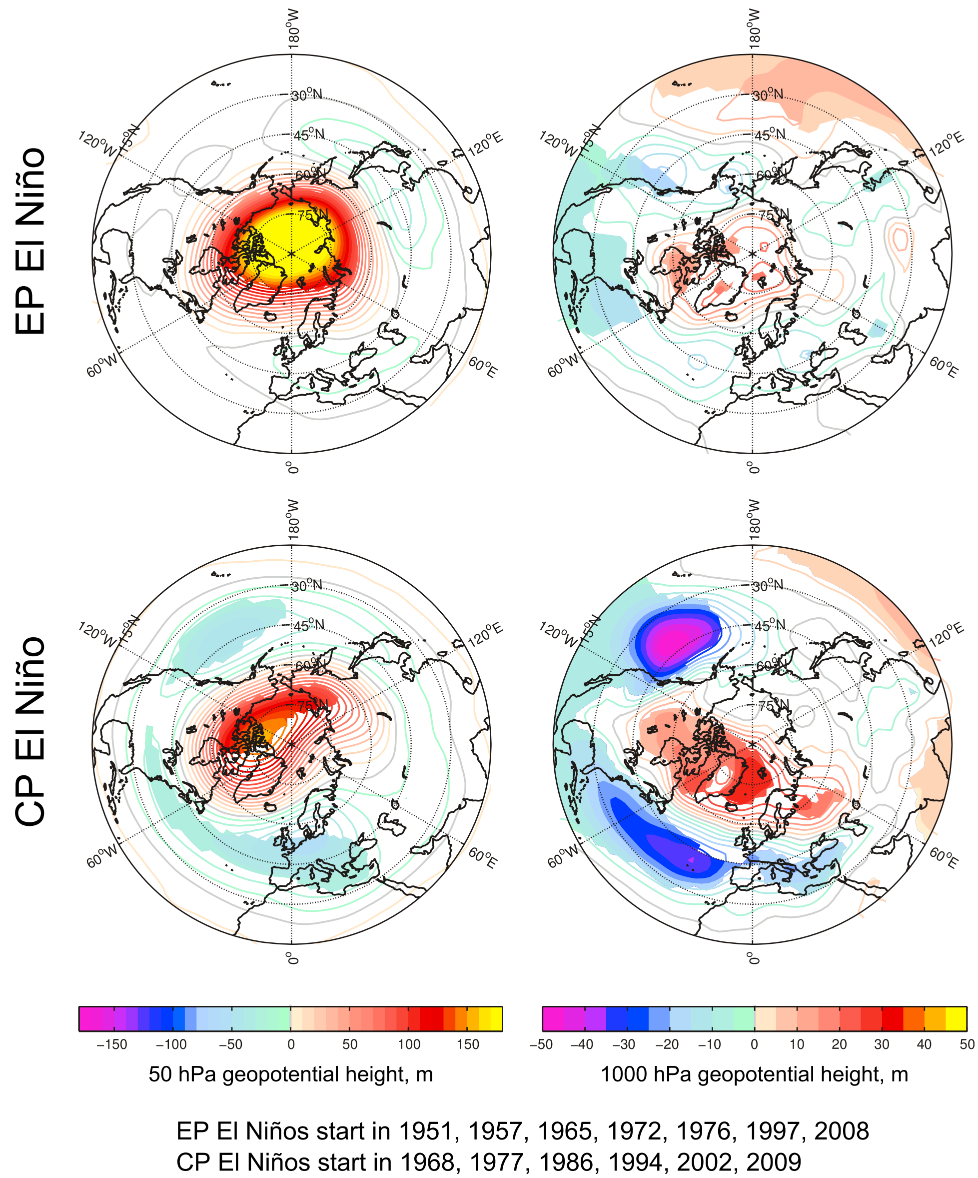

Figure 3. (left) Northern Hemisphere 50 hPa geopotential anomaly of EP (top) and CP (bottom) El Niño winters. (right) Same as Figure 3 (left) but for 1000 hPa geopotential height anomalies. Representation as in Figure 2.

$\beta^* > 0$. In case of a strong zonal wind shear (as is the case for jet streams), β^* dominates over the zonal wind effect. $\boldsymbol{K_s}$ plays the role of a refractive index for Rossby wave propagation. In strong westerly jets, $\boldsymbol{K_s}$ has a local maximum in the core of the jet. The region with maximum $\boldsymbol{K_s}$ acts as a zonally oriented waveguide.

[19] The effects of the STJ on planetary wave propagation and the resulting atmospheric anomalies depend on three elements: (1) the existence of the jet and its strength, position and extension since only a strong zonal jet can act as a zonal waveguide, (2) the relative geographical position of the meridional path of the Rossby waves to the STJ, which is determined by the longitude of the origin of the Rossby waves, since waves can only be trapped when their path crosses the STJ, and (3) the strength of Rossby waves generated at the equator by enhanced precipitation and related latent heat release since this determines the amplitude of the atmospheric disturbance.

