



Tropical stratosphere variability and extratropical teleconnections

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Thesis submitted in partial fulfilment of the requirements
for the degree of Doctor of Philosophy

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Abstract

The Quasi-Biennial Oscillation (QBO) is the dominant pattern of variability in the tropical stratosphere. Despite a well established theory regarding its generation in the atmosphere, the simulation in global climate models remains difficult. A set of metrics assessing the quality of model simulations is presented in this study. The QBO simulations in models submitted to the CMIP5 and CCMVal-2 intercomparison projects are characterised and compared to radiosonde observations and reanalysis datasets. Common model biases and their potential causes are addressed.

As the QBO has a long intrinsic period, knowing its influences on other parts of the climate system can be used to improve long range forecasts. These teleconnections of the QBO in observations are investigated using composite analysis, multilinear regression and a novel approach called causal effect networks (CEN). Findings from these analyses confirm previous results of the QBO modulating the stratospheric polar vortex and subsequently the North Atlantic Oscillation (NAO). They also suggest that it is important to take the equatorial zonal mean zonal wind vertical profile into account when studying teleconnections, rather than the more traditional method of using just one single level.

While QBO influences on the Northern Hemisphere winter polar vortex and the NAO are more clearly established, interactions within the tropics remain inconclusive. Regression analysis does not show a connection between the QBO and the MJO, whereas the CEN analysis does. Further studies are needed to understand the interaction mechanisms near the equator.

Finally, following the unprecedented disruption of the QBO cycle in the winter 2015/16, the event is described and a model analogue from the MPI-ESM-MR historical simulation is presented. Future implications are unclear, although model projections indicate more frequent QBO irregularities in a warming climate.

Acknowledgements

*“By three methods we may learn wisdom:
First, by reflection, which is noblest;
Second, by imitation, which is easiest;
and third by experience, which is the bitterest.”*
Master Kong

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I. Introduction

“Who ordered that?”

I. I. Rabi

This exclamation of surprise, which is credited to particle physicist I. I. Rabi on the discovery of the muon, could have been as well uttered by the first people investigating the upper atmosphere, Léon Teisserenc de Bort and Richard Aßmann. Their discovery of what became known as the stratosphere was indeed labelled as “the most surprising discovery in the whole history of meteorology” [Shaw, 1919, p. 225].

Up until the late 19th century, air measurements could be made only near the ground due to technical limitations of measuring and carrying devices. However, with the invention of the rubber balloon combined with self-recording instruments [Renard, 1892]¹, scientists could get a handle on the third dimension. So far this had only been possible from some higher mountains; those, however, bore the complication of their own micro-climate in contrast to the “free atmosphere” [Schmauß, 1933]. Gustave Hermite and Georges Besançon were the first to launch an unmanned balloon, the “Aérophile” (1893), followed by Aßmann’s “Cirrus” (1894) and very thorough explorations by Teisserenc de Bort (1898-1902) [Rochas, 2003].

What Aßmann and Teisserenc de Bort discovered when analysing their balloon data, was that the well-known temperature decline with altitude observed in mountain ascents stopped at about 10km height, giving rise to a layer where temperature remained seemingly constant, and even increasing slightly [Teisserenc de Bort, 1902]. Attributing this initially to measurement error due to solar radiation [Hermite, 1893; Teisserenc de Bort, 1898], after careful investigation (Teisserenc de Bort used data from 236 balloon soundings) the two researchers independently presented this “anomaly” to the German and French Academies of Science [Teisserenc de Bort, 1902; Aßmann, 1902]. It was subsequently coined the “isothermal layer” by Teisserenc de Bort [1902]².

¹Early manned balloon ascents to heights of over 10,000m have had limited success, with dangerous low levels of oxygen, that cost the life of the american aeronaut H. C. Gray, amongst others [Ryan, 2003].

²“zone isotherme” in the original French publication [Teisserenc de Bort, 1902].

The name “stratosphere” for the part of the atmosphere where temperature rises with height, was probably also introduced by Teisserenc de Bort [Shaw, 1919, p. 226]. It relates to the stratified characteristics of air with warmer layers above cooler ones, which is dynamically stable, hence no convection or turbulence occurring. This is in contrast to the “troposphere” (greek $\tau\rho\omicron\pi\acute{\eta}$ = turn, change), where the air is subject to vertical mixing. Despite the name, the stratosphere is all but quiet: there are radiative and chemical processes as well as dynamical ones, which are influenced by waves emerging from the troposphere.

A combination of chemical and dynamical processes play a role in the ozone budget of the atmosphere. While it was noted by Cornu [1879] that the incident solar spectrum near the Earth’s surface showed a huge discrepancy in the infrared and ultraviolet part and that the latter was due to absorption by ozone [Hartley, 1880], early investigations of the atmosphere did not find large amounts of O_3 in the lower atmosphere, the conclusion being that it had to be found higher up: in the stratosphere. With Gordon Miller Bourne Dobson inventing a spectrometer that could be used for ground based observations in 1924 and having these devices installed all over the world, the discovery of the height distribution of ozone peaking above the tropopause was not a surprise. What seemed strange, however, was the distribution in latitude with higher concentrations of ozone outside the tropics [Dobson *et al.*, 1929], where most of it should be produced due to the production process depending on UV radiation. Similarly, the helium and water vapour content in the atmosphere investigated by Brewer [1949], could only be explained by a large scale circulation involving motion through the stratosphere. Following their investigations, this phenomenon of tropical air moving poleward at high altitudes before descending in mid to high latitudes was later labelled the “Brewer-Dobson circulation” (c.f. Butchart [2014]).

Another conspicuous example of dynamics in the stratosphere - yet another surprising discovery - are sudden stratospheric warmings (SSWs), first observed in 1952 [Scherhag, 1952], which are thought to play a significant role in European winter climate [Baldwin & Dunkerton, 2001]. This phenomenon, characterised by a sudden rise in arctic stratospheric temperatures coinciding with circulation changes of the polar vortex, raised questions about dynamics in the upper atmosphere and its connection with the troposphere below.

In contrast to some other areas of physics, climate physicists face huge challenges in answering questions as the only available “laboratory” is Earth itself (or indeed other planets), so basic principles of experimentation (like localised control or repeatability) cannot be met. These scien-

tists' toolboxes are therefore computer models, which are used to represent the known processes in the Earth's system. However, not all processes and interactions are known, computing power is limited and a lot of the underlying equations lead to chaotic solutions. Even the best models will still be an approximation, as well as experiments in a confined and defined laboratory environment would be.

Looking at the representation of the stratosphere, for example, SSWs; though being well recognised in the measurement record, they are not well represented in global climate models (GCMs) without resolving the whole stratosphere [Charlton-Perez *et al.*, 2013]. Hence, influences from the stratosphere on the troposphere, are not represented by these so called “low top” models³, even though they have been recognised as being important for reliable assessment and prediction of tropospheric weather and climate. Taking this into account [IPCC, 2013], the number of GCMs resolving the whole stratosphere has risen from 4 (of 23) to 25 (of 37) models in the Coupled Model Intercomparison Project (CMIP) between phases 3 and 5 [Manzini, 2012].

Another important dynamical feature of the stratosphere is the Quasi-Biennial Oscillation (QBO), which has been observed for about 60 years now and is still a challenge to simulate [Hamilton *et al.*, 2015]. It appears as an oscillation with an approximate 28 month period in zonal wind shear zones descending through the tropical stratosphere between about 10 and 100hPa; however, influences on the (extra-)tropical troposphere have been found in numerous variables (c.f. Bhalme *et al.* [1987]; Randel & Wu [1996]; Huang *et al.* [2008] and references therein). Even the UK might attribute some credit of the extreme windy, warm and wet winter conditions during 2013/14 to the QBO [Met Office, 2014; Scaife *et al.*, 2014]. Although a number of studies suggest a connection between the QBO and surface patterns (c.f. Holton & Tan [1980]; Baldwin & Dunkerton [2001]; Claud & Terray [2006]; Anstey & Shepherd [2014] and references therein), several questions remain, for example: How strong is the connection? What is the mechanism behind the influence? Can models predict it?

Addressing these questions and whether models can simulate a QBO at all is the objective of the author's studies which are presented in the following chapters.

³If the highest level is below the stratopause (~ 1 hPa) models are classified as “low top”, otherwise as “high top”.

II. Background

The observed monthly mean zonal mean zonal wind, $\bar{u}$, in the tropical troposphere and stratosphere consists of different parts (Figure II.1). At ~ 1 hPa height, the 6-month cycle of the semiannual oscillation (SAO) dominates, below ~ 100 hPa, the oscillation is of an annual period (Annual Oscillation (AO), which also exhibits a high variance at about 1hPa [Butchart *et al.*, 2011]); in the lowermost troposphere, winds are constantly easterly. The part between ~ 100 hPa and ~ 10 hPa, however, exhibits an oscillation that is not directly coupled to an annual or semiannual cycle. Instead, its period is slightly longer than two years, hence the name “Quasi-Biennial Oscillation” (QBO, Figure II.2).

II.1. History of QBO Observations

The discovery of the QBO by Ebdon [1960] and Reed *et al.* [1961] continued the series of rather surprising observations in the stratosphere. The phenomenon was found in ballon-borne observations of the zonal wind patterns at the equator over a period of 5 years. Up to this point, observations have been very much singular in time: After the eruption of Mt. Krakatau in 1883, its plume circled the earth about 1-3 times from East to West¹, at about 30 km/h [Hamilton,

¹The terminology of wind movement (in contrast to ocean currents) depends on where the wind is coming from: Wind from East to West is labelled “easterly”, whereas wind from West to East is labelled “westerly”. Therefore, easterly zonal wind has negative and westerly zonal wind positive sign in local east, north, up coordinate systems.

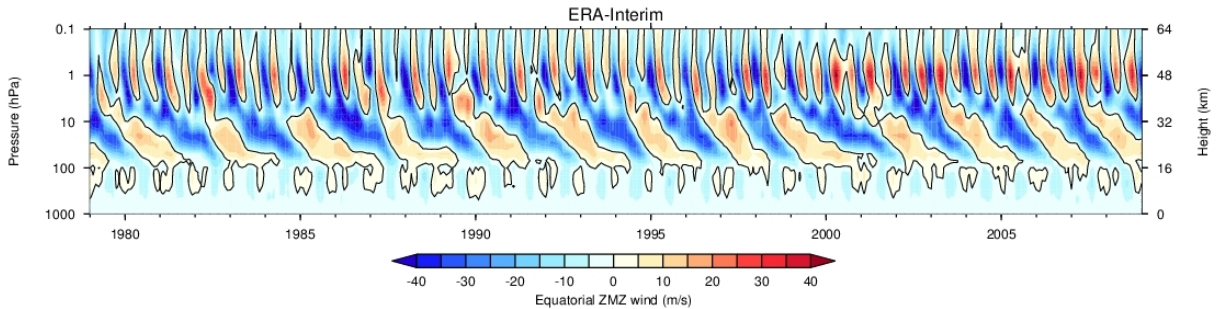


Figure II.1.: Equatorial zonal mean zonal wind time-height series from the ERA-Interim re-analysis (1979-2009) [Dee *et al.*, 2011]. Negative values of $\bar{u}$ are related to Easterly wind (blue) and positive values to Westerly wind (red). The zero wind line is shown in black.

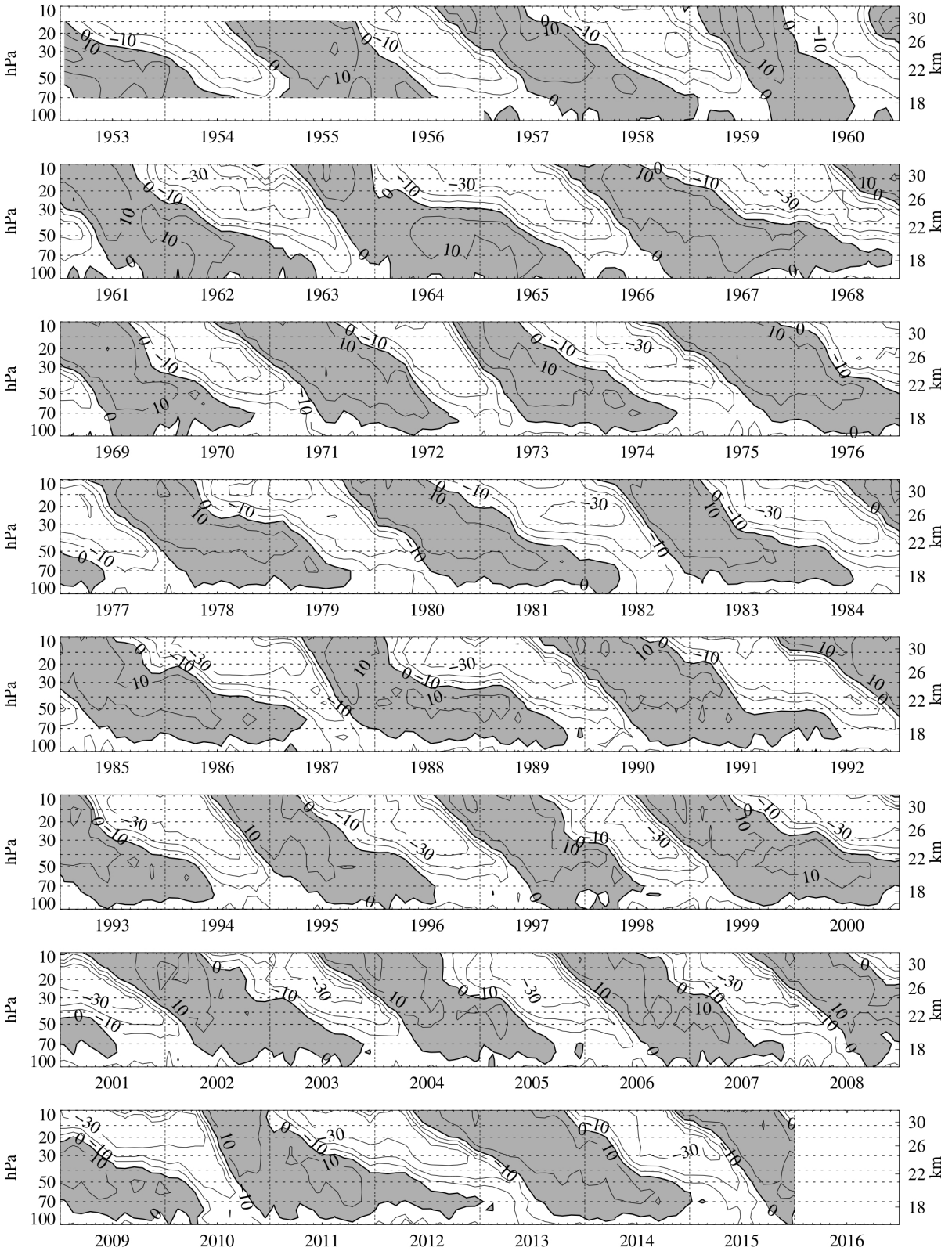


Figure II.2.: Zonal wind observations from equatorial stations: Canton Island, 3°S/172°W (Jan 1953 - Aug 1967), Gan/Maledive Islands, 1°S/73°E (Sep 1967 - Dec 1975) and Singapore, 1°N/104°E (since Jan 1976). Isopleths are at 10 m/s intervals; westerlies are shaded. Plot from Freie Universität Berlin [2016].

2012]. Afterwards, strong easterly winds were observed in the equatorial regions (Java, Lake Victoria, Batavia), that were subsequently named “Krakatau easterlies” [von Hann & Süring, 1915]. These winds seem to have been accepted as a permanent characteristic of the equatorial stratosphere, though von Hann & Süring [1915] noted that there are sometimes westerly winds embedded in the strong easterlies, which had been observed after the eruption of the Semeru (Java) in 1911 and also by Berson [1910] during his East Africa expedition. However, there was no explanation yet for these surprising westerlies [von Hann & Süring, 1915], that were named after Berson, who was the first to report them.

Having noted that “the salient facts about the general circulation of the tropical atmosphere below 13km are now known”, Palmer [1954] describes the general circulation in the tropical stratosphere as consisting of the Krakatau easterlies (between 15°North and South; height region between 30km and 40km) and the Berson westerlies (7°South to 7°North, from 20km to about 27km height, with the upper boundary varying from month to month). The first evidence of winds contradicting this general rule was produced by McCreary for monthly wind averages on Christmas Island. Between October 1956 and 1958 he observed a descending westerly wind regime that was followed by a descending easterly one. Looking at this very short time period, however, he could only draw the conclusion that “there is insufficient data for the equatorial stratosphere to warrant an estimation as to which year is more nearly normal.” [McCreary, 1959]. A similar short period (January 1957 to January 1958) was investigated by Ebdon [1960], who noted a reversal of the zonal wind direction in the stratosphere that appeared to be independent of longitude. From observations of the monthly mean zonal wind at Canton Island (50mb level) in 6 subsequent Januarys (1954-1959), Ebdon [1960] noted that there was a “remarkable 2-yr periodicity”, hence easterlies and westerlies would occur equally likely and thus both of McCreary’s states would be normal. This observation was backed up by Reed *et al.* [1961], who looked at stratospheric winds between July 1955 and February 1960 at 2 equatorial stations, Canton Island and Nairobi, Kenya. What they found became known as the Quasi-Biennial Oscillation: the monthly mean zonal wind does not exhibit a fixed pattern of Berson westerlies and Krakatau easterlies, but rather consists of descending and alternating shear zones, that emerge at about 10 hPa and diminish between 70hPa and 100hPa. Despite their short record, the characteristics of the “oscillation” Reed *et al.* [1961] described without making a claim about their general occurrence, have mostly been backed up by the observations up to the present day, such as:

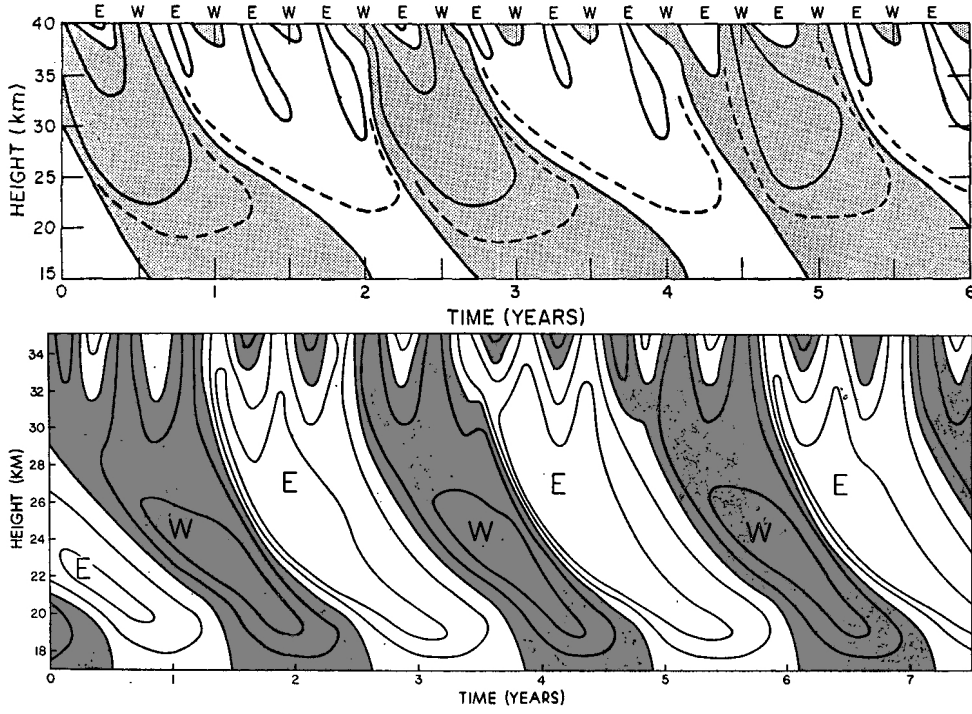


Figure II.3.: Time-height cross sections of equatorial zonal wind from (upper) Lindzen & Holton [1968] and (lower) Holton & Lindzen [1972] of early one-dimensional QBO modelling experiments. Contour lines are at 20 m/s (upper) and 10 m/s (lower), westerly flow is shaded.