[20] For Rossby waves associated with El Niños, obviously the East Asian STJ is relevant. It is strongest in winter (December to February/March) and, on average, extends with its core to just east of the date line (Figure 4, top). There are very clear differences between the mean anomaly patterns of 300 hPa zonal mean wind between EP (Figure 4, middle) and CP (Figure 4, bottom) El Niño winters. We find for all CP El Niños, except for 1994–1995 (not shown), an extension of positive zonal wind anomalies at the 300 hPa level in the subtropics from East Asia toward the subtropical Atlantic along the ~30°N latitude band. This is not the case for EP El Niños, when the westerly winds in the Northern Hemisphere subtropical upper troposphere are much weaker at the eastern tip of the East Asian STJ over the central Pacific.

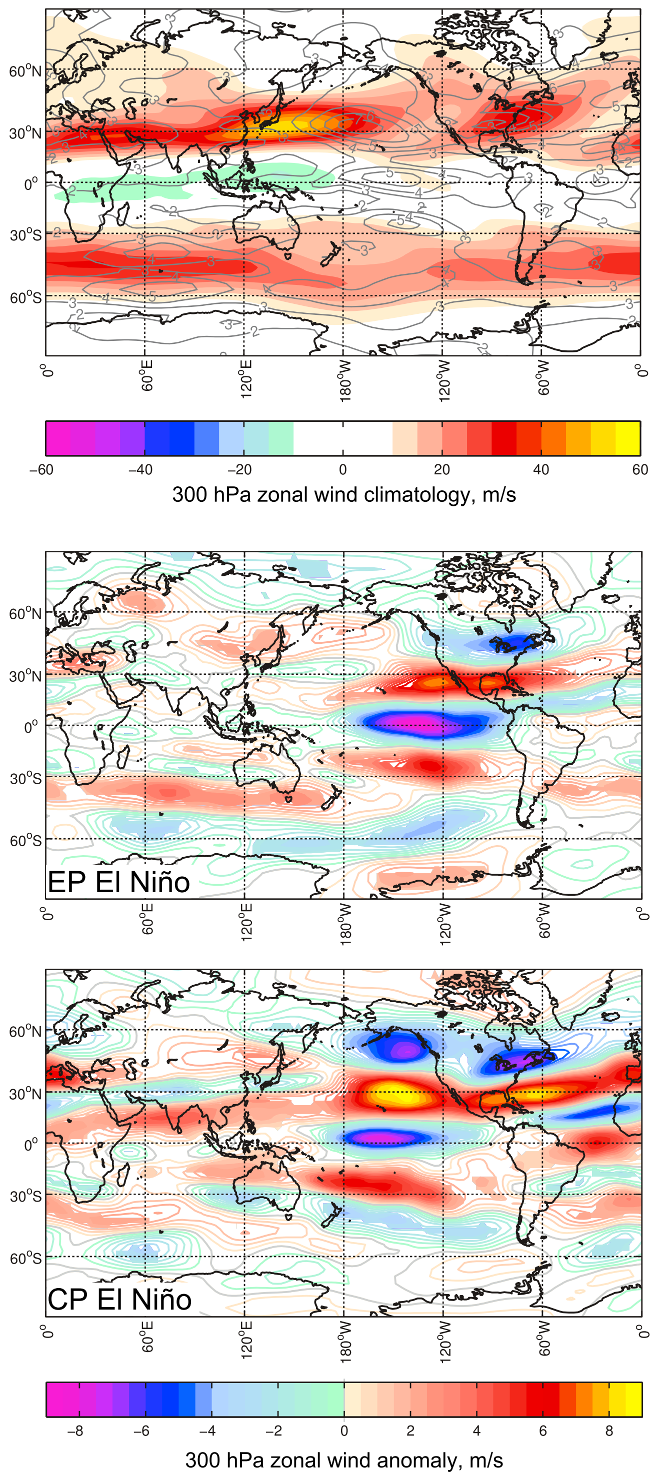

Figure 4. Winter zonal wind at 300 hPa: (top) climatology, and anomalies for (middle) CP and (bottom) EP El Niño winters. Figure 4 (top) shows long-term mean (shading) and standard deviation (gray line contours). Winters influenced by volcanic eruptions are excluded from the analysis. Figure 4 (middle) and Figure 4 (bottom): representation as in Figure 2.

[21] Eliassen-Palm (E-P) flux analysis (Figure 5) clearly indicates the differences between the two El Niño types in winter in the zonal mean zonal wind-forcing due to planetary waves. For CP El Niños the divergence of the meridional component of the E-P flux leads to a strengthening of the zonal wind in the upper troposphere near 30°N; i.e., the STJ is strongly reinforced, while upper tropospheric jets at higher northern latitudes are weakened. The patterns for EP El Niños are similar, but the effects are much weaker and are barely significant. Hence, during CP El Niños the STJ is elongated and strengthened, providing a potential zonal waveguide for Rossby waves originating in the equatorial Pacific.

[22] The possible key to the "subtropical bridge" is found in the zonal position and strength of precipitation anomalies associated with the two El Niño types (Figure 6). While for EP El Niños large positive precipitation anomalies in the equatorial Pacific are located east of the date line, they are stronger and more concentrated to the central Pacific,

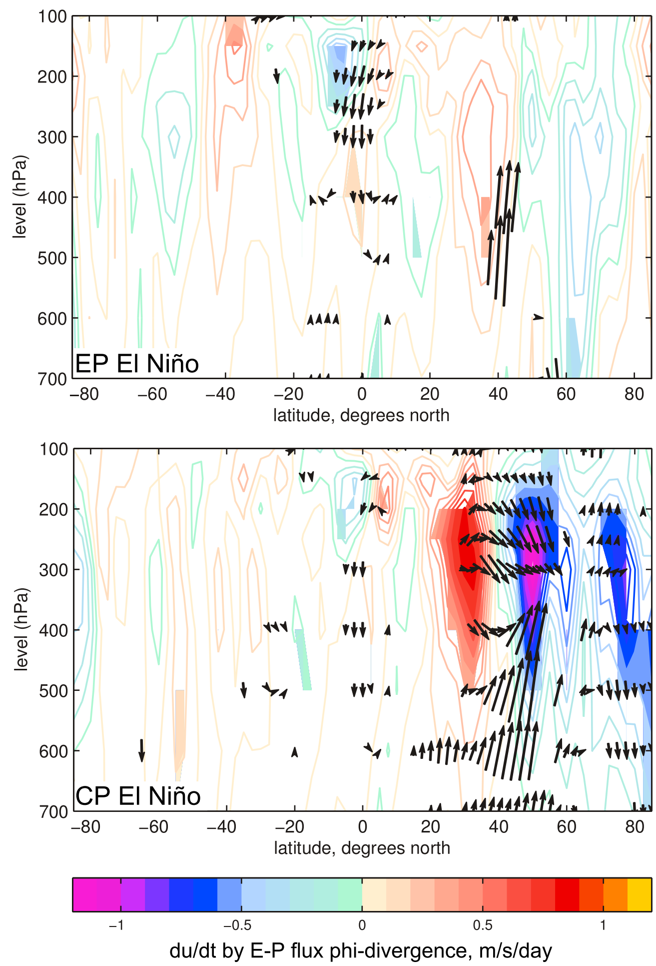

Figure 5. Eliassen-Palm (E-P) flux anomalies (arrows) and changes in the zonal wind variation (du/dt) due to the divergence of the meridional (phi-) component of the E-P flux for (top) CP and (bottom) EP El Niño winters. Only E-P flux anomalies significant at 95% confidence based on the Mann–Whitney U test are plotted. Zonal wind anomalies: representation as in Figure 2.

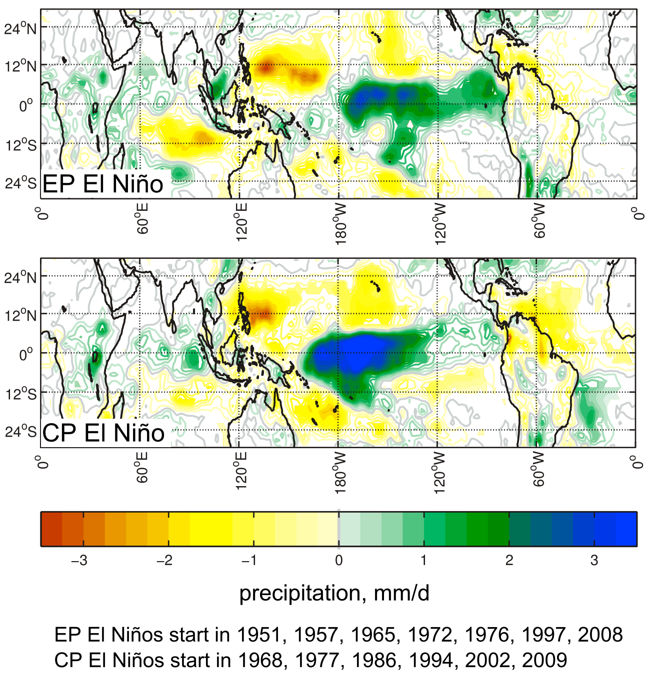

Figure 6. Tropical precipitation anomalies for (top) EP and (bottom) CP El Niños. Representation as in Figure 2.

straddling the date line for CP El Niños. During EP El Niños the STJ may not reach far enough into the East Pacific to trap meridionally propagating planetary waves generated by the East Pacific equatorial precipitation anomalies. During CP El Niños, however, the core of the slightly northward shifted East Asian STJ is stronger and extended over the central Pacific. This possibly enhances the chance of planetary waves generated by positive precipitation anomalies around the date line to become trapped during their meridional poleward propagation. Part of the perturbation response will then propagate zonally eastward along the STJ waveguide and cause perturbations over the Atlantic and Europe.