- The approximate period of 14 months for the wind reversal at each pressure level
- The descent rate of about 1km per month for the shear zones
- The weakening of the shear zones close to the tropopause
- The asymmetry in strength of the shear zones and period, depending on the height level
- The longitudinal symmetry
- The maximum amplitude at about 25km (~ 28 hPa)

These characteristics were further established in the longer observational record (1953-1995) including: an average period of 28.1 months, the asymmetry in the shear zones amplitudes, with stronger maximum easterly wind and slower descent of the easterly shear zones. The QBO was found to have a latitudinal extent of 12° half-width at half maximum and a vertical extent between 16 and 50km height [Baldwin *et al.*, 2001]. An updated analysis of the different characteristics of the stratospheric zonal wind record from 1956 to present as well as a more comprehensive set of characteristic QBO parameters are the subject of sections IV.1 and IV.3.

II.2. Development of a theory

Reed *et al.* [1961] not only described the stratospheric wind oscillation later to be known as the QBO, but also provided some theoretical arguments underpinning these observations. Most importantly, they posited the downward propagation of the shear zones to the action of atmospheric waves, basing this on the ozone distribution and that “mass transport is more likely to be upward than downward”. However, the explanation for a zonally symmetric ring of westerlies at the equator was still difficult; Reed *et al.* [1961] suggested tropical disturbances carrying westerly momentum as the cause, in analogy to the westerly jet stream in the mid-latitudes.

Despite Angell & Korshover [1963] noting the importance of studying “non-annual or non-diurnal oscillations in geophysics” and predicting a rational explanation for the quasi-biennial wind and temperature oscillations to be found soon, it was five years before Lindzen & Holton [1968] published “A Theory of the Quasi-Biennial Oscillation”. While previous models had failed to explain either the quasi-biennial period, the constant amplitude of the shear zones during downward propagation to about 23km height or the occurrence of westerlies at the equator, Lindzen & Holton [1968] were able to resolve these problems using their one-dimensional model of vertically propagating gravity waves interacting with the background zonal wind. In several two-dimensional numerical experiments they showed that an oscillation in zonal wind can be caused by equatorial gravity waves undergoing critical level absorption. However, the period of the oscillation was confined to a multiple of six months and the SAO in the upper stratosphere (above 40km) was necessary for the onset of a new shear zone. Re-examination of their theory [Holton & Lindzen, 1972] in the light of observations of waves at the equator led to a refinement regarding the interaction mechanism of the waves with the flow (thermal damping rather than critical level absorption), reinforcing the main hypothesis of momentum deposition from waves causing the QBO. Other important findings were that the wave spectrum range in phase speed could be smaller and the SAO was not required anymore for QBO generation [Holton & Lindzen, 1972]. The results of their early modeling studies can be seen in Figure II.3.

Some key features of the QBO could already be reproduced with these simplified models, like the longer easterly period in the upper atmosphere, the prevalence of the phase onset with the SAO in the upper stratosphere (compare Figure II.1), constant maximum amplitudes down to about 23km and, in the earlier model, the slower descent of the easterly shear zone. However, the asymmetry in the peak velocity of easterly and westerly shear zones is not well represented,

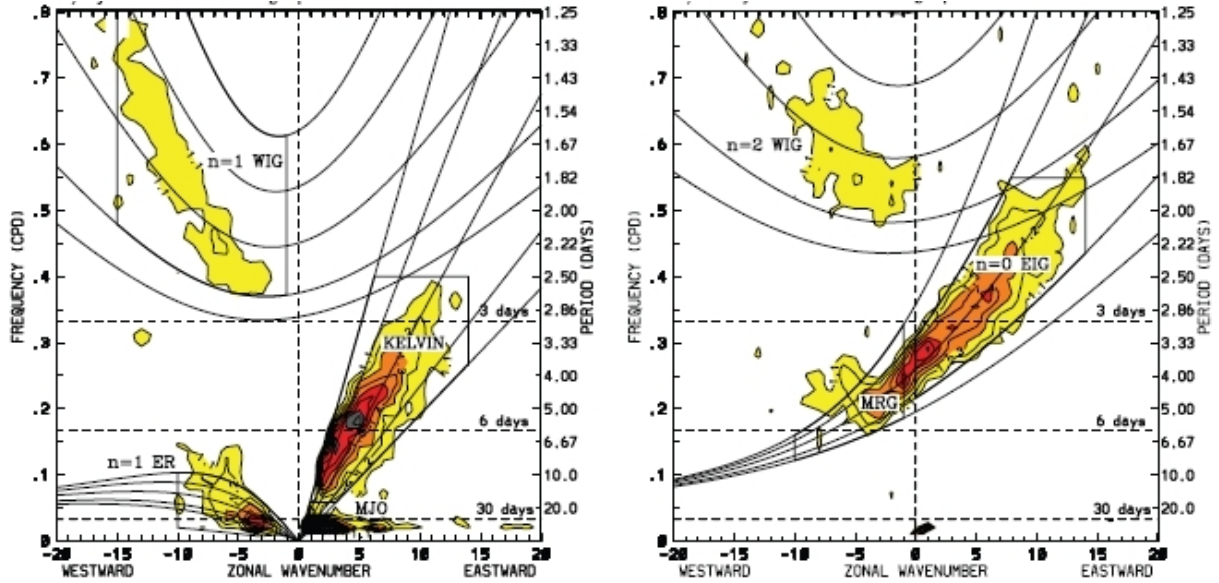


Figure II.4.: Symmetric (n odd, left) and antisymmetric (n even, right) wave spectra in the equatorial (15°S to 15°N) troposphere, derived from Cloud Archive User Services brightness temperature data. Dispersion curves are plotted for different wave types at several vertical wavenumbers. Contours show significant ($\geq 95\%$) signals in the wavenumber-frequency domain: Eastward/Westward Inertio Gravity waves (EIG/WIG), Kelvin waves, equatorial Rossby waves (ER) and mixed Rossby-Gravity waves (MRG). MJO denotes the Madden-Julian Oscillation. Figure from Kiladis *et al.* [2009].

neither are any irregularities of the cycles (c.f. shear zone descent rate, period, amplitude).

Parallel to the theory attributing the QBO generation to waves, the theory explaining the latter, was developed. Matsuno [1966] showed that the solutions to the linearized shallow-water equations for perturbations on a motionless basic state of mean depth h_e on the equatorial β -plane consist of different types of waves (defined by their dispersion relation): The planetary scale Kelvin and Rossby waves, as well as smaller scale Inertio-Gravity-waves. One special case is the mixed Rossby-Gravity wave, for which behaviour differs depending on frequency. An overview of the structure of these wave types is given in Figures II.4 and II.5. Early observations of Maruyama [1967] were consistent with the occurrence of a Rossby-Gravity wave at the equator that was identified to provide easterly acceleration to the QBO [Bretherton, 1969]. Westerly momentum, however, was thought to come from equatorial Kelvin waves [Wallace & Kousky, 1968].

The theory of wave-mean flow interaction was backed up by an impressive laboratory experiment: Plumb & McEwan [1978] forced a standing wave - two waves of equivalent amplitude and period, travelling in opposite direction - at the bottom of a cylinder filled with salt-stratified water. Their setup followed the idea of Plumb [1977], who showed that the process forcing the

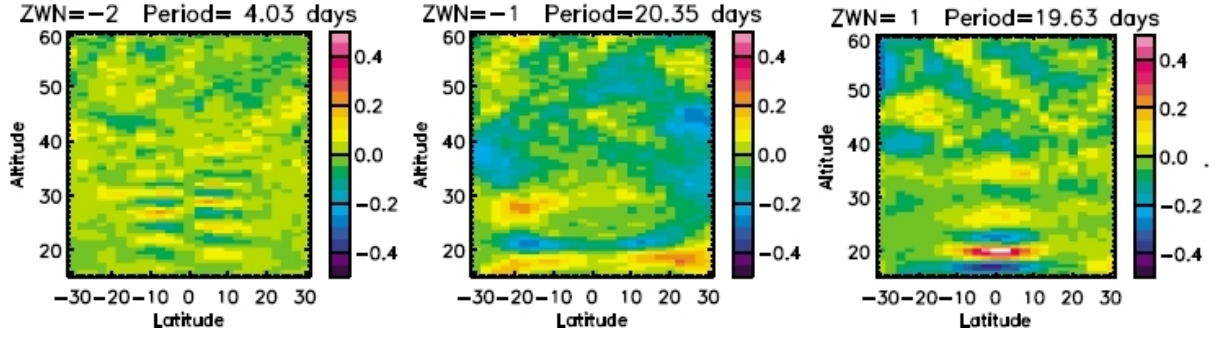


Figure II.5.: Latitude-height structure of different wave types derived from High Resolution Dynamics Limb Sounder temperature data. From left to right: Mixed Rossby-gravity wave at zonal wave number (ZWN) $k = -2$, equatorial Rossby wave ($k = -1$) and Kelvin wave ($k = 1$). The colour scale is temperature in K. The shown height range corresponds to about 100 hPa to 0.1 hPa, with the QBO being dominant in the region from 16-32km (10-70hPa). Figure from Alexander & Ortland [2010].

QBO was not exclusive to the equatorial stratosphere, but rather a general phenomenon of fluid dynamics, making it suitable for exploration in the laboratory. The experiment, which indeed produced an oscillatory mean flow and showed a good agreement with theoretical calculations, supported strongly the understanding of the mechanisms driving the QBO.

Wave-mean flow interaction is commonly accepted as the process generating the QBO. The generalised theory is also backed up by the observation of a quasi-quadrennial oscillation on Jupiter [Leovy *et al.*, 1991; Friedson, 1999] and an oscillation of about 15 years in Saturn’s equatorial stratosphere² [Fouchet *et al.*, 2008; Orton *et al.*, 2008]. The current understanding of the generation of an oscillation of zonal wind in the equatorial stratosphere and its implications for the temperature has been outlined by Baldwin *et al.* [2001]:

The principle underlying descending wind anomalies is exemplified by two vertically propagating internal gravity waves that are of the same amplitude, but opposite phase speeds c . These waves encounter a background state of zonal wind $\bar{u}$. In layers where $|\bar{u} - c|$ is small, the waves are damped, depositing momentum in the zonal flow. This leads to easterly momentum accelerating the air at the lower boundary of an easterly shear zone, with westerly waves passing through that layer freely. Figure II.6 schematically explains the process for one QBO cycle: At low levels within the stratosphere, viscous diffusion will damp out the anomalies, letting waves pass that build up a new shear zone above. In the upper stratosphere, the SAO might play a role in the onset of a new phase [Kuai *et al.*, 2009], although it is found not to be necessary in numerical simulations [Holton & Lindzen, 1972; Gray & Pyle, 1989].

²This is actually more a semi-annual oscillation in Saturn years, which are about 29 Earth years.

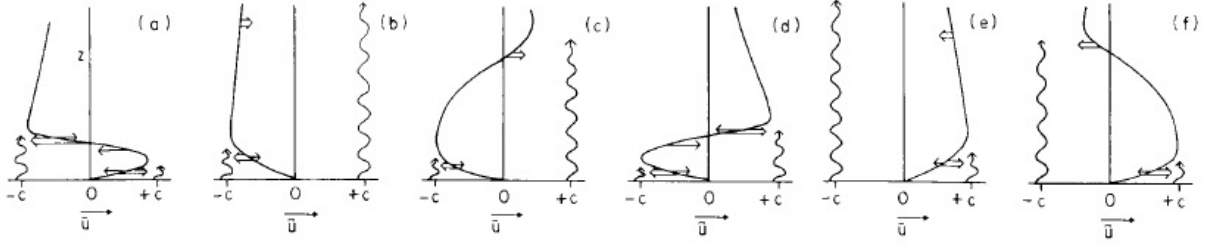


Figure II.6.: The descent of the QBO shear zones due to wave-mean flow interaction. Shown is a profile of zonal mean wind $\bar{u}$ that evolves with time from left to right. Two waves of phase speed $-c$ (eastward) and $+c$ (westward), depicted by wavy lines, travel upward and deposit momentum when their phase speed approaches the background wind speed. Acceleration of the zonal wind due to waves is indicated by double arrows, single arrows show viscous acceleration. Figure from Plumb [1984].

The latitudinal structure of the QBO, found to be approximately Gaussian, centered at the equator, and extending to about 12° North and South, might be explained by equatorially trapped waves (e.g. the Kelvin waves). Another possibility, pointed out by Haynes [1998], is the Coriolis force that plays an important role in rotating systems, in this case balancing part of the applied force. A dimensional analysis shows that for small values of the Coriolis parameter $2\Omega \sin \phi$ (where Ω is the rotation rate of the Earth and ϕ is latitude), the mean flow acceleration is basically accounted for by the applied force. For larger values, however, the Coriolis force cancels the applied forcing, leading to a change in meridional circulation instead; Lindzen & Holton [1968] already showed a reduced amplitude of the QBO at higher latitudes due to the Coriolis torque. Haynes [1998] predicted a latitudinal scale of about 10° with a simple model, using latitudinally symmetric wave forcing.

Longitudinally, the QBO is found to be approximately symmetric [Belmont & Dartt, 1968]; considering the processes involved, such as planetary scale waves and the Coriolis force, this would be expected. Some variations due to differences in the gravity wave spectrum, that might arise from different source regions, are small compared to the total amplitude of the oscillation and thus negligible.

In addition to the oscillation in zonal wind, an oscillation in temperature is observed [Angell & Korshover, 1964]. This is explained by the atmosphere being approximately in thermal wind balance [Andrews *et al.*, 1987a] near the equator:

$$\frac{\partial u}{\partial z} = -\frac{R}{H\beta} \frac{\partial^2 T}{\partial y^2} \underset{\text{Tropics}}{\sim} \frac{R}{H\beta} \frac{T}{L^2}$$

with zonal wind u , Temperature T , log-pressure height z , latitude y , gas constant for dry air R ,

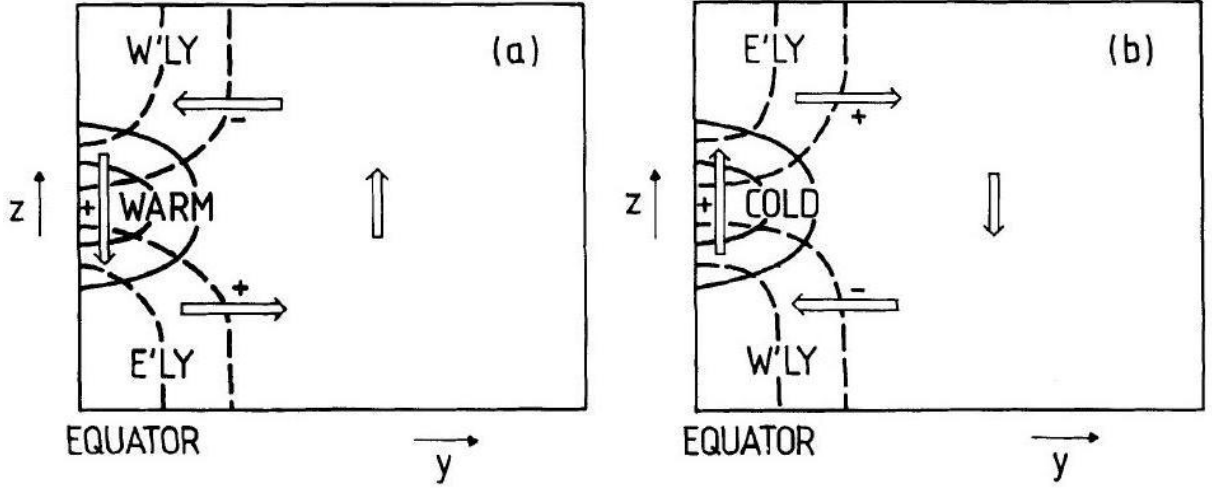


Figure II.7.: Schematics explaining the interaction of the QBO with the meridional circulation. Solid contours show temperature anomalies, dashed contours zonal wind anomalies. The arrows depict the meridional circulation, with plus and minus signs indicating the zonal wind acceleration. (a) is for a descending westerly, (b) for a descending easterly shear zone. Figure from Plumb & Bell [1982].

scale height $H \approx 7$ km, meridional scale L and latitudinal derivative of the Coriolis parameter β .

For the observed wind shear ($40\text{--}50 \frac{\text{m/s}}{\text{km}}$) between 10 and 70 hPa and meridional extent (10°), temperature anomalies of order $\mathcal{O}(1^\circ)$ would be expected, with the greatest anomalies coinciding with a maximum of $\frac{\partial u}{\partial z}$, occurring in the transition regions of the shear zones. Following this reasoning, descending westerly over easterly winds ($\frac{\partial u}{\partial z} > 0$) lead to a positive temperature anomaly at the equator that strengthens the descent in the meridional circulation (Figure II.7), whereas descending easterly shear zones have the opposite effect. This vertical advection of zonal momentum from the secondary circulation also explains why westerly shear zones descend faster than easterlies [Plumb & Bell, 1982].

Additionally, stratospheric temperatures are modulated by radiative processes involved in chemical reactions, such as ozone production and depletion. O_3 is produced by an oxygen radical O binding to molecular oxygen O_2 and destroyed by UV light in the reverse process. The presence of ozone leads to higher temperatures due to absorption of short- and longwave IR radiation, while its depletion has a cooling effect. From a modelling perspective, it is vital to distinguish the effects of dynamics and chemistry on temperature to better understand their relative importance.