[23] The different atmospheric effects during EP and CP El Niños are clearly apparent in the winter anomalies of the horizontal stream function at the 0.258 sigma level (i.e., the upper troposphere at roughly 250 hPa). In the case of CP El Niños (Figure 7, right), the pronounced chain of cores of significant negative stream function anomalies spreading around the Northern Hemisphere resembles a zonally damped wave number 4–5 pattern that essentially follows the shape of the circum-global STJ. The subtropical centers of the negative anomalies are found over the eastern North Pacific, the subtropical western North Atlantic, central Europe, northern India, and the final and weakest off Japan. Interestingly, there seems to be a bifurcation of the wave train eastward of Europe into one branch extending toward northern India and another toward west Siberia. We find also centers of enhanced variability in the climatology of upper tropospheric stream function resembling the positions of these waves (not shown). The bifurcation of eddies for storm-track variability as discussed by *Orlanski* [2003] or persistent snow cover anomalies establishing in autumn over Siberia [*Cohen et al.*, 2007] might serve as explanations of this feature, and will be investigated in a later study.

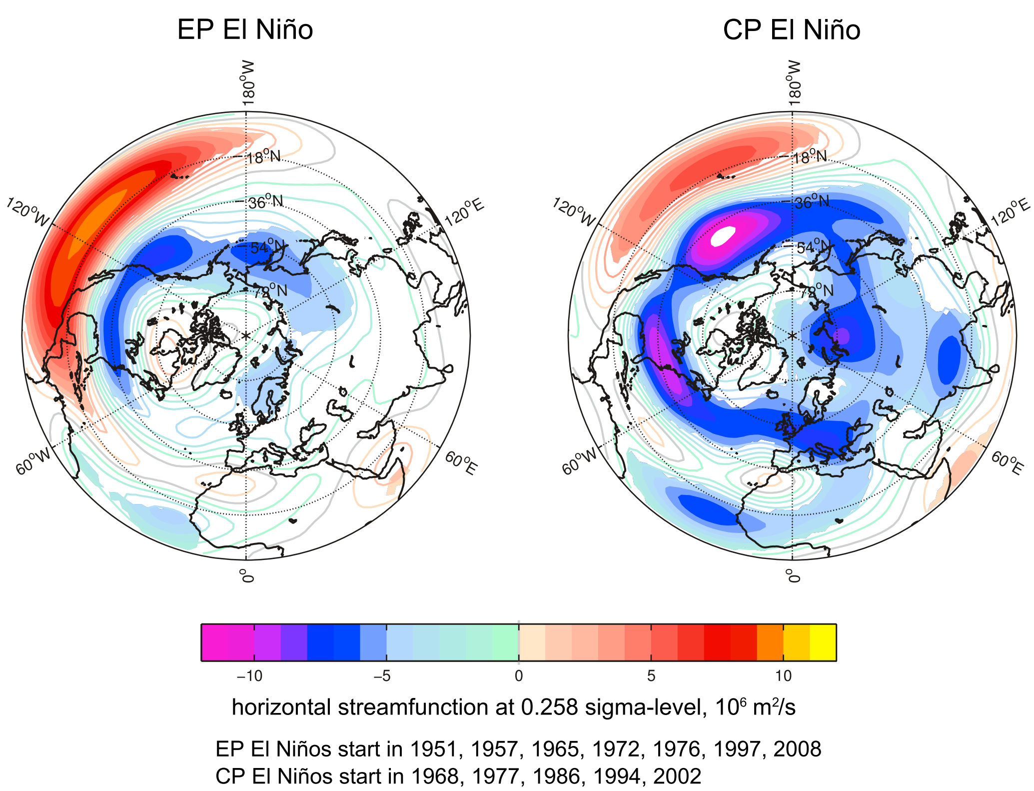

Figure 7. Horizontal stream function anomalies at the 0.258 sigma-level for (left) EP and (right) CP El Niño winters. Winters 2009–2010 are not included in the analysis.

[24] The strength of the stream function anomalies is strongest over the North Pacific suggesting that the wave train described above originates in the central North Pacific and then propagates eastward. In particular, the Atlantic centers weaken the Azores high and contribute significantly to the negative phase of the NAO. Such clear subtropical wave train is missing during the EP El Niño winters (Figure 7, left). The extension of the Asian STJ into the Atlantic is also missing during the 1994–1995 CP El Niño event, resulting in a positive pressure anomaly of the Azores High and a warm winter in Europe (not shown).

4. Discussion

[25] Our analysis suggests that there are distinct differences in the response of the global atmospheric circulation to the two different El Niño types. The results provide a possible explanation for the dual response patterns to El Niño that were diagnosed by *Gouirand and Moron* [2003] and *Brönnimann et al.* [2007]. While for EP El Niños there is a slight warming tendency over Europe, a field significance test for the Northern Hemisphere midlatitudes shows a high probability that these anomalies may occur by chance. A similar field significance test proves that midlatitude anomalies in near surface temperature and especially near surface pressure during CP El Niños are likely not due to chance. Europe becomes statistically significantly cooler in CP El Niño winters and a clear negative NAO phase is apparent from the pattern of statistically significant near surface pressure anomalies. During the CP El Niños 1914–1915, 1930–1931, and 1940–1941–1942, which are not included in our analysis, winters were as well characterized by negative values of the December to March NAO index (see J. Hurrell's PC-based index, available at http://www.cgd.ucar.edu/cas/jhurrell/indices.data.html#naopcdjf). During the 1930–1931 and 1940–1941–1942 CP El Niños, winters were also cold over central and eastern Europe, but winter 1914–1915 did not show unusual temperature anomalies over these regions.

[26] Our results are to a large degree consistent with the analysis of temperature and precipitation anomalies over the U.S. by *Larkin and Harrison* [2005], following the traditional El Niño definition by NOAA and a new one based on Niño3.4 SSTAs [*National Oceanic and Atmospheric Administration* (*NOAA*), 2003]. Nonetheless, some remarkable differences between their study and ours exist as well. The Nino3.4 index used by *Larkin and Harrison* [2005] only covers SSTAs between 170°W to 120°W and, thus, neglects information at and to the west of the date line. In their list, the El Niños starting in 1963, 1968, 1977, 1986, and 1994 are consistent with our CP event definition, but they define the El Niños starting in 1987 and 1991 as EP events, while these are mixed and CP El Niños, respectively, according to our definition. Supporting our definition, *Kug et al.* [2009] also define these two events as "mixed." Most importantly, *Larkin and Harrison* [2005] do not account for the effects of the three big volcanic eruptions that coincided with El Niños. Differences in the global surface air temperature anomaly patterns between the two El Niño types become more apparent when the volcanically disturbed winters are excluded from the analysis since the effect of volcanic aerosol loading is toward a strengthening of the stratospheric polar vortex, thus counteracting and in part overriding the effect of El Niño.

[27] The CP El Niño winter 1994–1995 proves to be an outlier in all investigated aspects. This event particularly increases the variability of the atmospheric response diagnosed for CP El Niños, but does not undermine the general validity of our arguments. The causes for the different response of the atmosphere to this particular El Niño event remain to be investigated.

[28] The different climate response to the two El Niño types can possibly be explained by the different location of anomalous convective heating along the equatorial Pacific relative to the STJ. Due to their more westward origin during CP El Niños, poleward propagating waves generated by enhanced precipitation at the equator may in part be trapped in the East Asian STJ and propagate zonally eastward along the subtropical waveguide. If the East Asian STJ is enhanced and extended eastward, as is the case for CP El Niños (Figures 4 and 5), this mechanism is further facilitated, leading to a zonal subtropical circum-global wave number 5 wave train in the upper troposphere as already discussed by *Branstator* [2002]. This wave train is responsible for a weakening of the Azores High, thus leading to a strong and significantly negative NAO phase during CP El Niño winters (Figure 3).

[29] Deep tropical convection preferentially occurs over SSTs exceeding a threshold of 26°C–28°C. The rather strong SSTAs during EP El Niños are located in the cool cold tongue area of the eastern Pacific. SSTAs during CP El Niños are generally weaker than those during EP El Niños, but they are added to the climatologically warmer waters usually characterizing the eastern edge of the Pacific warm pool. They lead to much stronger precipitation anomalies (Figure 6) and, hence, to excitation of stronger Rossby waves. Resulting from the meridional propagation of these stronger waves, as a local response over the North Pacific, a strong upper tropospheric cyclonic anomaly develops (Figure 7), which at its southern edge maintains the subtropical zonal waveguide. During EP El Niño winters, convective heating anomalies and related Rossby wave generation are weaker and farther east. This, as a local effect of meridionally propagating wave disturbances, corresponds to weaker and eastward shifted cyclonic anomalies over the North Pacific compared to CP El Niño winters; the STJ is not extended and the subtropical waveguide is not active. The described hypothetical mechanism may be regarded as nonlinear since the perturbation response is affected by modifications of the propagation properties resulting from the perturbations themselves. We are currently planning model experiments to further investigate this hypothesis.

[30] This mechanism (we call it "subtropical bridge") provides an alternative paradigm for interpreting Northern Hemisphere teleconnections, which contradicts the idea that the dynamical coupling of the troposphere with the stratosphere (i.e., the "stratospheric bridge") is the dominant factor regulating the joint variability in the Pacific and Euro-Atlantic sectors. Our hypothesis is supported by the clearer assessment of the relative importance of the stratospheric polar vortex for the NAO-like tropospheric variability during El Niño events, which is allowed by the separation between EP and CP El Niños. Indeed we find that both El Niño types lead to a weaker than normal winter polar

vortex in the lower stratosphere, a behavior that is stronger and statistically more significant for EP El Niños. Nonetheless, even under the comparatively weaker polar vortex conditions observed during EP El Niño winters, there is only a marginal tendency toward a negative NAO phase. This seems to arise especially from the development of a high pressure anomaly at Arctic latitudes indirectly affecting the Icelandic Low. Notably, *Castanheira and Graf* [2003] have shown that a significant correlation between the Aleutian Low and the Icelandic Low is only observed under strong stratospheric polar vortex conditions, but not when the polar vortex is weak. Since the "subtropical bridge" is not active in EP El Niño winters, a NAO response during EP El Niños is rather unlikely. This is confirmed by winter European surface air temperature anomalies which are slightly positive, but do not pass the attribution test during EP El Niño winters despite the concomitant weak stratospheric polar vortex.