Figure II.8 shows ozone measurements from radiosondes and atmospheric radars over Thumba (8.5°N , 77°E) for the period 2002–2007. The highest O_3 concentration can be seen in the lower

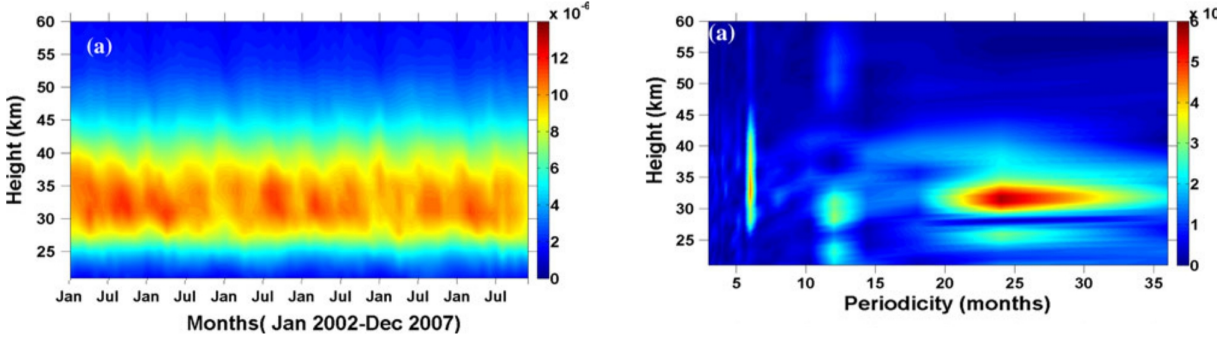


Figure II.8.: Left: Time-height series of ozone mixing ratio in the atmosphere 2002-2007 over Thumba (8.5°N, 77°E) as measured by radiosondes and atmospheric radars. Right: Fourier spectra of the timeseries. The height range shown corresponds to 55hPa - 0.2hPa. Figure from [Kumar *et al.*, 2011].

and middle stratosphere (the so called ozone layer). Fourier analysis of this timeseries shows strong semiannual modulation in the stratosphere, an annual oscillation in the lower stratosphere and a broad spectral peak at the QBO periods (22-30 months). The QBO modulation of ozone has been found in satellite data (e.g. Randel & Wu [1996]; Huang *et al.* [2008]) as well. Two peaks in height occur in the ozone Fourier spectrum at QBO periods. Huang *et al.* [2008] studied the relationship between ozone and temperature and find a positive correlation between the two in the height region of 30-80km, and an anticorrelation below 30km. The difference between the two parts of stratosphere can be explained by the domination of (fast) radiative processes in the upper stratosphere and slower temperature-dependent ones in the lower stratosphere. This finding can be used to identify ozone-temperature feedbacks as well as the relative importance of chemistry and dynamics as described by Brasseur & Solomon [2005].

II.3. Modelling the QBO

After the first successful simulations in simplified two- and one-dimensional models [Lindzen & Holton, 1968; Holton & Lindzen, 1972], understanding of the principle dynamics driving the QBO seemed to be established and supported by observational data. However, these early models were confined to tropical latitudes and, more importantly, represented dynamical aspects only. Gray & Pyle [1989] introduced a parametrization of the QBO into a more realistic global 2D model that included a full radiation scheme and a comprehensive stratospheric ozone scheme to try to explain the observed QBO signal in column ozone and extratropical temperatures. Their model included the effects of two Kelvin waves (with different phase speed) and a Rossby-gravity wave, which provided the momentum for zonal wind acceleration. Their modelled oscillation (Figure

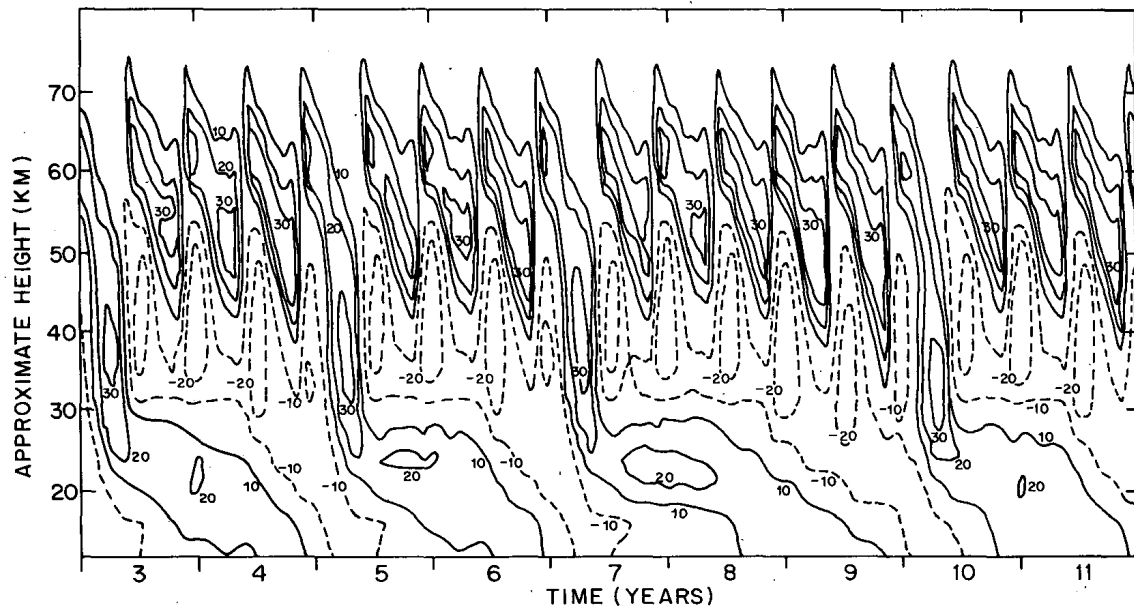


Figure II.9.: Time-height cross sections of equatorial zonal wind from Gray & Pyle [1989] two-dimensional dynamical/chemical QBO modelling experiment. Contour lines are at 10 m/s, the zero contour has been omitted.

II.9) resembled the observed QBO; the modelled semiannual oscillation and variations in ozone were in good agreement with the observations as well. However, the amplitude (and thus the momentum flux) of the waves incorporated in the simulation exceeded the observed values by a factor of three. Further, the peak amplitude of the westerlies exceeded the one of the easterlies, while the observations show the opposite.

The addition of chemistry was not the only change to the early dynamical models that required higher momentum transfer by equatorial waves: the extension to global domains meant including the effects of mid-latitude planetary scale (Rossby) waves and hence a representation of the Brewer-Dobson circulation. This results in air undergoing an upward motion in the equatorial lower stratosphere (rate of upwelling ~ 1 km/month), opposing the descent of wind regimes from above [Dunkerton, 1991]. Observations, however, showed that Kelvin and Rossby-gravity waves alone cannot provide the necessary momentum flux for forcing the QBO winds, which led Dunkerton [1997] to the conclusion that a broad spectrum of gravity waves, already proposed by Lindzen & Holton [1968], was required to provide the additional momentum. Although observations at that time could not resolve these waves, present day satellite observations indeed show a broad spectrum of equatorial and extratropical waves (Figure II.4, II.5).

However, even with these advancements in theory, one problem remained when it came to simulating a QBO in a Global Climate Model (GCM): parts of the wave spectrum cannot be

Wavetype	Kelvin	Rossby	Gravity (lf/hf)		Mixed Rossby-Gravity
Period	$\gtrsim 3$ days	–	1-3 days	$\lesssim 1$ day	$\gtrsim 3$ day
Zonal wavenumber	1-4	–	4-40	> 40	1-4
Zonal wavelength	$\gtrsim 10,000$ km	–	1000-10,000 km	$\lesssim 10$ km	$\gtrsim 10,000$ km
Propagation	eastward	westward	east- or westward		westward
Momentum flux	eastward	east- or westward	in wave direction		eastward

Table II.1.: Characteristics of the relevant waves for QBO forcing as described by Baldwin *et al.* [2001]. Propagation and momentum flux are describing the direction of phase and group velocity in zonal direction, respectively. Rossby waves are thought only to play a minor role in QBO forcing [Takahashi & Holton, 1991]. Gravity waves are divided into lower frequency (lf) inertia-gravity waves and high frequency (hf) gravity waves.

resolved because either the relevant wavelengths are shorter than the grid cell dimensions or their sources, like topography, are not well represented. This means that the effect of these waves (i.e. momentum transfer in wave-mean flow interaction) has to be parametrised. Currently, two main parametrisations are employed in many GCMs that simulate a QBO: the Doppler Spread Parametrisation (DSP) scheme [Hines, 1997a,b] and the ultrasimple spectral parametrisation (USSP) scheme [Warner & McIntyre, 2001] (Figure II.10). They differ in what part of the wave spectrum is absorbed during upward propagation and how vertical propagation, dissipation and mean flow forcing is modelled.

Another issue affecting QBO simulation is that the stratosphere needs to be simulated to high levels (Osprey *et al.* [2013] found no QBO in the HadGEM2-CC model up to 39km, but one in the configuration up to 60km), which is computationally expensive.

With these challenges, it took over 40 years from the discovery of the QBO until the first simulation in a full 3D GCM [Takahashi, 1996]. Following this initial success, a number of studies describing simulations that captured a QBO signal were, among others, Horinouchi & Yoden [1998]; Takahashi [1999]; Scaife *et al.* [2000] and Hamilton *et al.* [2001]. Despite these studies representing the QBO and other aspects of variability fairly well, it is not clear which aspects of the model configurations are required to achieve a good QBO. Important factors include, but are probably not limited to: resolution (horizontal and vertical), parameterised gravity wave forcing [Giorgetta *et al.*, 2006] and placement of model lid [Osprey *et al.*, 2013].

Within the recent big model comparison collaborations, the Coupled Model Intercomparison Project, phase 5 (CMIP5 - leading to the 4th IPCC Assessment report) and the Chemistry-Climate Model Validation Activity (CCMVal-2), only four out of more than thirty (CMIP5) and five out of fourteen models (CCMVal-2) internally produced a pattern of alternating easterly

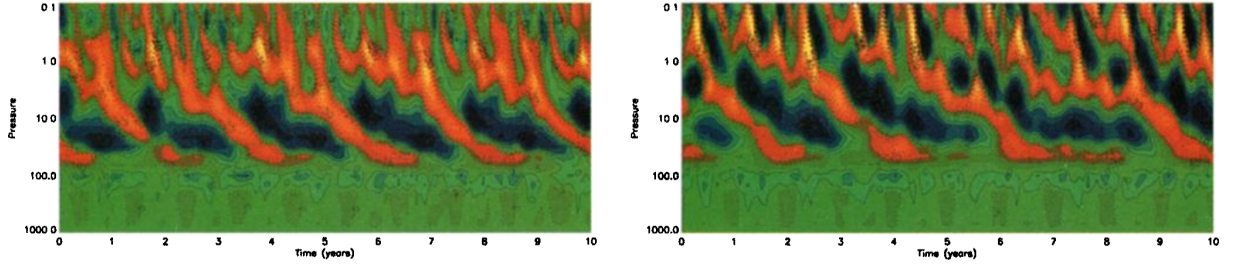


Figure II.10.: Equatorial zonal mean zonal wind from simulations of the Unified Model for the USSP parametrisation (left) and the DSP parametrisation (right). Red colours show westerly, green/blue colours easterly wind. Plot from [Scaife *et al.*, 2000].

and westerly shear zones in the stratosphere. To what extent can these indeed be called a “quasi-biennial oscillation”? Which features of the observed QBO do they represent? What common biases in the models are there and how do simulated oscillations compare to reanalyses and observations? These questions are the subject of chapter IV.

II.4. Possible QBO teleconnections

Biennial variations in surface observations were described as early as 1884, with Clayton [1884, 1885] noticing variations in surface pressure and rainfall at various well separated stations in North America between 1874 and 1883. He recognised a cycle of 11 years, that he assigned to the sun-spot cycle, and one of 25 months, for which he found no explanation. His observations were part of a larger inventory by Landsberg [1962], who put together a list of 43 timeseries for which an approximate 2-year cycle (between 1.9 and 2.7 years, with most of the periods slightly longer than two years) were identified. These cycles occurred in precipitation, snow cover, sea surface temperature, tree rings, floods and atmospheric pressure compiled from different locations. Landsberg already speculated on the question whether “these biennial pulses have anything to do with the recently observed remarkable wind shifts in the stratosphere”.

However, even if two processes happen on the same timescale and are found to be correlated (allowing for possible lags), they may not necessarily be connected. The oscillation in temperature seen in the tropical stratosphere is mainly caused by the dynamical QBO and can be explained adiabatically by the thermal wind relationship (section II.2). Similarly, oscillations in trace gases like ozone [Angell & Korshover, 1964] can be linked to the dynamical processes causing the QBO and the subsequent tropical upwelling due to the associated modulation of the secondary circulation. The further away from the tropical stratosphere, the harder it is to explain any supposed connection. One of the well-known examples illustrating this is the

modulation of the frequency of Sudden Stratospheric Warmings (SSWs) by different phases of the QBO; in observations from 1962 to 1977 more warmings are found to occur during a QBO easterly phase than in its westerly phase.

Holton & Tan [1980] suggested that the QBO acted as a wave guide for planetary scale waves, which cause SSWs. They further assert that the impact of the weakened polar vortex during an SSW then influences the surface temperature and pressure as well as geopotential height, which would explain the observed connection between the QBO and the high latitude near surface circulation. The process of wave propagation modulation by the QBO and its influence on the Northern Hemisphere polar vortex has become known as the Holton-Tan effect.

Despite numerous studies (c.f. Anstey & Shepherd [2014]), an adequate explanation of the exact process remains elusive and since the investigation of Holton & Tan [1980] (Figure II.11), the reported effect has varied in strength. Lu *et al.* [2008] showed changes in the strength of the correlation of the QBO and the polar vortex, which suggests other processes are at work. These could be for example the solar cycle [Labitzke *et al.*, 2006], the Pacific Decadal Oscillation [Jadin *et al.*, 2010], large scale circulation changes [Lu *et al.*, 2014] or another pathway of influence from the tropical stratosphere to the extratropical surface. The latter possibility is described by Simpson *et al.* [2009] as variation in the tropospheric eddy momentum flux following meridional circulation changes.

When looking at the teleconnection between the QBO and the Northern Hemisphere surface in winter, it is important to take into account other influences. One of the biggest factors is the El Niño-Southern Oscillation (ENSO), which is characterised by variations in sea surface temperature in the equatorial Pacific [Trenberth, 1997]. As with the QBO teleconnections, ENSO influences might be mediated via the troposphere or stratosphere [Butler *et al.*, 2014]. There are numerous studies looking at the modulation of ENSO influences by the QBO or the other way round [Wei *et al.*, 2007; Calvo *et al.*, 2009; Garfinkel & Hartmann, 2010; Richter *et al.*, 2015; Hansen *et al.*, 2016], but there might also be a direct interaction between the QBO and ENSO due to the influence of both on deep convection at the equator [Gray *et al.*, 1992; Schirber, 2015].

Another teleconnection, that has attracted the interest of researchers recently, is between the QBO and the Madden-Julian oscillation (MJO). The latter is an eastward propagating modulation of convection in the equatorial Indian and Pacific Oceans [Madden & Julian, 1971, 1972]. It can be seen in the tropical wave spectrum (Figure II.4), acting on timescales of 30-60

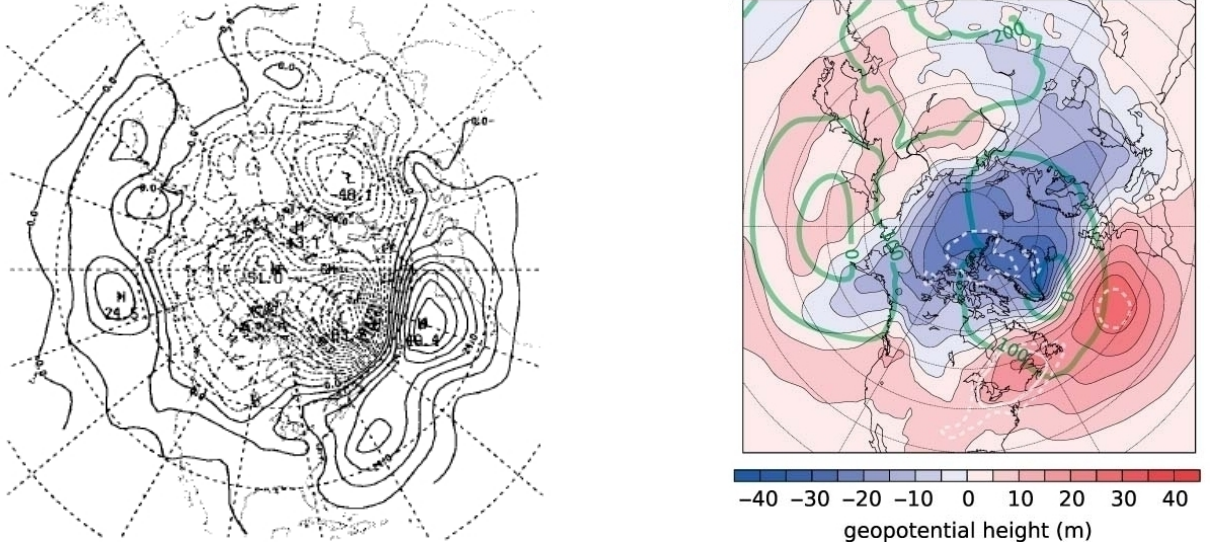


Figure II.11.: Left: Geopotential height difference between QBO Westerly and Easterly phase at 1000 hPa (surface) in the first paper suggesting an influence of the QBO on the extratropics [Holton & Tan, 1980]. Right: Updated analysis by Anstey & Shepherd [2014]. January-February mean of the geopotential height QBO W-E difference. Green contours show the mean climatology, white lines regions of significance (dashed 90%, solid 95%); ECMWF reanalysis 1957-2011. In both cases, the equatorial zonal wind at 50hPa has been used to define the QBO phases.

days, and is the most influential pattern of convective variability.

It has been pointed out by Liu *et al.* [2014] that the MJO can also be linked to different types of Sudden Stratospheric Warmings, with its influence being modulated by the QBO. However, the authors suggest further investigation of the exact mechanism. Yoo & Son [2016] find that the QBO is an important factor in predicting the December-February MJO. In light of the MJO itself influencing the upward fluxes of gravity waves from the troposphere into the stratosphere [Moss *et al.*, 2016], which are important for generating the QBO, the nature of the QBO-MJO interaction and their connection with the extratropics is not obvious.

Soon after the discovery of the QBO, a possible connection between the stratospheric wind and the Indian Monsoon has been pointed out (see Bhalme *et al.* [1987] and references therein). These early studies discuss a spectral peak in the rainfall pattern at QBO frequencies and examine the correlation of the stratospheric zonal wind with rainfall intensity. Despite significant correlations of the 10hPa wind of the preceeding January with the rainfall in June to September (Asian Monsoon season), Bhalme *et al.* [1987] also state that the QBO is unlikely to be the only influential factor, with ENSO playing an important role, but do not offer a mechanism of QBO→Monsoon influence. More recently, Read & Castrejón-Pita [2012] questioned the adequacy of using the QBO as Monsoon predictor, attributing the ostensible link to separate synchronisation of the QBO and the tropical biennial oscillation, which is associated with the

Monsoon, to a semi-annual cycle.

Since the Monsoon plays a crucial role for agriculture in India, improving its prediction is an important objective in modelling. With more data being available, Claud & Terray [2006] revisited the connection between the QBO and the Indian Monsoon. Their correlation and composite analysis of data from 1953-2001 confirm the modulation of the Monsoon by the QBO and the usefulness of including stratospheric equatorial wind in long-range forecasting, especially for the late Monsoon season (August and September). However, they also point out the importance of exploring the additional influence from ENSO, as do Liess & Geller [2012]. Their analysis, that included precipitation, wind and geopotential height fields, suggests that the modulation of the tropical tropopause height/temperature and deep convection by the QBO leads to changes in convection over India which then causes differences in precipitation. The simplified mechanism is as follows: QBO west phase $\rightarrow$ larger stratospheric equatorial temperatures compared to subtropics $\rightarrow$ warmer/lower tropopause along equator $\rightarrow$ enhanced convection over India $\rightarrow$ increased Monsoon precipitation. Inoue & Takahashi [2013] suggest that the QBO modulates the tropospheric circulation over Asia by influencing wave activity and convection in autumn, which then leads to changes in Monsoon rainfall.