[31] As suggested by *Vecchi et al.* [2006], one possible result of global warming is a progressive weakening of trade winds, which may lead to increased relative occurrence of CP El Niños [*Graf*, 1986] due to reduced wind-driven surface divergence at the equatorial Pacific. Hence, even though climate projections tend to indicate increasing global surface temperatures for the upcoming decades, the probability of occurrence of relatively cold winters in Europe may increase due to the effect of more CP El Niños on the strength of the Azores High, and hence on the NAO, via the subtropical bridge. Observations in recent years are at least not in contradiction with this suggestion.

[32] Some studies [e.g., *Merkel and Latif*, 2002] rather convincingly supporting the idea of El Niño having influences on European weather are based on model simulations that are biased toward CP El Niños [*Guilyardi*, 2006] with SSTAs more concentrated toward the central Pacific rather than along the East Pacific cold tongue region. The associated lack of capability to simulate a change in the prevailing El Niño type could have a detrimental effect on the capability of these models to accurately simulate future regional climate variability.

[33] Our results encourage further investigations on the efficiency of the subtropical waveguide for teleconnections originating from tropical convection anomalies at the eastern edge of the equatorial Pacific warm pool region and impacting on the strength of the Azores High and, hence, on the NAO index. We may speculate that as both model and reanalysis data are characterized by a positive bias in the climatological tropical deep convection over the Pacific warm pool [*Uppala et al.*, 2005; *Wagner*, 2009], the subtropical waveguide may be active quasi-permanently, leading to similar negative effects on the mean strength of the Azores High during boreal winter. This means that, independent of El Niño, those models systematically showing too high precipitation over the Pacific warm pool may systematically underestimate the strength of the Azores High in winter.

[34] Finally, since the strength of the lower stratospheric polar vortex and the STJ are inversely correlated as seen by the "tropospheric interannual oscillation" of *Chen et al.* [2002], it will be of great interest to revisit the mechanisms found behind the dynamical coupling between stratospheric and tropospheric circulation. It may well turn out that there is a "stratospheric bridge" link under very strong stratospheric polar vortex conditions due to reflection of planetary waves originating from the midlatitude troposphere impacting the strength of the Icelandic Low as suggested by *Castanheira and Graf* [2003]. Conversely, for weak polar vortex conditions and, hence, for a strong STJ, the Azores High may be affected by central Pacific convective heating anomalies via the subtropical bridge.

[35] The pattern of CP El Niño SSTAs matches the warm phase pattern of the Pacific Decadal Oscillation. It will be interesting to investigate to what degree the results of *Brönnimann et al.* [2007] and *Zanchettin et al.* [2008], who suggested a variable influence of El Niños on Europe climate and weather depending on decadal phases of SSTAs in the extratropical North Pacific, and the multidecadal variations of links between tropical Pacific SSTAs and tropical North Atlantic SSTs [*Walter and Graf*, 2002] can be explained by low frequency variability of active and inactive subtropical bridge affecting the strength of the Azores High. If the mechanism could be proved also for such longer time scales, it could be important for decadal climate predictability.

[36] **Acknowledgments.** This work was performed during visits of H.F.G. at the Max Planck Institute for Meteorology, Hamburg. H.F.G. thanks MPI for the support. D.Z. acknowledges funding by the ENIGMA project from the Max Planck Society. The authors are grateful to National Oceanic and Atmospheric Administration Earth System Research Laboratory/Physical Sciences Division for providing their reanalysis products at http://www.esrl.noaa.gov/psd/cgibin/data/composites/printpage.pl. The authors thank Felix Buntzel for providing his source code for the E-P flux diagnostics calculation and Sajjad Saeed for his helpful comments and suggestions. Two anonymous reviewers gave valuable advice that helped clarify the presentation.

References

Ambrizzi, T., and B. J. Hoskins (1997), Stationary Rossby-wave propagation in a baroclinic atmosphere, *Q. J. R. Meteorol. Soc.*, *123*(540), 919–928, doi:10.1002/qj.49712354007.

Andrews, D. G., J. R. Holton, and C. B. Leovy (1987), *Middle Atmosphere Dynamics*, 489 pp., Academic, San Diego, Calif.

Ashok, K., S. K. Behera, S. A. Rao, H. Weng, and T. Yamagata (2007), El Niño Modoki and its possible teleconnection, *J. Geophys. Res.*, *112*, C11007, doi:10.1029/2006JC003798.

Bell, C. J., L. J. Gray, A. J. Charlton-Perez, M. M. Joshi, and A. A. Scaife (2009), Stratospheric communication of El Niño teleconnections to European winter, *J. Clim.*, *22*(15), 4083–4096, doi:10.1175/2009JCLI2717.1.

Branstator, G. (2002), Circumglobal teleconnections, the jet stream waveguide, and the North Atlantic oscillation, *J. Clim.*, *15*, 1893–1910, doi:10.1175/1520-0442(2002)015<1893:CTTJSW>2.0.CO;2.

Brönnimann, S. (2007), The impact of El Niño-Southern Oscillation on European climate, *Rev. Geophys.*, *45*, RG3003, doi:10.1029/2006RG000199.

Brönnimann, S., E. Xoplaki, C. Casty, A. Pauling, and J. Luterbacher (2007), ENSO influence on Europe during the last centuries, *Clim. Dyn.*, *28*, 181–197, doi:10.1007/s00382-006-0175-z.

Cagnazzo, C., and E. Manzini (2009), Impact of the stratosphere on the winter tropospheric teleconnections between ENSO and the North Atlantic and European region, *J. Clim.*, *22*(5), 1223–1238, doi:10.1175/2008JCLI2549.1.

Castanheira, J. M., and H.-F. Graf (2003), 2003: North Pacific–North Atlantic relationships under stratospheric control?, *J. Geophys. Res.*, *108*(D1), 4036, doi:10.1029/2002JD002754.

Chen, W., H.-F. Graf, and M. Takahashi (2002), Observed interannual oscillations of planetary wave forcing in the Northern Hemisphere winter, *Geophys. Res. Lett.*, *29*(22), 2073, doi:10.1029/2002GL016062.

Cohen, J., M. Barlow, P. J. Kushner, and K. Saito (2007), Stratosphere-troposphere coupling and links with Eurasian land surface variability, *J. Clim.*, *20*, 5335–5343, doi:10.1175/2007JCLI1725.1.

Di Lorenzo, K., M. Cobb, J. Furtado, N. Schneider, B. Anderson, A. Bracco, M. A. Alexander, and D. Vimont (2010), Central Pacific El Niño and decadal climate change in the North Pacific, *Nat. Geosci.*, *3*, 762–765, doi:10.1038/NGEO984.

Eshel, G., M. A. Cane, and B. F. Farrell (2000), Forecasting eastern Mediterranean droughts, *Mon. Weather Rev.*, *128*, 3618–3630, doi:10.1175/1520-0493(2001)129<3618:FEMD>2.0.CO;2.

Gouirand, I., and V. Moron (2003), Variability of the impact of El Niño–Southern Oscillation on sea-level pressure anomalies over the North Atlantic in January to March (1874–1996), *Int. J. Climatol.*, *23*(13), 1549–1566, doi:10.1002/joc.963.

Graf, H.-F. (1986), On El Niño/Southern Oscillation and Northern Hemispheric temperature, *Gerlands Beitr. Geophys.*, *95*, 63–75.

Guilyardi, E. (2006), El Niño–mean state–seasonal cycle interactions in a multi-model ensemble, *Clim. Dyn.*, *26*, 329–348, doi:10.1007/s00382-005-0084-6.

Halpert, M. S., and C. F. Ropelewski (1992), Surface temperature patterns associated with the Southern oscillation, *J. Clim.*, *5*, 577–593, doi:10.1175/1520-0442(1992)005<0577:STPAWT>2.0.CO;2.

Ineson, S., and A. A. Scaife (2009), The role of the stratosphere in the European climate response to El Niño, *Nat. Geosci.*, *2*(1), 32–36, doi:10.1038/ngeo381.

Kalnay, E., et al. (1996), The NCEP/NCAR 40-year reanalysis project, *Bull. Am. Meteorol. Soc.*, *77*, 437–471, doi:10.1175/1520-0477(1996)077<0437:TNYRP>2.0.CO;2.

Kao, H.-Y., and J.-Y. Yu (2009), Contrasting eastern-Pacific and central-Pacific types of ENSO, *J. Clim.*, *22*, 615–632, doi:10.1175/2008JCLI2309.1.

Kaplan, A., M. Cane, Y. Kushnir, A. Clement, M. Blumenthal, and B. Rajagopalan (1998), Analyses of global sea surface temperature 1856–1991, *J. Geophys. Res.*, *103*, 18,567–18,589, doi:10.1029/97JC01736.

Kiem, A. S., and S. W. Franks (2004), Multi-decadal variability of drought risk eastern Australia, *Hydrol. Process.*, *18*, 2039–2050, doi:10.1002/hyp.1460.

Kiem, A. S., S. W. Franks, and G. Kuczera (2003), Multi-decadal variability of flood risk, *Geophys. Res. Lett.*, *30*(2), 1035, doi:10.1029/2002GL015992.

Kug, J.-S., F.-F. Jin, and S.-I. An (2009), Two types of El Niño events: Cold tongue El Niño and warm pool El Niño, *J. Clim.*, *22*, 1499–1515, doi:10.1175/2008JCLI2624.1.