In sections V.2.1 and V.2.2, composite and regression analyses are employed, as well as a recently developed method (chapter VII) based on causal effect networks [Runge *et al.*, 2014] to investigate the different proposed QBO teleconnections (section VII). With the knowledge of the mechanism of the interaction between tropical stratosphere and extratropical circulation, it might be possible to improve model predictions [Scaife *et al.*, 2014]; anticipating this, Clayton already hoped to “be able to foreknow the coming weather, with considerable accuracy for long periods in advance” [Clayton, 1884].

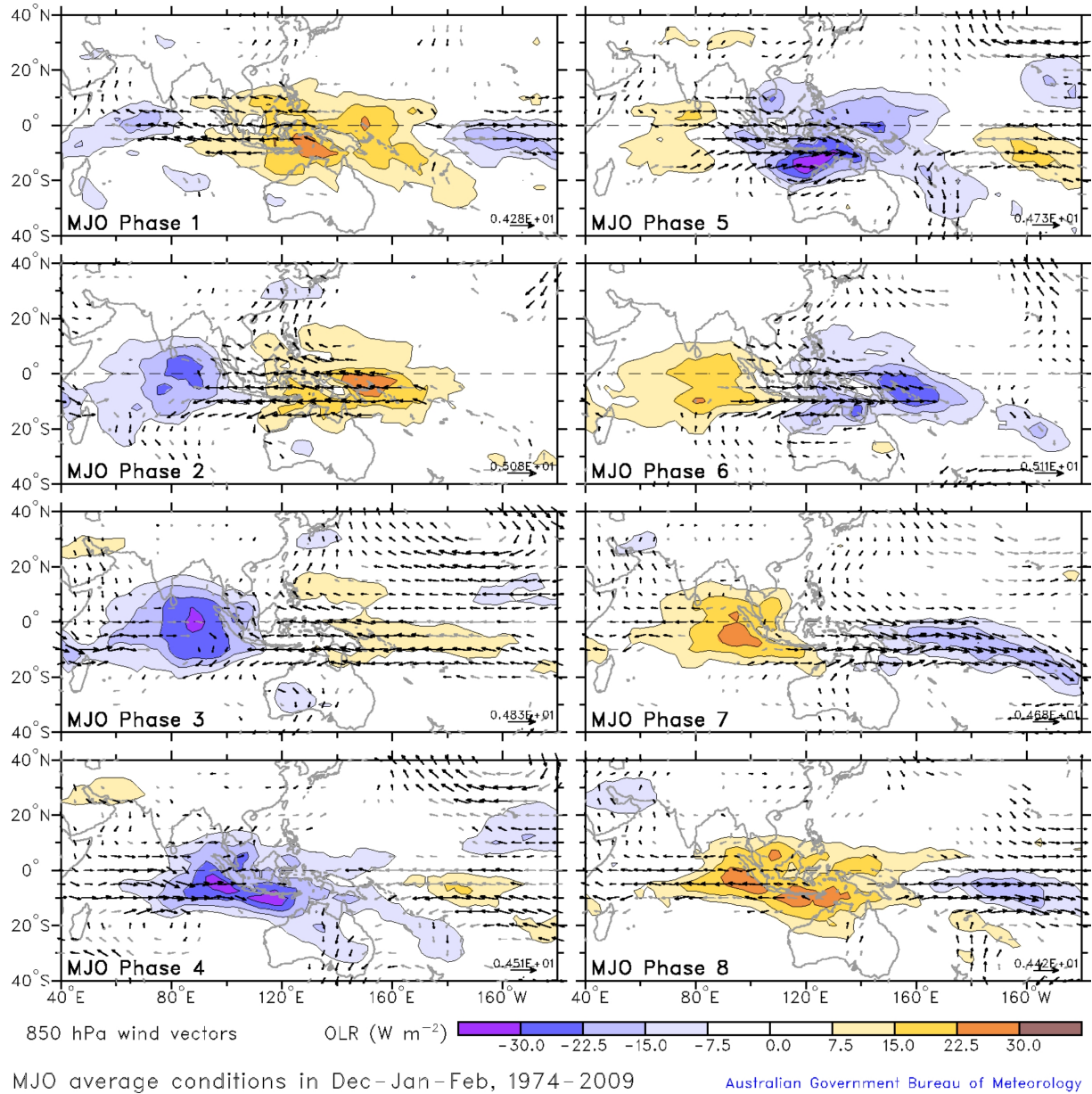


Figure II.12.: The wintertime (DJF) Madden-Julian Oscillation, average from 1974-2009. The anomaly of outgoing longwave radiation (OLR), in colours, is a measurement of cloudiness and thus convection. Blue colours show less OLR (stronger convection), yellow less active conditions. Vectors depict the anomaly in mean circulation at 850hPa. The classification into 8 phases depends on the center of the anomalies. Plot from Australian Government Bureau of Meteorology [2016].

III. Data and Methods

III.1. Data

III.1.1. Observations

The Quasi-Biennial-Oscillation (QBO) Data Series is used as the observational dataset of equatorial zonal wind [Freie Universität Berlin, 2016]. Data are available from 1953 onwards from three different stations: Canton Island, 3°S/172°W (Jan 1953 - Aug 1967), Gan/Maledive Islands, 1°S/73°E (Sep 1967 - Dec 1975) and Singapore, 1°N/104°E (since Jan 1976). The values for zonal wind, u , are provided as monthly means from daily radiosonde measurements at the pressure levels 70, 50, 40, 30, 20, 15, and 10 hPa, the latter since 1956. The dataset is updated regularly since the initial work by Naujokat [1986]. In this thesis, the monthly mean wind between 10 and 70hPa from 1956-2015 is used. Figure II.2 shows the data starting from 1953. The 100 hPa level, that is shown from 1957 onwards, is only available in the dataset from Singapore (1987 onwards). The QBO observational data will be used as a reference in the model comparison (Section IV.3), as far as possible, and to assess the surface impacts in the regression and composite analysis (Sections V.2.1, V.2.2).

For the latter, two datasets of surface parameters are analysed: mean sea level pressure from the HadSLP2 dataset [Allan & Ansell, 2006] and sea surface temperature from the HadISST [Rayner *et al.*, 2003] dataset [MetOffice, 2014]. HadSLP covers the time from 1850 to 2011, with data before 2004 originating from HadSLP2, and the more recent data from HadSLP2r. The dataset comprises 2228 land stations, with 615 providing data covering more than 100 years. The land based stations are complemented by marine observations from the International Comprehensive Ocean Atmosphere Data Set [Worley *et al.*, 2005] and the Marine Data Bank from the UK MetOffice.

Index	Time coverage	Source	Daily data
QBO	1956-2014	FU Berlin	ERA Interim
Solar (Sunspot number)	1700-2010	WDC-SILSO, Royal Observatory of Belgium, Brussels	-
Volcano	1850-2011	GISS Sato Index [Sato <i>et al.</i> , 1993]	-
North Atlantic Oscillation (NAO)	1950-2014	NOAA Center for Weather and Climate Prediction	NOAA
Arctic Oscillation (AO)	1950-2014	NOAA Center for Weather and Climate Prediction	NOAA
Antarctic Oscillation (AAO)	1979-2014	NOAA Center for Weather and Climate Prediction	-
Pacific North America Pattern (PNA)	1950-2014	NOAA Center for Weather and Climate Prediction	NOAA
Pacific Decadal Oscillation (PDO)	1854-2015	NOAA Centers for Environmental Information	NOAA
El Niño Southern Oscillation (ENSO)	1870-2012	HadISST, Nino 3.4 region (120W°-170°W and 5°N-5°S)	NOAA (from 1981 onwards)
Madden-Julian Oscillation (MJO)	1979-2013	NOAA Earth System Research Laboratory	NOAA
Central Indian Monsoon rainfall	1871-2015	Indian Institute of Tropical Meteorology	IPRC
Polar Vortex (upper stratosphere)	1979-2010	ERA Interim, 60°N-70°N, 5-50hPa	ERA Interim
Polar Vortex (lower stratosphere)	1979-2010	ERA Interim, 60°N-70°N, 100-200hPa	ERA Interim
Subtropical Jet	1979-2010	ERA Interim, 25°N-35°N, 300-150hPa	ERA Interim

Table III.1.: Overview of the time series used in regression and causal effect network analysis. Most of the analysis was done on monthly mean data, for which the sources are listed; no daily data of a solar or volcanic index was used. The jet indices were defined as mean of the zonal mean zonal wind in the specified regions using ERA-Interim data. The daily QBO index was derived from ERA-Interim as well since there is no easily available daily observational data. The daily ENSO index starts in 1981. The monthly central Indian rainfall as Monsoon index was taken from the Indian Institute of Tropical Meteorology (IITM), the daily index from the International Pacific Research Centre (IPRC). The NAM, AAO, AO, NAO and PNA indices are derived from EOF analysis of geopotential height anomalies. As MJO index, the OLR MJO index (OMI), based on outgoing longwave radiation (OLR) is used. For a more detailed description of the derivation of the various indices refer to the pages of the National Oceanic and Atmospheric Administration (NOAA, <https://www.esrl.noaa.gov/psd/data/climateindices/list/>). Monthly mean time series of the indices are shown in Figure III.1.

The data was quality controlled, gridded, interpolated and blended together to form the final dataset of monthly mean sea level pressure with a spatial resolution of 5° [Allan & Ansell, 2006]. HadISST combines observations of sea ice, sometimes digitized from hand-drawn charts, and in situ sea surface temperature measurements from ships (Marine Data Bank) as well as satellite measurements. Data undergo a similar process as the mean sea level pressure data, available as monthly mean sea surface temperature, 1871-2011 with 1° spacial resolution. A full account of the incorporated measurements and method is given in Rayner *et al.* [2003].

Various indices representing forcings and influenced phenomena are used in regression and causal effect network analyses. Table III.1 summarizes the origin and time span of daily and monthly datasets and Figure III.1 shows the time series of the monthly data that were used in at least one of the analyses.

Where indices for the subtropical jet and the polar jet are used, these are calculated as the mean of $\bar{u}$ from the ERA-Interim reanalysis in defined regions (details in Table III.1). The areas defining these indices are shown in Figure III.2. The subtropical jet area was chosen based on the mean and standard deviation of the zonal mean zonal wind field. For the stratospheric polar vortex, different indices are used in the literature, usually at or bounded around 60°N , 10hPa. This area was expanded to cover more of the vortex area; the lower polar vortex was defined to distinguish between stratospheric and tropospheric mediated QBO influences.

III.1.2. Global Climate Models

Zonal mean zonal wind $\bar{u}$ and temperature $\bar{T}$ are assessed for the models that internally produce a QBO, which were submitted to the Coupled Model Intercomparison Project, phase 5 [World Climate Research Programme, 2010] and the Chemistry-Climate Model Validation 2 [SPARC CCMVal, 2010]. An overview is given in Tables III.2, A.1 and A.2. These models include HadGEM2-CC (and its predecessor HadGEM1), MIROC-ESM(-CHEM), MPI-ESM-MR (CMIP5), EMAC, MRI, UMSLIMCAT, UMUKCA-METO and UMUKCA-UCAM (CCMVal). The historical CMIP5 **r1i1p** run is taken; timespans for the individual models are listed in table III.2. In the historical run, known forcings such as the solar cycle, volcanic eruptions, greenhouse gas concentration, CO_2 and other emissions, ozone distribution and land use, are included to get a most realistic reproduction of the past climate. Forcing from sub-grid scale gravity waves is included using two different parametrising schemes (Table III.2).

To assess the internal natural variability in HadGEM2-CC, which is also used for model studies,

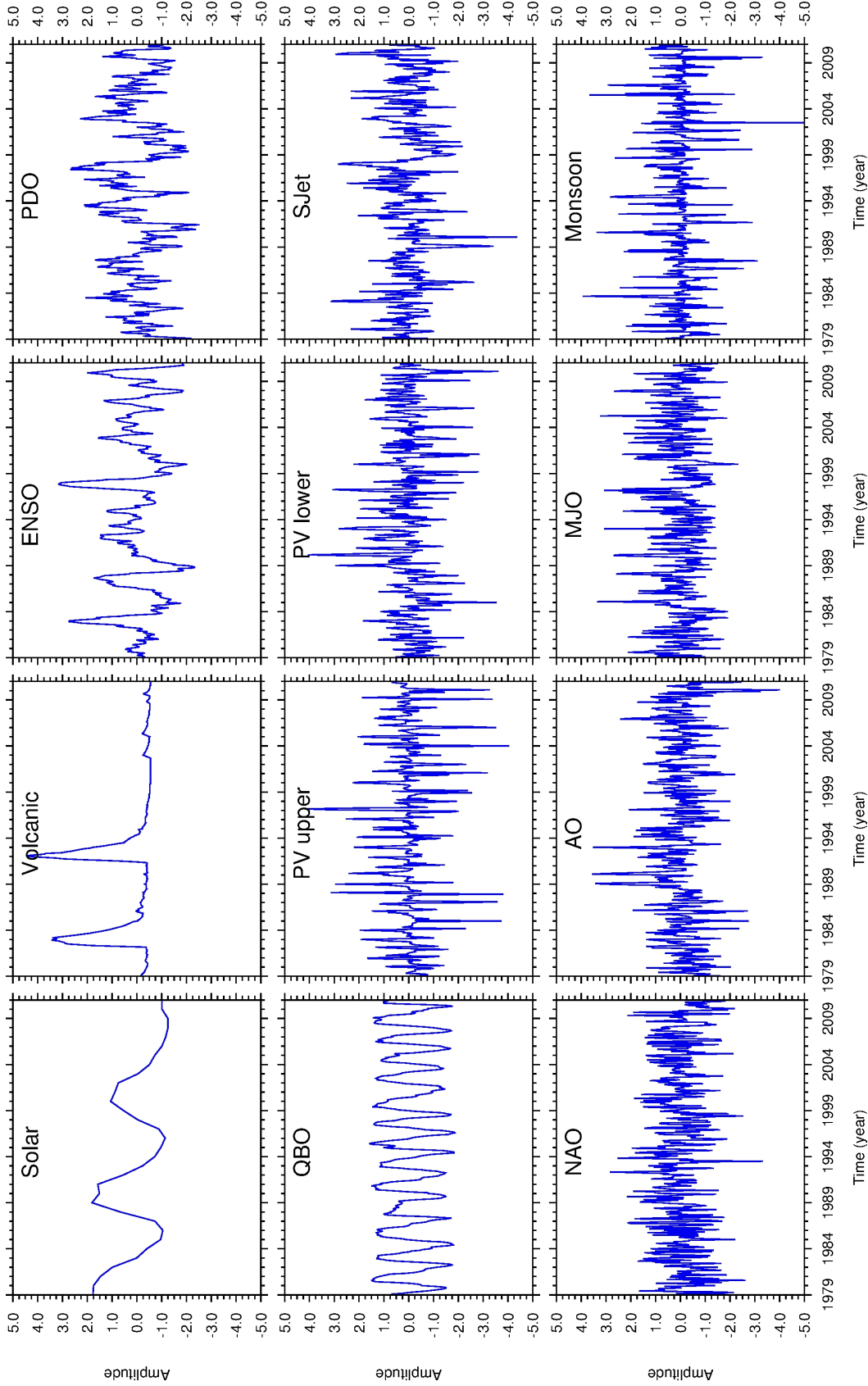


Figure III.1.: Monthly mean time series of climate indices used in regression and causal effect network analysis, as described in Table III.1. All time series are normalized for plotting. The QBO index shown is the first EOF of equatorial $\bar{u}$ between 10hPa and 70hPa. Apart from Solar and Volcanic forcing indices, time series are deseasonalized and a linear trend removed.

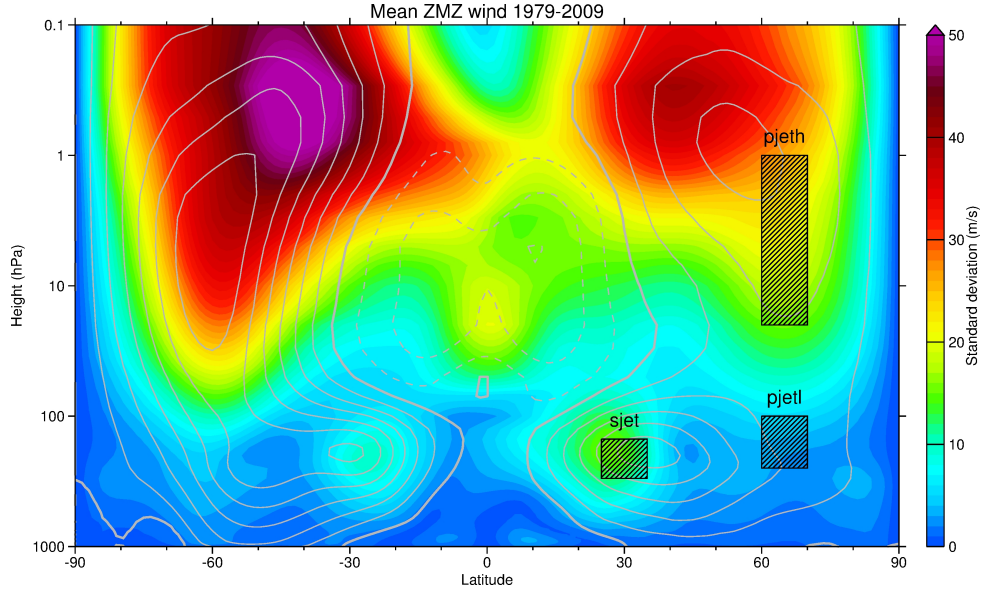


Figure III.2.: Latitude-height plot of areas from which the jet indices are calculated. Grey contours show the mean of $\bar{u}$ from 1979 to 2009 from the ERA-Interim reanalysis, spaced at 10 m/s, negative contours are dashed and the zero contour is thicker. Colours show the standard deviation of the wind field in m/s. The areas chosen for the subtropical jet (sjet) and the upper/lower polar vortex indices (pjeth, pjetl) are highlighted.

500 years of the control run (MetUM experiments `akgie`, `alsza` and `alszb`) will be analysed. The sea surface temperature (SST), mean sea level pressure (MSLP) and zonal wind u fields are used for this purpose. In contrast to the historical runs, the control run does not include varying “external” forcings mentioned above; terms like solar irradiance are instead kept constant, CO_2 concentration is fixed at the pre-industrial level.

III.1.3. Reanalyses

Reanalyses combine measurement and climate model data. In contrast to observations, which are local, reanalyses data cover the whole globe, and a large (vertical) part of the atmosphere as well. Table III.3 lists the datasets assessed in the QBOi project study (Section IV.1) and their main references. Monthly mean zonal mean wind and temperature from 1979-2009 (except ERA40, covering 1958-2001) are analysed in the same way as the GCMs to provide a reference of the models’ performance regarding QBO simulation.

To a certain extent reanalyses can serve as reference for the real world as the data is produced by combining output from a climate model run in hindcast mode (i.e. forecasting of past events) with real measurements of meteorological variables. Assimilating the different observations into a climate model has the advantage of yielding a global dataset that is physically consistent (instead of just being a numerical interpolation). However, this method can also include some

Model	Reference	Resolution	Wave parametrisation	Length
MIROC-ESM-CHEM	Watanabe <i>et al.</i> [2011], Watanabe & Kawatani [2012]	T42 L68	Hines	156 years
MPI-ESM-MR	Schmidt <i>et al.</i> [2013], Krismer & Giorgetta [2014]	T63 L95	Hines	156 years
HadGEM2-CC	Osprey <i>et al.</i> [2013]	1.25° x 1.875° L60	W & M	374 years
HadGEM1	Osprey <i>et al.</i> [2010]	2.50° x 3.75° L60	W & M	50 years
CMCC-CMS	Manzini <i>et al.</i> [2006], Giorgetta <i>et al.</i> [2006]	T63 L95	Hines	156 years
EMAC	Jöckel <i>et al.</i> [2006]	T42 L90	Hines	41 years
MRI	Shibata & Deushi [2008a], Shibata & Deushi [2008b]	T42 L68	Hines	47 years
UMSLIMCAT	Tian & Chipperfield [2005]	2.50° x 3.75° L64	W & M	55 years
UMUKCA-METO	Morgenstern <i>et al.</i> [2009]	2.50° x 3.75° CP60	W & M	47 years
UMUKCA-UCAM	Morgenstern <i>et al.</i> [2009]	2.50° x 3.75° CP60	W & M	45 years

Table III.2.: Overview of the climate models used in the study. For more details on the models and the modeling institutions see Tables A.1 and A.2. The wave parametrizations used were either the DSP scheme [Hines, 1997a,b] (Hines) or the USSP scheme [Warner & McIntyre, 2001] (W & M).

Reanalysis	Reference	Resolution of forecast model
ERA40	Uppala <i>et al.</i> [2005]	T _L 159 and N80 reduced Gaussian, L60
ERA Interim	Uppala <i>et al.</i> [2005]	T _L 255 and N128 reduced Gaussian, L60
MERRA	Rienecker <i>et al.</i> [2011]	0.66° lon x 0.5° lat; 72 sigma levels
JRA25	Onogi <i>et al.</i> [2007]	T106 L40
JRA55	Ebita <i>et al.</i> [2001]	T _L 319 L60
CFSR	Saha <i>et al.</i> [2010]	T382 L64
NCEP1	Kalnay <i>et al.</i> [1996] Kistler <i>et al.</i> [2001]	T62 L28
NCEP2	Kanamitsu <i>et al.</i> [2002]	T62 L28

Table III.3.: Overview of the reanalysis datasets used in the study. The period is 1979-2009 for all reanalyses except ERA40, which covers 1958-2001.

error sources, like the underlying observational error and gaps in coverage (spatial and temporal), shortcomings of the climate model used, unintentional errors from subjective decisions (e.g. which observations to use) and a background error arising from using the analysed state as new input for the subsequent forecast. These caveats, as well as some suggestions on how to overcome them, are described in general in Thorne & Vose [2010], and for the tropical stratosphere in particular in Fujiwara *et al.* [2016].

III.2. Methods

III.2.1. Composite analysis

One common method to look for signals in environmental data is to compare the means of fields (in the atmospheric sciences these would be variables like wind speed, temperature, pressure etc.) in well-defined states. This could include comparing the mean state of single months or different seasons. More common, however, is looking at the state of a phenomenon such as the QBO, sorting the data into categories before comparing their respective means. In the case of the QBO, the natural choice is contrasting “East” and “West” years, taking $\bar{u}$ at a certain height (usually 30 or 50 hPa) as the selection parameter. Having two distinct states, significant differences can be identified using a two-sided Student-*t*-test. Alternatively, significance can be assessed comparing the mean state of the selected sample to the underlying distribution of states using a Monte-Carlo test (bootstrapping).

III.2.2. Regression analysis

Multilinear regression is used to model the influence of several explanatory variables x_i on the measured variable y . In the case of this study, x_i and y are functions of time and thus the relation can be written as $y(t) = \sum_{i=1}^n \beta_i x_i(t) + \epsilon(t)$ with regression coefficients β_i and the residual $\epsilon(t)$. The “best fit” for the coefficients $\vec{\beta}$ is calculated by minimizing $|\epsilon|^2$ (which is the difference between the measured value y and the regression model $\sum_i \beta_i x_i$ at each point in time. This regression is called “linear” because $y(t)$ only depends linearly on the coefficient vector $\vec{\beta}$.

Testing the significance σ_i of β_i is done by a one-sided Student-*t*-test with the null-hypothesis of β_i equaling 0. In the simplest case of only 1 explanatory variable x , the regression is equal to a linear fit $y = \beta x + \epsilon$, where β describes the slope of the line (equal to the correlation of x and y) and ϵ is the vertical distance between the measured point y and the fit βx at each single

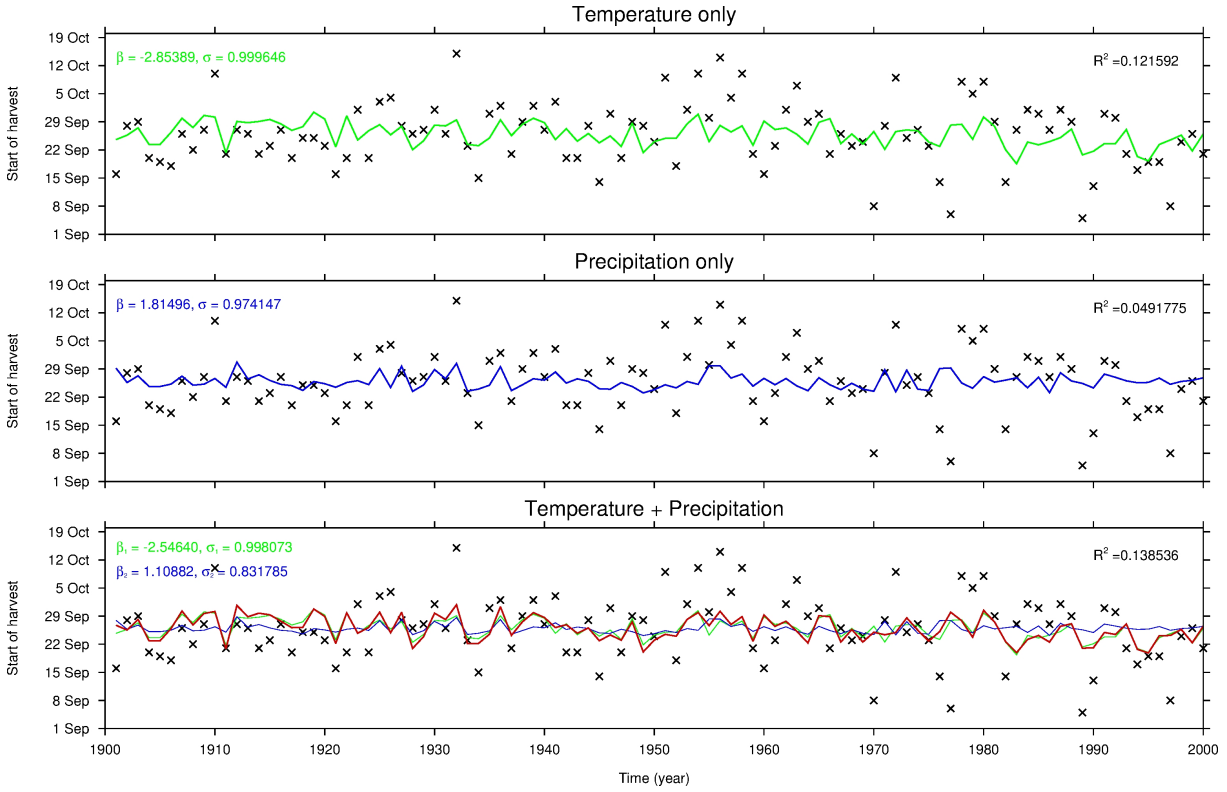


Figure III.3.: Examples of linear regression analysis: The upper and middle panel show $\beta_i x_i(t)$ from a model of 1 explanatory variable (coloured line) and the lower a model of 2 explanatory variables (green: temperature, blue: precipitation, red: sum of both) which is fitted to the same dataset (black crosses). The values for the regression coefficients β_i are given as well as their significance σ_i according to a t -test. The R^2 values state the amount of variance accounted for by each model.

point. To assess the quality of the fit, the coefficient of determination R^2 can be used. It is defined as $R^2 = 1 - \frac{|\epsilon|^2}{|\Delta_y(t)|^2}$ with $\Delta_y(t) = y(t) - \bar{y}$ representing the variation in the measured variable y ($\bar{y}$ denoting its mean). R^2 thus compares the variability of both the observations and the regression model; by construction $0 \leq R^2 \leq 1$ with $R^2 = 1$ describing a perfect regression model.

Figure III.3 exemplifies this method using the start of the grape harvest in the Bordeaux region from 1901 to 2000 [Conseil des Grand Crus Classés en 1855, 2008] with explanatory variables being the monthly mean temperature and the monthly mean regional precipitation in July at 0° longitude, 45° latitude. Precipitation and temperature data is taken from National Center for Atmospheric Research Staff (Eds.) [2014]. Three regression analyses are carried out, two using only one of the explanatory variables and one using both.

The interpretation of this analysis is that higher July temperatures lead to an earlier harvest date (negative β), whereas higher precipitation leads to a later harvest date (positive β). In both cases, the regression coefficients are significantly different from 0 on the 95% level, in case of temperature even on the 99% level. The model incorporating both predictors is better than using only a single predictor (higher R^2 value), however, the amount of variance accounted for by these two variables is only about 14% of the total variance, so a more refined model would be needed to make better predictions of the harvest date.

When comparing the influence of several explanatory variables x_i on y , all time series are normalised (zero mean and amplitude of one standard deviation). The calculated regression coefficients β_i are then scaled again by multiplication of the ratio of the amplitude of the variable and its standard deviation $\beta'_i = \frac{x_{max} - x_{min}}{\sigma(x_i)}$. Hence, β'_i accounts for the amplitude of y attributed to variable x_i , which makes it easier to compare influences of different forcings and to the composite analysis.

The measurement variables $y(t)$ examined in this study are wind (u, v, w), SST and MSLP as well as zonal means of u , the residual circulation ($\bar{v}^*, \bar{w}^*$) and the Eliassen-Palm (EP) flux (F_ϕ, F_z). The explanatory variables are the solar cycle (total irradiance), the Nino 3.4 index, volcanoes (i.e. aerosol optical depth), rising CO_2 levels (i.e. a linear trend) and a QBO index that is either $\bar{u}$ at a certain height or an index calculated from an EOF analysis. A time series from 1979-2010 of the analysis indices is shown in Figure III.1.

III.2.3. Causal effect networks

One problem with the two methods described above is that they cannot distinguish correlation from causality or detect indirect chains of influence (where one process influences another via modulation of a third process) and limitations in assessing nonlinear systems [Sugihara *et al.*, 2012]. Lagged correlation/compositing attempts to address the problem of establishing the directionality of the influence, however these methods can produce misleading results when strongly autocorrelated time series are present [Runge *et al.*, 2014]. Thus, a new method to assess strength, direction and the time lag of interactions has been employed. Ebert-Uphoff & Deng [2012b,a] suggested graphical models to address these difficulties, and Runge *et al.* [2012a,b, 2014] further developed these ideas, implementing the algorithms in the software package “TiGraMITE” [Runge, J., 2015].

The underlying principle of causal effect networks (CEN) is to distinguish between direct links $X \rightarrow Y$, indirect links $X \rightarrow Z \rightarrow Y$ and a common driver Z of both X and Y using conditional independence. This means testing the processes X and Y for conditional independence, i.e. whether $P(X \cap Y|Z) = P(X|Z)P(Y|Z)$ holds for the probability function P . The algorithm used for the latter consists of several steps as shown in Figure III.4: First, all links between the variables X , Y and Z are tested up to a lag time τ_{max} , that can be set, for the existence of an unconditional link. Links that have been found to be significant (yielding a nonzero cross correlation coefficient that is significant to a set level) are tested for their conditional dependence using a modified Peter-Clark (PC) algorithm as described in Runge *et al.* [2014].

As climate data brings a time ordering, the direction of links at $\tau > 0$ is clear, leading to the identification of “parents”. Links at $\tau = 0$ are classified as “neighbours” and their direction might be resolved on a shorter timescale. All links are then assessed for their conditionality on a combination of the possible parent processes, testing for a nonzero partial correlation; the latter is defined as the cross correlation between the residuals in a regression onto the same regressors (ϵ and $\vec{x}$ from the description in the previous section). This means testing for the existence of a link $X_{t-\tau} \rightarrow Y_t$ excluding influences of X at other lags τ' , autocorrelation effects $Y_{t-\tau'} \rightarrow Y_t$ and third party influences $Z \rightarrow Y$.

In the next step, all identified conditionally independent links are assessed for their strength and significance in a multiple linear regression model. The explanatory variables $\vec{x}$ are the parents of a process that were identified in the step above, the regression coefficients $\vec{\beta}$ represent the strength of the link. The regression is done with each of the time series X , Y and Z and the re-

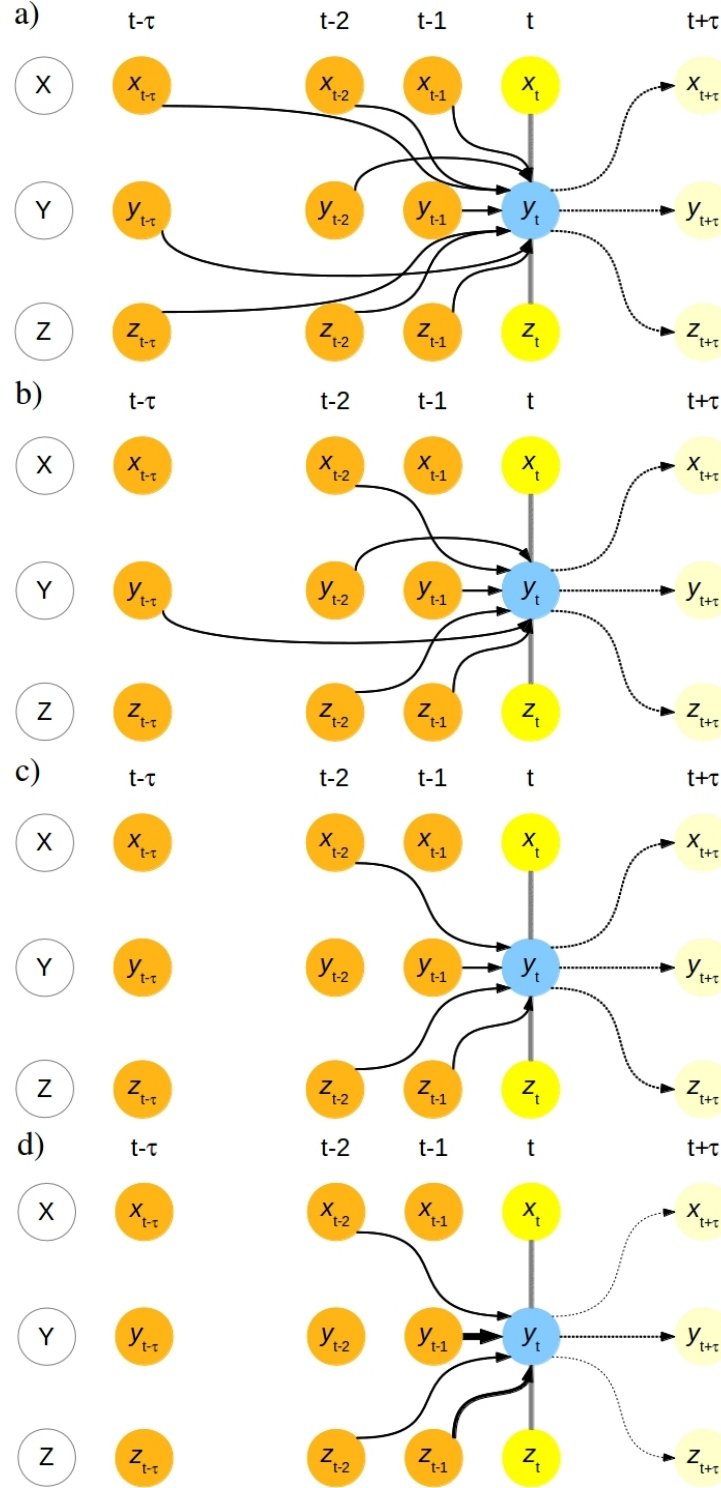
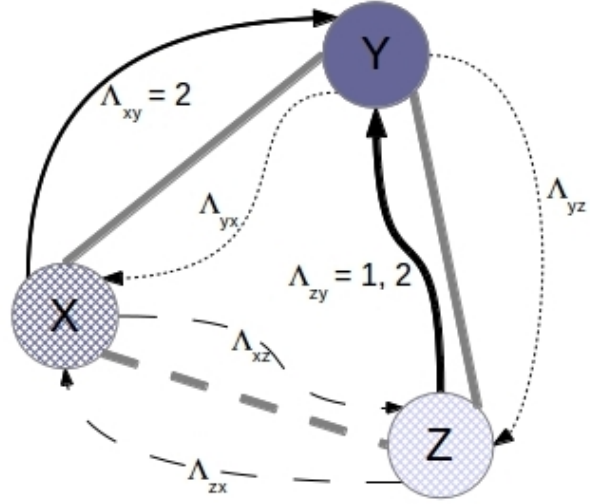


Figure III.4.: Steps of analysis using the Causal Effect Networks software package TiGraMITE: a) Example of connections from time series X , Y and Z to y_t . Only states at lags $\tau > 1$ can have a direct influence and are thus classified as “parents”. There might be a connection to the “neighbours” of x_t and z_t to y_t that can potentially show up as a directional link on higher time resolution. The maximum lag τ to be taken into account can be varied. All possible links are tested for the strength of their cross correlation. b) Possible links obtained by a modified PC algorithm, testing for a nonzero cross correlation between the time series at defined lags τ : Autocorrelation of Y over a time range of $\tau \geq 2$, links from X to Y with lags 1 and 2 and links from Z to Y at various lags. c) Connections of the time series after testing the proposed links for independence. d) Links that have been identified as significant and conditionally independent: Strong autocorrelation of Y , influence X and Y with lag $\Lambda_{xy} = 2$ and influence Z on Y with $\Lambda_{zy} = 1$ and $\Lambda_{zy} = 2$ with the former being stronger than the latter. Links between $z_{t-\tau}$ and x_{t-1} to y_t were classified as conditional and thus removed; this can be due to autocorrelation of the X and Z .

Figure III.5: Display of the connections from the time series of Figure III.4 d) in a network graph. The autocorrelation of the processes is depicted by the colour of the circles, the strength of the most significant links by the thickness of the line; associated lags Λ_{ij} are given. The dashed links/hatched colours and Λ values are not shown in the panels of Figure III.4 (for clarity), but are calculated in the process. There might be more than one link that fulfils the significance criteria, in which case the corresponding lags are displayed in order of link strength, here $\Lambda_{zy} = 1, 2$.



spective parent processes; the links, their strength and significance can be depicted in a network graph, as exemplified in Figure III.5. There are different methods to establish links and test for their independence. The basic one uses linear methods, giving the advantages of defining and interpreting a significance level. However, the causal effect network can also be used in conjunction with nonlinear methods from information theory, namely momentary information transfer, as detailed in Runge *et al.* [2012a,b, 2014].