Larkin, N. K., and D. E. Harrison (2005), On the definition of El Niño and associated seasonal average U.S. weather anomalies, *Geophys. Res. Lett.*, *32*, L13705, doi:10.1029/2005GL022738.

Merkel, U., and M. Latif (2002), A high resolution AGCM study of the El Niño impact on the North Atlantic/European sector, *Geophys. Res. Lett.*, *29*(9), 1291, doi:10.1029/2001GL013726.

National Oceanic and Atmospheric Administration (NOAA) (2003), NOAA gets U.S. consensus for El Niño/La Niña index, definitions, press release, Camp Springs, Md., 30 Sept. 2003. [Available at http://www.noaanews.noaa.gov/stories/s2095.htm.]

Orlanski, I. (2003), Bifurcation in eddy life cycles: Implications for storm track variability, *J. Atmos. Sci.*, *60*(8), 993–1023, doi:10.1175/1520-0469(2003)60<993:BIELCI>2.0.CO;2.

Perlwitz, J., and H.-F. Graf (1995), The statistical connection between tropospheric and stratospheric circulation of the Northern Hemisphere in winter, *J. Clim.*, *8*, 2281–2295, doi:10.1175/1520-0442(1995)008<2281:TSCBTA>2.0.CO;2.

Ren, H.-L., and F.-F. Jin (2011), Niño indices for two types of ENSO, *Geophys. Res. Lett.*, *38*, L04704, doi:10.1029/2010GL046031.

Ropelewski, C. F., and M. S. Halpert (1987), Global and regional scale precipitation patterns associated with the El Niño/Southern Oscillation (ENSO), *Mon. Weather Rev.*, *115*, 1606–1626, doi:10.1175/1520-0493(1987)115<1606:GARSPP>2.0.CO;2.

Steel, R. G. D., and J. H. Torrie (1960), *Principles and Procedures of Statistics*, 481 pp., McGraw-Hill, New York.

Trenberth, K., and C. J. Guillemot (1998), Evaluation of the atmospheric moisture and hydrological cycle in the NCEP/NCAR reanalysis, *Clim. Dyn.*, *14*, 213–231, doi:10.1007/s003820050219.

Uppala, S. M., et al. (2005), The ERA-40 re-analysis, *Q. J. R. Meteorol. Soc.*, *131*, 2961–3012, doi:10.1256/qj.04.176.

Vecchi, G. A., B. J. Soden, A. T. Wittenberg, I. M. Held, A. Leetmaa, and M. J. Harrison (2006), Weakening of tropical Pacific atmospheric circulation due to anthropogenic forcing, *Nature*, *441*, 73–76, doi:10.1038/nature04744.

Wagner, T. M. (2009), A dynamical convective cloud field model and the effects of aerosols, Ph.D. dissertation, Univ. of Cambridge, Cambridge, U. K.

Walter, K., and H.-F. Graf (2002), On the changing nature of the regional connection between the North Atlantic oscillation and sea surface temperature, *J. Geophys. Res.*, *107*(D17), 4338, doi:10.1029/2001JD000850.

Weng, H., K. Ashok, S. K. Behera, S. A. Rao, and T. Yamagata (2007), Impacts of recent El Niño Modoki dry/wet conditions in the Pacific Rim during Boreal summer, *Clim. Dyn.*, *29*(2–3), 113–129, doi:10.1007/s00382-007-0234-0.

Yeh, S. W., J. S. Kug, B. Dewitte, M. H. Kwon, B. P. Kirkman, and F. F. Jin (2009), El Niño in a changing climate, *Nature*, *461*, 511–514, doi:10.1038/nature08316.

Zanchettin, D., S. W. Franks, P. Traverso, and M. Tomasino (2008), On ENSO impacts on European wintertime rainfalls and their modulation by the NAO and the Pacific multidecadal variability described through the PDO index, *Int. J. Climatol.*, *28*(8), 995–1006, doi:10.1002/joc.1601.

Zanchettin, D., C. Timmreck, H.-F. Graf, A. Rubino, S. Lorenz, K. Lohmann, K. Krueger, and J. H. Jungclaus (2012), Bi-decadal variability excited in the coupled ocean–atmosphere system by strong tropical volcanic eruptions, *Clim. Dyn.*, doi:10.1007/s00382-011-1167-1, in press.

Zhang, C. (1993), Large-scale variability of atmospheric deep convection in relation to sea surface temperature in the tropics, *J. Clim.*, *6*, 1898–1913, doi:10.1175/1520-0442(1993)006<1898:LSVOAD>2.0.CO;2.

H.-F. Graf, Center for Atmospheric Science, University of Cambridge, Downing Place, Cambridge CB2 3EN, UK. (hfg21@cam.ac.uk)

D. Zanchettin, Max-Planck-Institute for Meteorology, Bundesstrasse 54, D-20146 Hamburg, Germany.