III.2.4. Fourier analysis

Fourier analysis is a common method for decomposing a time series into periodic signals of different frequency, phase and amplitude. The basis is the Fourier transformation that transfers an integrable function $f(t) : \mathbb{R} \rightarrow \mathbb{C}$ from the time domain into the frequency domain: $\mathcal{F}(f(t)) = \hat{f}(\omega) = \int_{-\infty}^{\infty} f(t)e^{i\omega t} dt$ (This notation uses the angular frequency ω , other conventions use frequency, $\xi = \frac{\omega}{2\pi}$, which leads to an integration factor of $\frac{1}{2\pi}$).

Real signal analysis, however, works slightly differently, as the input $f(t)$ is sampled at discrete time steps ΔT and only defined on an interval $T = N\Delta T$. The vector $f(t)$, which consists of the values of the measured variable at each point in time t_n , is transformed into a vector of N elements: $\mathcal{F}_k(f(t_n)) = \sum_{n=0}^{N-1} f(t_n)e^{-2\pi i k \frac{n}{N}}$ with $k \in 0, \dots, N-1$. $\vec{\mathcal{F}}$ represents the Fourier amplitudes at distinct angular frequencies $\omega_n = 2\pi i \frac{n}{N}$ with n an integer, $|n| \leq \frac{N}{2}$. The maximum detectable frequency (Nyquist-Frequency) in the Fourier spectrum corresponds to the sampling interval as $f_{max} = \frac{2}{\Delta T}$.

An example of the principles of Fourier analysis is shown in Figure III.6: A signal, composed of 20 waves with different (random) amplitudes $5 \leq A < 10$ and frequencies $\omega = n$,

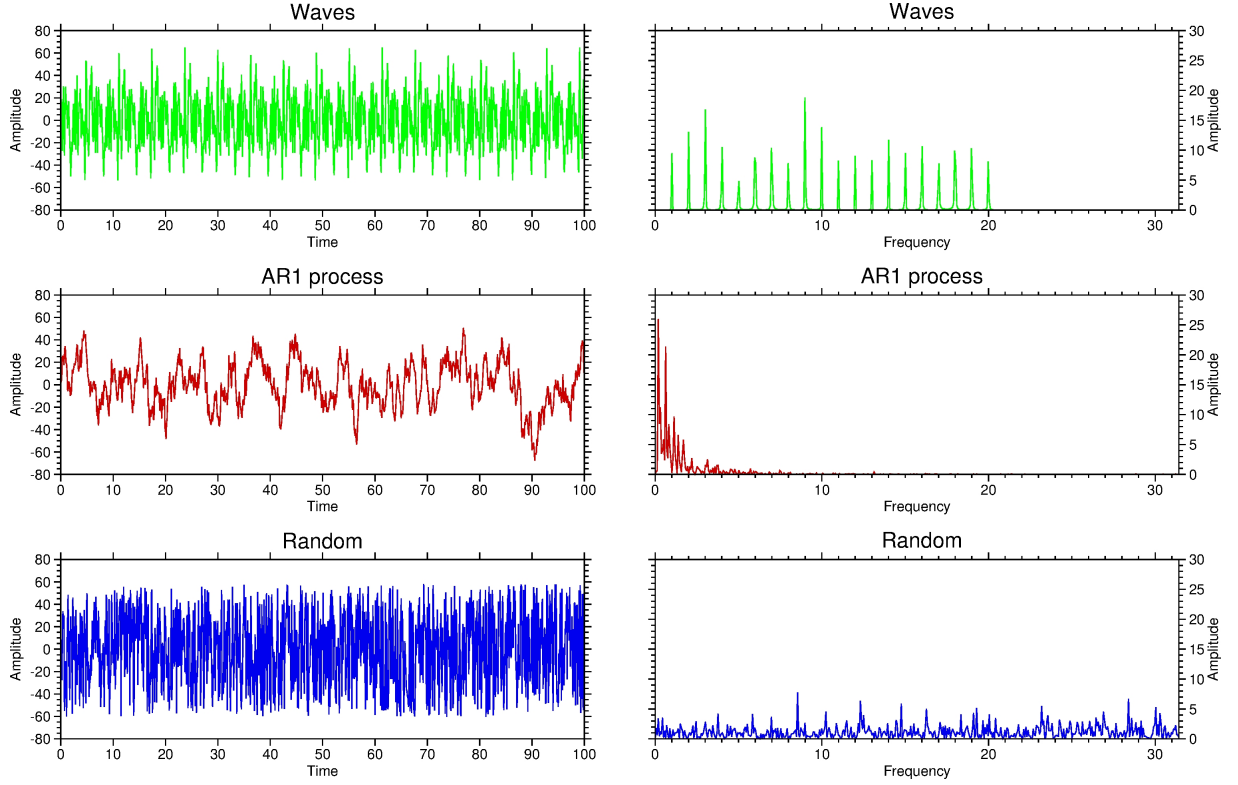


Figure III.6.: Examples of the effect of a Fourier transformation on three different time series of length $T = 1000 \cdot \Delta T$ (left) and the corresponding Fourier spectrum $|\mathcal{F}(f(t))|^2$ (right) as calculated via a Fast Fourier Transform (FFT).

First row: A sum of 20 waves $f(t) = \sum_{n=1}^{20} A \sin(nt + \varphi)$ with random amplitudes $5 \leq A < 10$ and random phase shifts $0 \leq \varphi < 2\pi$.

Second row: An autoregressive process of order 1 (AR1), also known as random walk.

Third row: 1000 random numbers from a uniform distribution (white noise/AR0 process).

All functions were shifted to have a mean of 0 and the same peak to peak amplitude for better comparability. Note the different scale for the Fourier amplitudes $|\mathcal{F}|^2$ compared to the amplitudes $f(t)$.

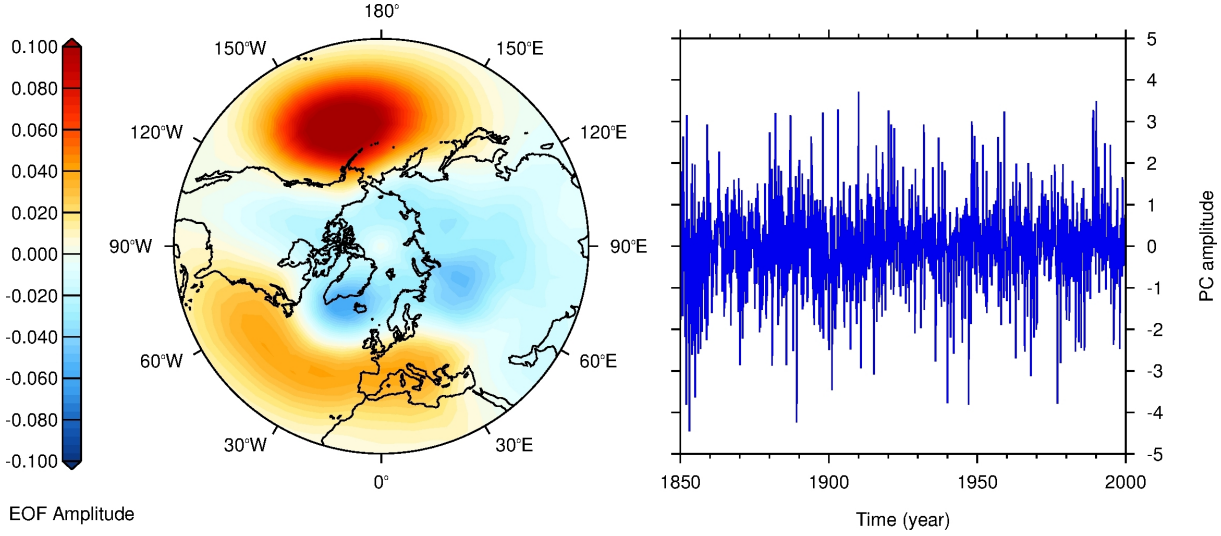


Figure III.7.: First EOF of monthly mean MSLP anomalies in the Northern Hemisphere (20°-90°), calculated from the HadSLP monthly mean MSLP anomaly, and its associated PC time series. The first EOF is usually identified as the Northern Annular Mode (NAM) or the Arctic Oscillation (AO).

$f(t) = \sum_{n=1}^{20} A \sin(nt + \varphi)$, is plotted as a function of time t , being sampled at $\Delta T = 0.1$, $t_{k=0}^{999} = k\Delta T$, and its Fourier transform as a function of frequency. The latter shows sharp peaks (which would be δ -functions in continuous time series) at the corresponding frequencies of the waves $\nu_n = \frac{n}{2\pi}$. As comparison, the Fourier transforms of white noise (AR0 process) and a random walk (AR1 process) are shown as well. The AR1 process consists of a series of points $x_{i=0}^{999}$ with $x_n = \phi \cdot x_{n-1} + \epsilon_n$. The autoregressive parameter $\phi = 0.9$ was set, the “error” values ϵ_n were drawn from a normal distribution ($\mu = 0$, $\sigma = 1$).

The Fourier power spectra show very characteristic shapes for the different time series: Distinct peaks at certain frequencies (combination of waves), a Lorentzian shape (AR1 process) and a uniform distribution (random input). The maximum detectable frequency is $\omega_{max} = 2\pi f_{max} = \frac{2\pi}{2\Delta T}$, in this case ($\Delta T = 0.1$) $\omega_{Nyquist} = 10\pi$. A combination of these patterns (as well as higher order autoregressive processes) might be found in atmospheric data and produce quite characteristic spectra.

This type of analysis is useful in the atmospheric sciences to extract periodic signals, like the diurnal or the annual cycle. In this study, the Fourier analysis is used to assess the spectra of equatorial wind and temperature, searching for an approximate biennial cycle and elaborating its spatial structure.

III.2.5. EOF analysis

The decomposition of a time-dependent function f , defined on various gridpoints, like height, into so called empirical orthogonal functions (EOFs) is similar to that of Fourier analysis. Instead of prescribing the basis functions as waves ($\sin(\omega t)$ and $\cos(\omega t)$), these functions are determined from the field itself, hence are called “empirical”. This can be useful for detecting larger scale patterns in time and space that are not necessarily sinusoidal waves.

The basic principle is to decompose f into spatial and temporal patterns: Let f be defined on N points in time and M gridpoints. Then the anomalies of f , $f' = f - \bar{f}_t$, which have zero temporal mean at each spatial point, can be written into a $N \times M$ matrix F . The eigenvectors $\vec{e}_n$ of the covariance matrix $C = \frac{1}{N-1} F^T F$, for which $C\vec{e}_n = \lambda_n \vec{e}_n$, are the spatial EOFs. They are sorted by descending values of λ_n , so that the first EOF accounts for most of the variance in the dataset. To each vector, there is a corresponding principal component (PC) time series, that is calculated as $\vec{p}_n = F\vec{e}_n$. The anomaly field F can be reconstructed as $F = P^T(t)\vec{\lambda}^T \mathbb{I}_M E$ with $P(t)$ being a $N \times M$ matrix, with rows $\vec{p}_n$, $\vec{\lambda}$ the vector of the eigenvalues λ_n and E being an $M \times M$ matrix with EOFs $\vec{e}_n$ as rows.

By construction, the M vectors e_n are normalised and orthogonal, hence serve as a basis in the spatial domain. However, if the separation between two eigenvalues is less than the uncertainty $\delta\lambda_n \approx \lambda_n(\frac{2}{n})^{\frac{1}{2}}$, then the two corresponding eigenvectors are degenerate [North *et al.*, 1982] and any linear combination of the two would be equally valid for forming the subspace of these eigenvectors.

For a lot of phenomena, the first EOFs (that account for most of the variability) are used as approximation for future climate projections or to derive climate indices, like the North Atlantic Oscillation (NAO) or the Northern and Southern Annular Mode (NAM¹/SAM). Figure III.7 shows the first EOF of Northern Hemisphere MSLP anomalies, with a consistent pressure anomaly over the pole and one of opposite sign mostly surrounding it; EOF 1 accounts for 16.9% of the variance. The shown pattern with negative anomalies over the pole would be referred to as a positive NAM/AO. A related index, the North Atlantic Oscillation (NAO) is calculated as the difference in pressure over Reykjavic and Lisbon, which are close to the peaks of the first EOF. Alternatively, the NAO index can be derived from rotated principal component analysis, which is a variation of EOF analysis [Barnston & Livezey, 1987].

In terms of the QBO, the first two EOFs capture most of the variance of the raw zms wind

¹Also referred to as the Arctic Oscillation (AO)

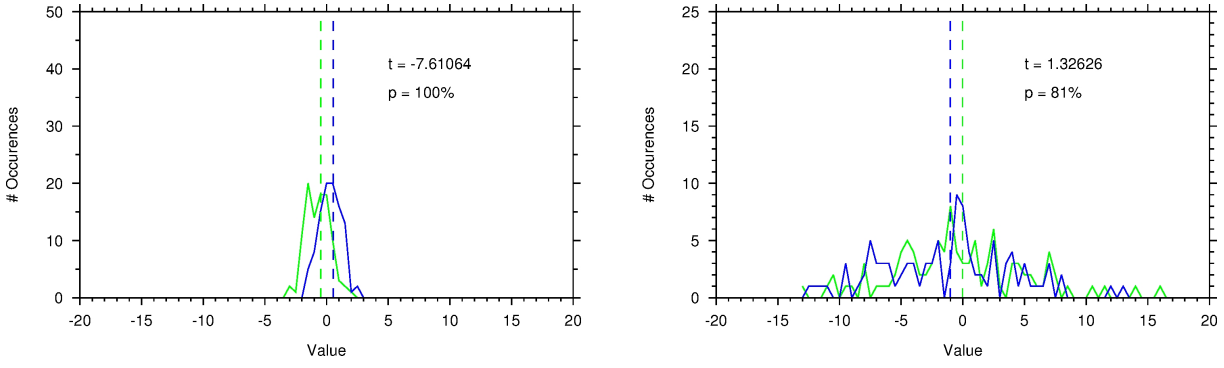


Figure III.8.: Examples for 2-sided Student’s- t tests on 2 Gaussian distributions ($n_i=100$ for all) with respective means (dashed lines). Left: small standard deviation of each distribution ($\sigma_1 = 0.99$, $\sigma_2 = 0.88$), right: high standard deviation ($\sigma_1 = 5.35$, $\sigma_2 = 5.25$) For each plot, the t -values as well as the associated p -values (in % as measure of significance) are given.

(71%) and can be used to derive a couple of characteristic parameters (Section IV.4), and serve as an index of a regression analysis (Section V.2.2).

III.2.6. Statistical tests

When performing correlation, composite or regression analysis it is vital to not only identify the strength of a connection (e.g. regression coefficient), but also the robustness of it. The latter is done using a significance test. There are many ways to test significance; the most commonly used ones in atmospheric science (and this thesis) are the Student’s- t test and “bootstrapping”, which is also referred to as Monte-Carlo testing. An extensive explanation of these methods is given in Wilks [2011].

Student’s t -test

There are two types of Student’s t -test: one- and two-sided. The most commonly used is a one-sided t -test in which the null hypothesis that a sample with mean $\bar{x}$ has been drawn from a population with a specified mean μ_0 . For large enough samples and a set of independent datapoints to draw from, the test statistic is $t = \frac{\bar{x} - \mu_0}{\sqrt{Var(\bar{x})}}$ with the variance of the sampling distribution of n observations in the denominator being estimated as $Var(\bar{x}) = \frac{\sigma^2}{n}$, where $\sigma = \sqrt{\frac{1}{n-1} \sum_{k=0}^n (x_k - \bar{x})^2}$ is the square root of the sample variance. The significance is then estimated from the calculated value t by comparing it against the Student’s- t -distribution with $n - 1$ degrees of freedom and a p -value is assigned; small values (usually $p < 0.05$ (95% level) or $p < 0.01$ (99% level)) leading to rejection of the null hypothesis. In the case of regression analysis, the significance of the value of coefficient β_i is tested against the null hypothesis $\beta_i = 0$.

Two-sided t -tests aim to compare whether two sample distributions are significantly different. Here $t = \frac{\bar{x}_1 - \bar{x}_2}{\sigma_{12} \sqrt{\frac{1}{n_1} + \frac{1}{n_2}}}$ and $\sigma_{12} = \sqrt{\frac{(n_1-1)\sigma_1^2 + (n_2-1)\sigma_2^2}{n_1+n_2-2}}$. n_i are the number of points in each sample, $\bar{x}_i$ denote the sample means and σ_i the sample standard deviations as above. This test takes into account the difference in the mean of the samples, as well as their combined standard deviation. The comparison of the means is not enough if the two distributions are broad, as illustrated in Figure III.8. Two Gaussian distributions (with low and high σ) are compared under the assumption of the sample variances being equal. The distributions have in both cases the same difference of the means $\bar{x}_1 - \bar{x}_2$, but have a different combined standard deviation σ_{12} .

While the difference of the mean in the case of a small sample variance is significant, in the case of a high sample variance the hypothesis of the distribution means being different would be rejected on the 95% level. Note that the test statistic t depends on the difference in means, the standard deviation and the number of points in each sample, so that in the second case of large standard deviations, the means might still be significantly different, if the sample size was much larger.

In the context of atmospheric science, the two-sided t -test is often used to compare data selected by one of two states, e.g. positive/negative sign of the NAM, the sign of the equatorial zonal mean zonal wind at a certain height or a positive/negative zonal wind anomaly of the polar vortex.

Bootstrapping

While the t -test is a so called parametric test (assuming an underlying distribution), there are non-parametric methods to test for significance as well. One of those is the Monte Carlo, or bootstrapping test. It is applied when the underlying distribution is not known or does not fulfil requirements for a parametric test, or if the sample sizes are not big enough for the central limit theorem to apply. The underlying null hypothesis states that a sample selected applying certain criteria does not differ from a random sample drawn from the same dataset. Let the size of the selected sample be n_S ; the question becomes whether its mean $\bar{x}_S$ is significantly different from the mean of a randomly drawn sample.

For the bootstrap sample, there are n_S random datapoints selected from the basic distribution, with repetition of points possible (bootstrapping with replacement). Then, the mean of this sample is one of the entries of the bootstrap distribution. This process is repeated up to the desired number of bootstrap samples, n_B . Often the value $n_B = 10,000$ is chosen, to give an

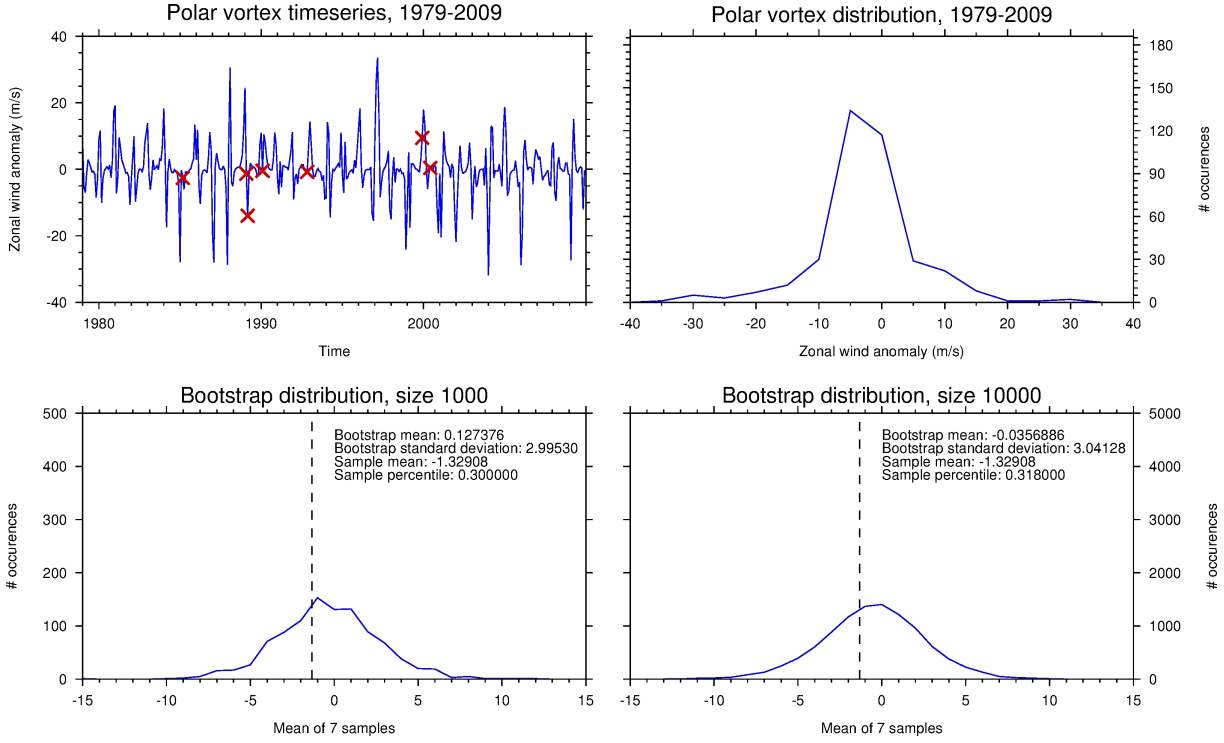


Figure III.9.: Example of a bootstrap test. Upper left: Time series of monthly polar vortex anomalies (60°N, 10 hPa) and selected times of strong subtropical jet anomalies (red crosses). Upper right: Distribution of polar vortex $\bar{u}$ anomalies. Lower: Distribution of bootstrap samples, including their mean, standard deviation and percentile of the mean of the selected vortex anomalies. The dotted line indicates the mean of the selected vortex anomalies.

accuracy of 0.1% on the sample percentile. Within this distribution of random means, the place of the sample mean, is evaluated. If $\bar{x}_S$ is among the lowest or highest 2.5% of the values, i.e. outside the 95% confidence level, the sample mean can be seen as significantly different from a random sample mean.

Note that the mean of the bootstrap distribution will depend on the mean of the underlying dataset only, its variance σ^2 , however, depends on the sample size n_S , $\sigma^2 \sim \frac{1}{n_S}$. The latter provides a scaling factor for plots for better comparability of samples of different sizes. Figure III.9 exemplifies the use of a bootstrap test and dependence on sample size.

The sample of polar vortex $\bar{u}$ anomalies was drawn based on times of a strong subtropical jet (resulting in 7 datapoints) and its mean compared to the distribution of $n_b = 1,000$ and $n_b = 10,000$ random 7-point samples. In both cases, the selected anomalies would not be regarded as significantly different from a random sample on the 95% level (would have to be in the lowest 2.5% or the highest 97.5% of the bootstrap distribution), however the allocated sample percentile is different within the standard deviation and error of the two different bootstrap sizes.

IV. The QBOi project

Despite a sound understanding of the principles behind the generation of the QBO in the atmosphere (section II.2) and successful simulation in simplified models (section II.3), an adequate representation in global climate models remains a challenge. Nonetheless, some GCMs are able to produce an oscillation of zonal wind in the equatorial stratosphere. To assess how realistic these simulations are, and to be able to compare models against each other, a set of characteristic parameters was developed as part of the QBOi project¹, an international collaboration of eleven modelling groups with the aim of “Improving the Quasi-Biennial Oscillation in Global Climate Models” . QBOi is an activity organised by the Stratosphere-troposphere Processes and their Role in Climate (SPARC) project of the world climate research programme (WCRP).

IV.1. Development of a characteristic parameter set

Two fields are assessed for QBO characteristics: zonal mean zonal (ZMZ) wind, $\bar{u}$ (Figure IV.1) and zonal mean temperature (Figure IV.2). For all of the definitions, monthly mean data from climate models and reanalyses (Overview in Tables III.2, III.3) are used. Model ensemble means are calculated as arithmetic mean of all models, reanalysis ensemble means refer to the most up to date datasets (ERA-Interim, MERRA, JRA55).

For an assessment of the modes of temporal variability in the tropical stratosphere, Fourier analysis is applied to the time-height series of $\bar{u}$ (see Figure IV.3 showing results for ERA-Interim). The QBO is identified as the broad spectral peaks around 2-3 years. However, for some models (CMCC-CSM, UМУKCA-UCAM, UМУKCA-METO, UMSLIMCAT), the QBO period is clearly longer, so that power at lower frequencies is assigned to the QBO in these cases. Features that can be seen in the time-height cross sections (Figures IV.1, IV.2) reflect in the Fourier spectra: The QBO dominates the zonal wind in the middle stratosphere, whereas the SAO can be seen in the upper stratosphere. Below the tropopause, the oscillation is annual in

¹<http://users.ox.ac.uk/~astr0092/QBOi.html>,
biennial-oscillation/

<http://www.sparc-climate.org/activities/quasi-biennial-oscillation/>

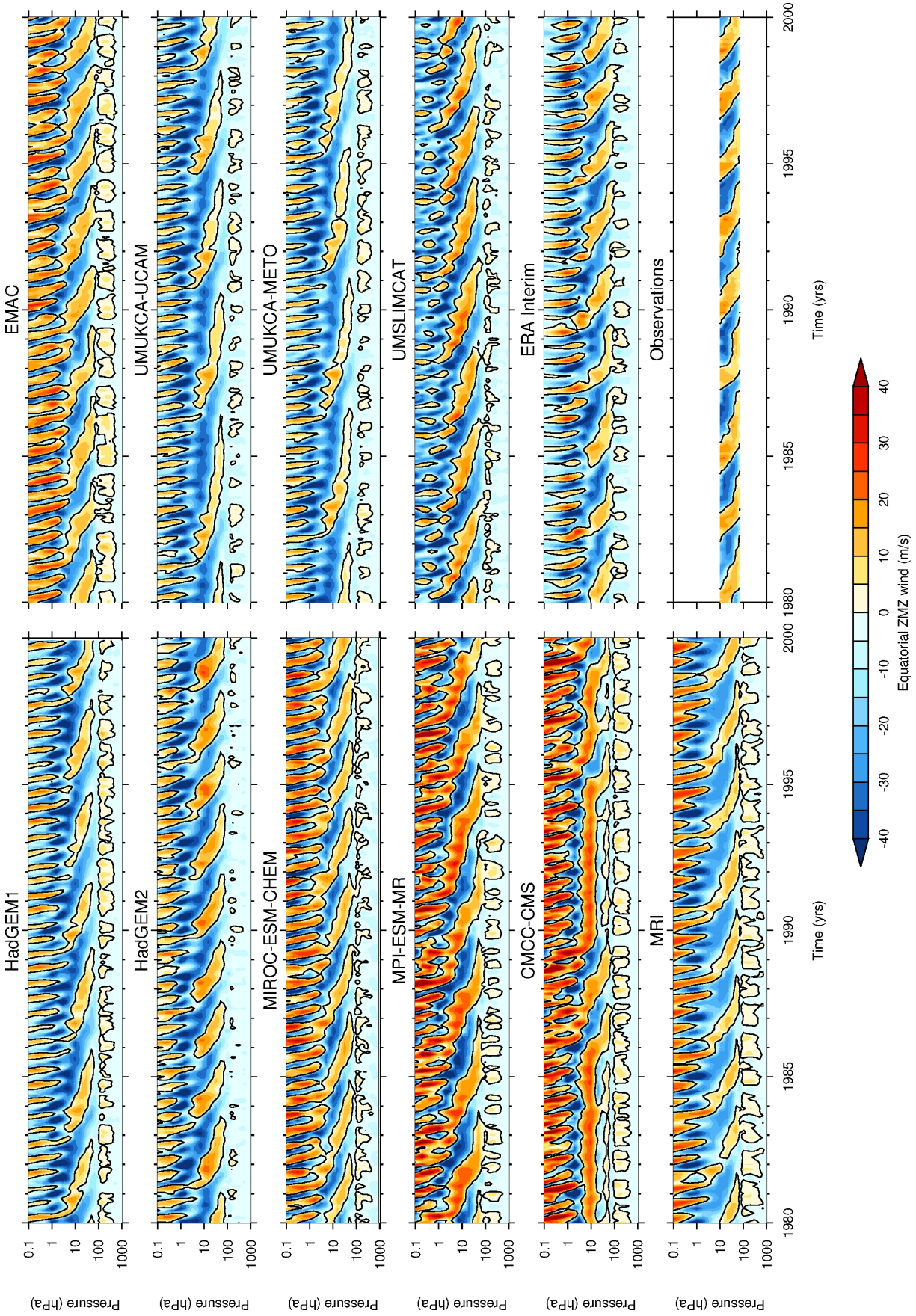


Figure IV.1.: Equatorial zonal mean zonal wind time-height series from models and the ERA-Interim reanalysis, 1980-2005. Easterly wind zones are blue, westerly shear zones red. The zero wind line is shown in black. The observational dataset from Freie Universität Berlin [2016] is shown on the bottom right for levels 10-70hPa.

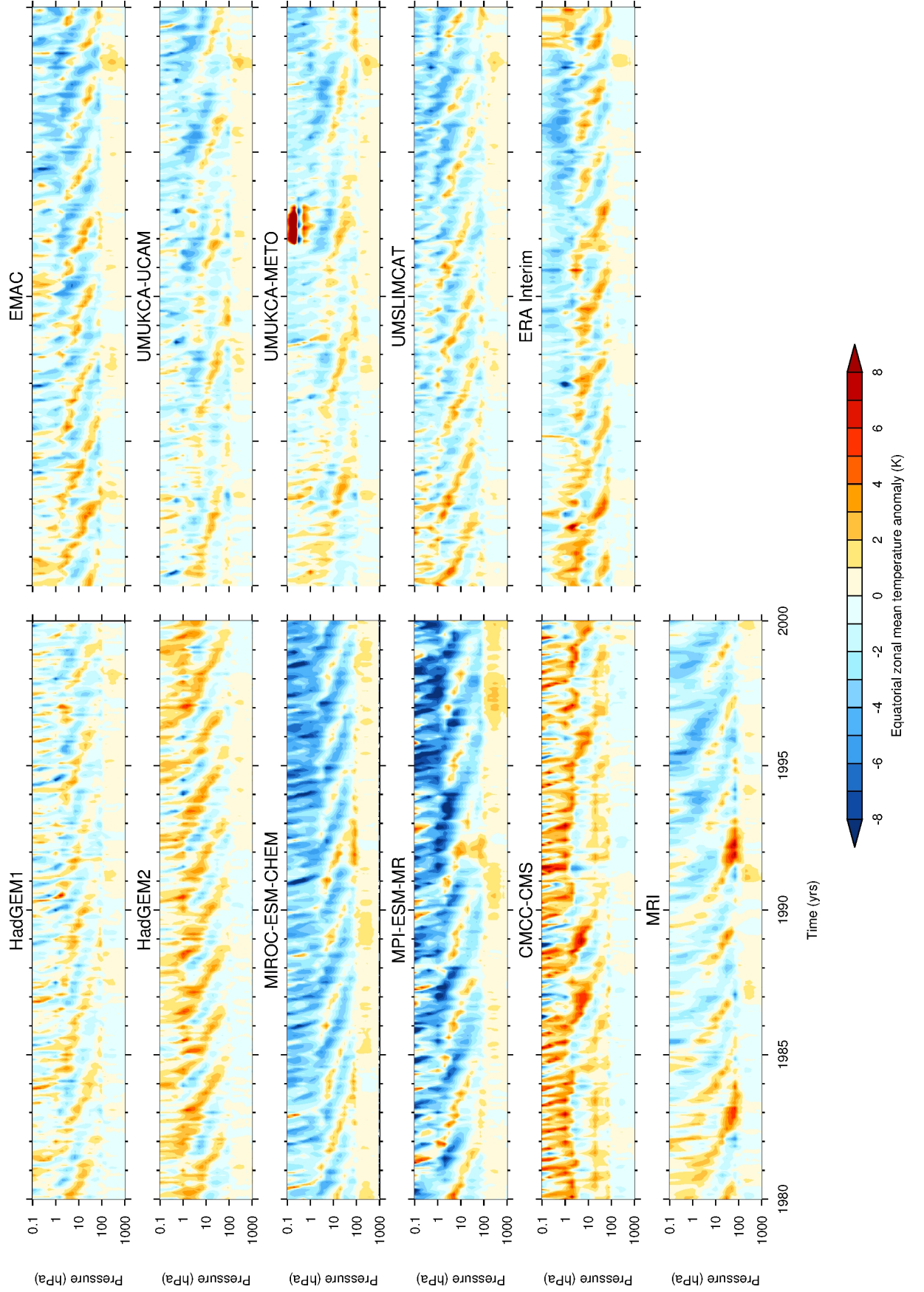


Figure IV.2.: Time-height series of zonal mean temperature anomalies (colours) from all models and the ERA Interim reanalysis.

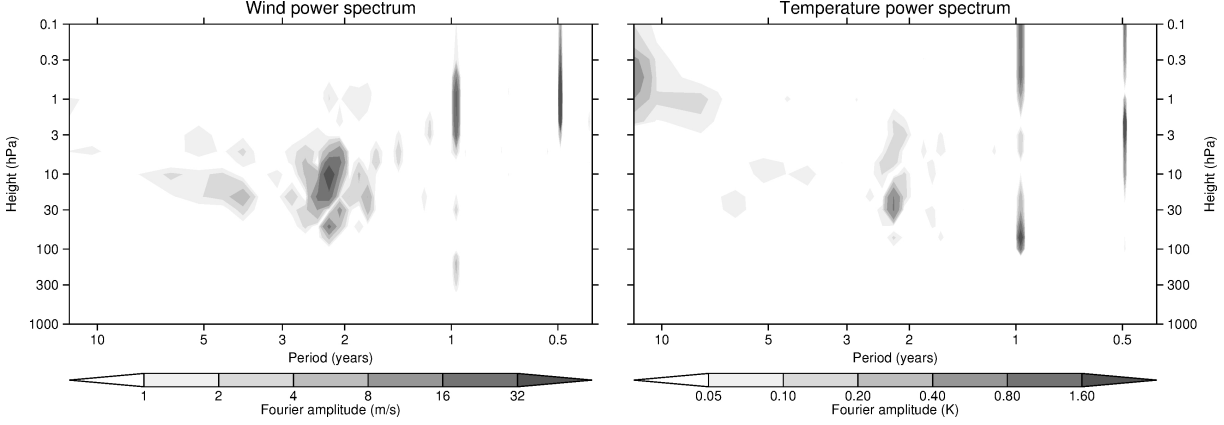


Figure IV.3.: Equatorial zonal mean zonal wind (left) and temperature (right) period-height Fourier spectrum from the ERA-Interim reanalysis, 1979-2009. This is the Fourier transformation of the $\bar{u}$ and T time-height series shown in Figures IV.1 and IV.2 and updates the ERA40 analysis of Pascoe *et al.* [2005].

the upper troposphere, with constant easterlies in the lower troposphere. A QBO-like structure can be found in the temperature field as well, however, the annual cycle clearly modulates it and is the dominant source of variability. Power at periods $\gtrsim 10$ years, especially in the temperature spectrum, can be attributed to the solar cycle.

Figure IV.4 shows the derivation of some characteristic parameters, starting from the equatorial ZMZ wind Fourier spectrum, using HadGEM2-CC as an example: The harmonics with periods around 2 years, which correspond to the QBO period, are averaged at each point in latitude and height to obtain the QBO amplitude, which is the sum of the QBO harmonics divided by the sum of all harmonics (the latter equalling the field variance σ^2): $A_{QBO} = \frac{1}{\sigma^2} \sum_{f_{QBO,min}}^{f_{QBO,max}} A^2(f)$, $f_{QBO,min} = \frac{1}{\tau_{max}}$ and $f_{QBO,max} = \frac{1}{\tau_{min}}$ are the frequencies determined by the minimum and maximum of the QBO periods τ .

This gives the latitude-height spectrum, from which the maximum Fourier zonal wind amplitude u_{max} and the respective height h_{max} , at which it occurs, is extracted. The cross section at the equator (between $\pm 5^\circ$) gives the height profile that is used to find the vertical extent (Full Width Half Maximum) and the depth/lowest level (height of 10% of the maximum amplitude) of the QBO.

A cross section at h_{max} leads to the latitudinal profile, from which the QBO extent is calculated: A Gaussian curve is fitted to the profile, and its Full Width Half Maximum (FWHM), equal to $2\sqrt{2\ln(2)}\sigma$ with standard deviation σ , is taken as the QBO width. The timeseries at h_{max} is used to calculate the QBO period: The times of phase change are found by linearly interpolating wind between two neighbouring points of opposite sign. The time between every

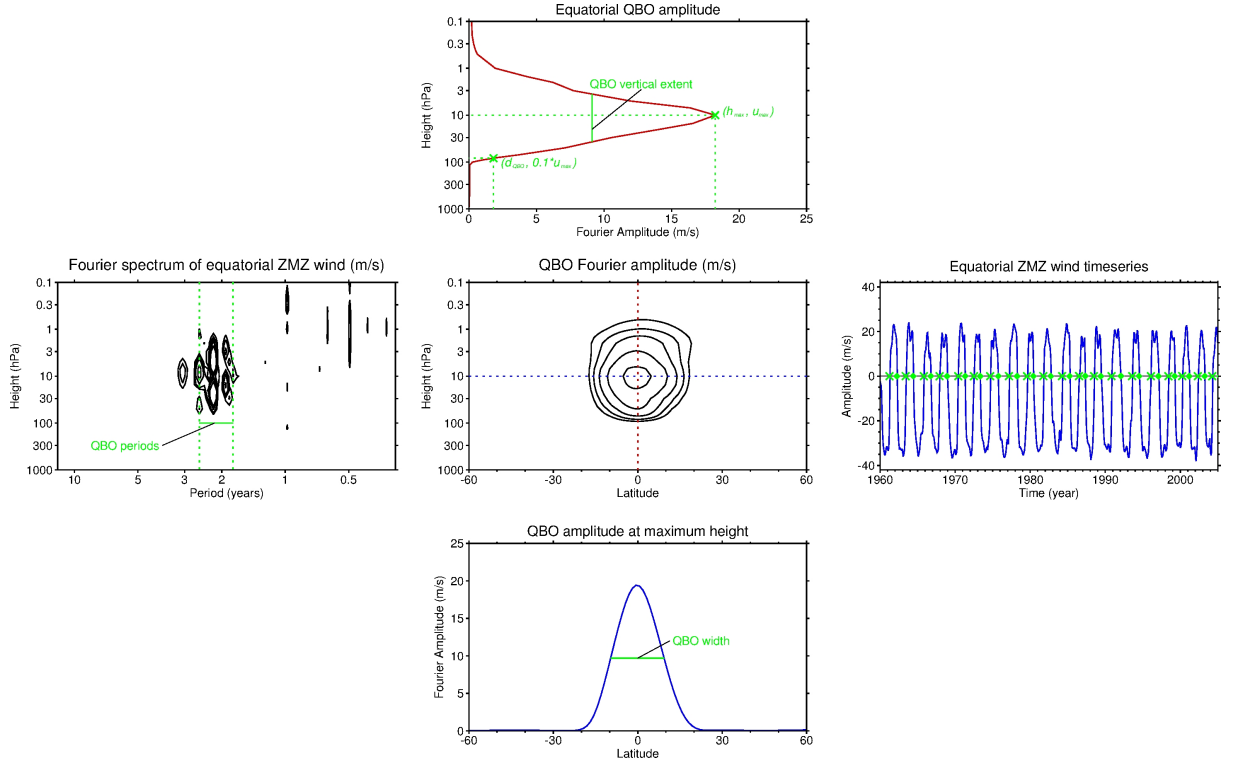


Figure IV.4.: Derivation of QBO $\bar{u}$ characteristic parameters, using HadGEM2-CC (akg11) as an example:

Middle row, left: Fourier spectrum of equatorial zonal mean zonal wind (as Figure IV.3). Contours are drawn at 1, 2, 4, 8 and 16 m/s. The Fourier harmonics around 2 years (periods derived from the timeseries of $\bar{u}$, middle row, right) are averaged to give the latitude-height QBO amplitude (middle, same contours). From this, a cross section at the equator (red) is taken to derive the QBO height profile (upper) and one at 10 hPa (blue) for the latitude profile (lower). From the height profile, the vertical extend (FWHM), the depth d_{QBO} as well as the maximum Fourier amplitude (u_{max}) can be identified. The latitude cross section at h_{max} serves to identify the latitudinal extent (FWHM) of the QBO.

From the $\bar{u}$ timeseries at h_{max} , the period of each single QBO cycle is calculated as the difference between every other zero crossing (crosses). See text for more details.

other zero crossing gives the period of one QBO cycle, and the minimum, maximum and mean period are identified as characteristic QBO parameters. An asymmetry of the phase lengths at this height can be seen in Figure IV.4, middle row, right panel. From the time of the zero crossing, the mean amplitudes of westerlies/easterlies can be identified: The maximum and minimum of each cycle is taken to be the West/East amplitude of this cycle, respectively, and the values for all cycles are averaged to give the overall West/East amplitude.

The annual synchronisation of the QBO is assessed by looking at the timing of phase changes (Figure IV.5). However, with some models and all the reanalyses only covering 30 years of data (about 12 QBO cycles), this analysis is not yet statistically conclusive. There seems to be some synchronisation, with the a preference of phase change in autumn and spring in the

observations. However, only 28 cycles contribute to this statistic, so it might not be significant. Interestingly, the phasing in reanalyses is different (though consistent in the three sets that were assessed) from the observations, favouring West-East changes in October, but not showing the peak in May. East to West changes in observations occur slightly more often in November, in reanalyses it is December and the preferred time in models is May. The discrepancy between reanalyses and observations is very likely to come from the underlying climate model - as can be seen, the phase changes occur mainly in spring and autumn, but do not necessarily coincide with the timing of observations. Kawatani *et al.* [2016] explain this with the disagreement in the representation of quasi-stationary planetary waves. Another reason for the distribution of the phase changes in models, that is similar for the West-East and East-West transitions is the lack of coupling of the wave parametrization to the actual wave sources, such as deep convection.

Another feature that has been noted in simplified model studies [Rajendran *et al.*, 2016] and suggested from observations [Kuai *et al.*, 2009; Gabis & Troshichev, 2011; Read & Castrejón-Pita, 2012] is a potential period locking of the QBO with the annual or semiannual cycle. Despite the mean period (28 months) not being an integer or half integer of an annual cycle, there might still be a preference for certain periods that fulfil phase-synchronisation (24, 30, 36 months). Figure IV.6 shows the distributions of QBO periods for models, reanalyses and observations. It peaks at 25-27 months for the models with a long tail towards high periods and no clear peak at certain periods. The distributions of the observed period and the ones in reanalyses seem to suggest some phase locking - however, with 36 (reanalyses) and 28 (observations) cycles, the results are not yet statistically conclusive and the most common periods are in the interval from 25-26 months (Observations) and 26-27 months (Reanalyses), not around 24, what would be expected from a synchronisation with the annual cycle. One reason for the discrepancy in the reanalyses might be the same as for the differing timing of phase changes.

One problem when comparing studies about QBO-annual cycle synchronisation are the different methods that are used, either by looking at the timing of phase changes or at the actual length of the periods. Furthermore, some studies look at different heights [Pascoe *et al.*, 2005; Kuai *et al.*, 2009], or deseasonalize wind to define their metrics. Subtracting monthly means does not lead to huge differences in the wind field, but with a resolution of the synchronisation measure of 1 month, it can shift the phase change into the next/previous calendar month. This is most important for the phase change measure, but less so for the actual period length which does not depend on calendar months.

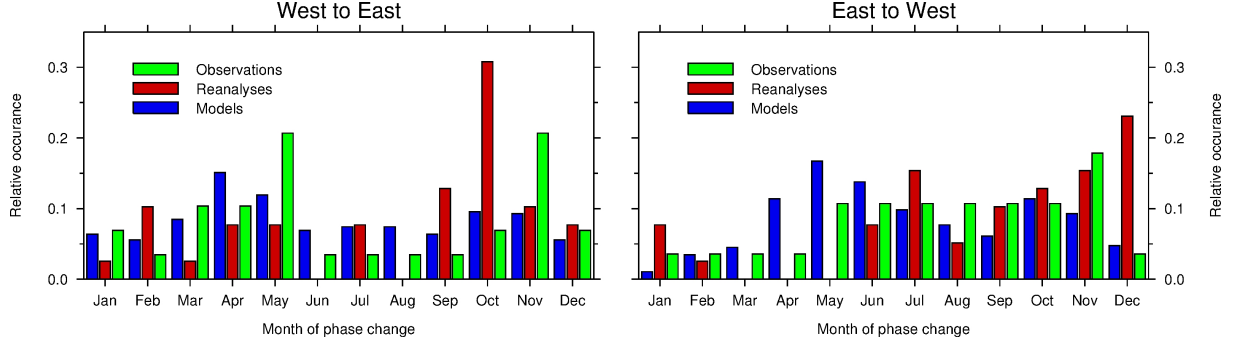


Figure IV.5.: Phase changes of the QBO at the height of the maximum. Left: Timing of phase change from West to East, Right: Timing of phase change from East to West. Blue shows the monthly distribution of phase changes from all models, red from ERA-Interim, MERRA and JRA55 reanalyses and green Singapore Observations. There are 415 QBO cycles contributing to the model statistic and 36 to the reanalysis statistic; the observations comprise 28 cycles.

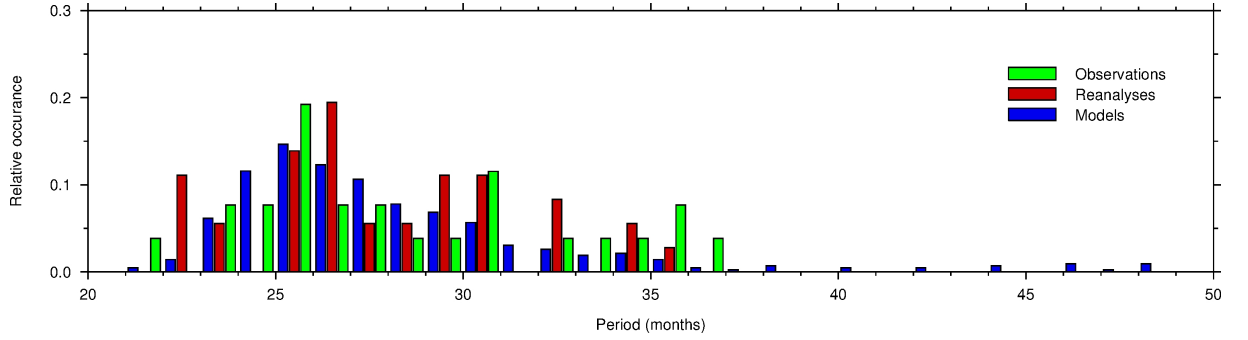


Figure IV.6.: Length of QBO period (1 month intervals) at height of maximum in models (blue), reanalyses (red) and observations (green). Long model periods are from UMUKCA-UCAM and -METO.

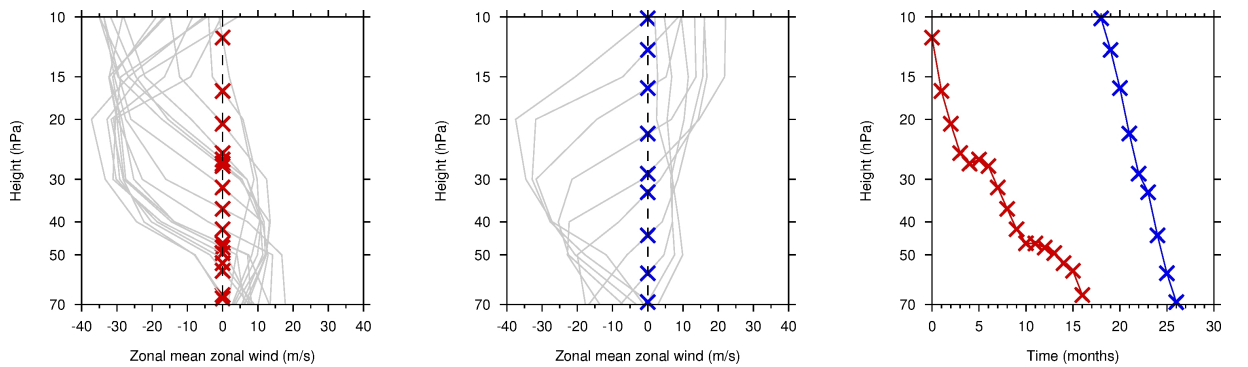


Figure IV.7.: Illustration of the derivation of the QBO descent rates: Subsequent height profiles of monthly mean equatorial ZMZ wind (grey lines) are used to identify the height of phase change where $\bar{u} = 0$ (crosses). Depending on the sign of the wind above the phase change (negative: red crosses (left), positive: blue crosses (middle)), the rates are assigned to the easterly or westerly descent rate, respectively.

Right: Time and height of zero crossings from one example QBO cycle (1964-1966). Red symbolizes the descending easterly, blue the descending westerly shear zone, as in the profiles shown in the other two panels.

Model/Reanalysis	Height of maximum (hPa) *	Fourier amplitude (m/s)	Latitudinal extent (°) *	Vertical extent (km) *	Lowest level (hPa)	Period (months)		Amplitude (m/s)			Descent rate (km/month)	
						Min	Max	Mean	East	West	East	West
HadGEM1	10	14.7 ± 1.0	17.9	18.1	78	26.2 ± 3.1	41.3 ± 2.8	33.8 ± 4.4	-35.5 ± 1.4	16.0 ± 2.2	0.67 ± 0.50	0.72 ± 0.53
HadGEM2-CC	10	18.6 ± 0.9	18.7	16.7	83	21.1 ± 1.3	31.2 ± 1.4	26.5 ± 2.5	-34.8 ± 1.5	21.4 ± 2.3	0.76 ± 0.52	1.10 ± 0.50
MIROC-ESM-CHEM	15	16.0 ± 0.2	19.6	17.9	76	22.3 ± 0.8	31.8 ± 0.9	26.1 ± 2.0	-34.7 ± 2.1	14.4 ± 1.7	0.68 ± 0.39	0.85 ± 0.55
MPLESM-MR	10	22.1 ± 0.5	20.2	18.3	84	23.5 ± 1.7	57.0 ± 7.8	30.3 ± 4.8	-38.1 ± 1.9	26.1 ± 1.5	0.58 ± 0.49	0.90 ± 0.68
CMCC-CMS	10	14.4 ± 0.8	18.8	19.0	68	22.6 ± 2.2	153.5 ± 18.6	47.9 ± 29.9	-28.3 ± 7.6	26.4 ± 2.0	0.44 ± 0.55	0.59 ± 1.41
EMAC	15	13.2 ± 0.6	21.0	18.5	76	24.1 ± 1.8	35.0 ± 2.1	28.4 ± 3.3	-29.0 ± 2.3	14.3 ± 1.4	0.73 ± 0.63	0.96 ± 0.59
MRI	20	13.3 ± 1.0	20.6	21.5	81	25.3 ± 1.8	38.3 ± 2.3	28.3 ± 3.1	-27.4 ± 1.1	12.5 ± 1.3	0.57 ± 0.44	1.03 ± 0.88
UMSLIMCAT	10	15.2 ± 1.2	18.8	22.7	77	28.5 ± 1.9	35.1 ± 1.6	31.2 ± 1.8	-36.7 ± 2.4	23.0 ± 3.2	0.83 ± 0.58	0.57 ± 0.45
UMUKCA-METO	10	11.1 ± 2.1	18.2	15.6	80	53.2 ± 7.1	62.4 ± 4.4	56.8 ± 3.3	-34.0 ± 1.8	17.0 ± 1.5	0.25 ± 0.47	0.52 ± 0.53
UMUKCA-UCAM	10	12.9 ± 0.8	18.2	16.7	87	36.0 ± 6.1	66.3 ± 4.5	49.5 ± 9.5	-34.8 ± 1.4	17.8 ± 0.9	0.33 ± 0.46	0.56 ± 0.45
ERA 40	20	14.6 ± 0.6	21.7	17.3	89	19.6 ± 1.7	34.7 ± 2.0	28.3 ± 3.8	-35.6 ± 2.4	13.4 ± 2.3	0.74 ± 0.71	1.32 ± 0.72
ERA-Interim	20	14.6 ± 1.0	20.9	15.1	86	22.9 ± 1.7	34.7 ± 2.5	28.0 ± 3.5	-34.5 ± 2.3	16.0 ± 1.4	0.79 ± 0.78	1.28 ± 0.71
MERRA	20	15.1 ± 1.1	20.8	17.7	89	22.5 ± 1.5	34.8 ± 1.8	28.0 ± 3.7	-34.6 ± 2.5	16.4 ± 1.9	0.74 ± 0.69	1.16 ± 0.68
JRA25	20	14.1 ± 1.3	20.9	19.4	87	22.9 ± 2.1	35.2 ± 2.6	28.0 ± 3.6	-35.4 ± 2.1	16.4 ± 2.0	0.73 ± 0.68	1.56 ± 0.68
JRA55	20	14.5 ± 1.2	21.3	17.7	88	22.9 ± 1.6	35.6 ± 2.9	28.0 ± 3.7	-35.6 ± 2.3	16.1 ± 1.5	0.69 ± 0.67	1.44 ± 0.59
CFSR	20	14.0 ± 0.8	21.4	16.5	83	23.2 ± 1.6	34.8 ± 2.2	27.9 ± 3.5	-34.9 ± 2.3	14.7 ± 2.8	0.75 ± 0.83	1.18 ± 0.65
NCEP1	20	11.7 ± 1.1	20.1	-	80	22.9 ± 1.5	34.5 ± 2.8	27.9 ± 3.5	-27.9 ± 3.4	15.3 ± 1.4	0.82 ± 0.79	1.55 ± 0.68
NCEP2	10	12.1 ± 0.8	20.6	-	81	22.5 ± 1.6	35.1 ± 2.7	27.7 ± 4.2	-35.6 ± 3.8	10.0 ± 3.5	0.84 ± 0.84	1.68 ± 0.87
Observations	15	15.7 ± 0.6	-	-	-	20.4 ± 1.8	36.9 ± 2.3	28.2 ± 4.5	-32.8 ± 3.2	18.3 ± 2.8	0.66 ± 0.75	1.24 ± 0.65

Table IV.1.: Overview of the calculated QBO parameters for all models and reanalyses. Values and their errors as described in section IV.2 are given. Where possible, the value from the observations is given as well. However, these only consist of one timeseries at the equator (hence no latitudinal information) and are available between 10 and 70hPa, so the depth and vertical extent could not be assessed. The level of maximum QBO in observations, 15hPa, is not a grid level in the reanalyses output; this is the most likely cause for the discrepancy between Observations and reanalyses. The highest level NCEP1 and NCEP2 is 10hPa, which accounts for the missing value in vertical extent. A comparison of the model and reanalysis means is given in section IV.3.

* The error of these parameters is mainly determined by the grid spacing (refer to Tables III.2 and III.3).

A slightly more complicated, but nonetheless important, parameter is the QBO descent rate - or rather, the mean descent rates for each of the phases. Figure IV.7 illustrates how these are defined: The vertical profile of $\bar{u}$ between 10 and 70hPa is taken and the height (in km) of the phase change ($\bar{u} = 0$) is obtained in each month. The difference in height of phase change Δh , divided by the time interval $\Delta t = 1\text{month}$ gives the descent rate $\delta_i = \frac{\Delta h_i}{\Delta t}$ in km/year for each timestep. Depending on the sign of the wind above the zero crossing, the rate is either identified as belonging to a descending easterly ($\bar{u} < 0$) or westerly ($\bar{u} > 0$) shear zone and the average of all respective rates defines the mean easterly/westerly descent rate. The descent of westerly shear zones is more uniform across different heights, whereas descent of easterly shear zones can be delayed by a couple of months, especially between 30-50hPa ($\sim 20\text{-}24\text{km}$). The latter is known as “stalling” of the QBO and accounts both for a slower mean and a greater variance of the easterly descent rate. Figure IV.7 shows one profile of a descending easterly and westerly shear zone exemplifying this difference.

For comparison, all characteristic parameters derived from ZMZ wind for models, reanalyses and observations (if possible) are listed in Table IV.1. Performance of models relative to reanalyses/observations is discussed in section IV.3.

For the temperature analysis, the anomalies at each height are calculated first by subtracting the temporal mean over all datapoints from the timeseries. The time-height series of temperature anomalies for all the models and the ERA-Interim reanalysis is shown in Figure IV.2. The same procedure as for the zonal mean zonal wind is applied to derive the Fourier spectrum, maximum amplitude, height of maximum in T , latitudinal extent and depth. Table IV.2 lists the characteristic parameters for temperature calculated from each model and reanalysis dataset.

IV.2. Error estimation

For metrics which can be calculated for each individual QBO cycle (period, easterly/westerly amplitude, easterly/westerly descent rate), the error of the respective metric is defined as the standard deviation $\sigma^2 = \frac{1}{n} \sum_{i=1}^n (x_i - \bar{x})^2$ of the mean. As this approach is not possible for a single value metric, a different method, suggested by Christiansen [2010], is applied: First, a wind timeseries of 1000 randomly chosen QBO cycles is constructed, with one cycle being defined as the timeseries between two subsequent wind minima. From this timeseries, 100 surrogates of

Model/Reanalysis	Height of maximum (hPa) *	Fourier amplitude (K)	Latitudinal extent (°) *	Vertical extent (km) *	Lowest Level (hPa) *
HadGEM1	15	0.7 ± 0.1	12.9	20.0	89
HadGEM2-CC	6	1.0 ± 0.1	14.4	19.2	96
MIROC-ESM-CHEM	7	1.4 ± 0.0	13.8	16.3	69
MPI-ESM-MR	5	1.7 ± 0.0	15.2	20.8	82
CMCC-CMS	5	1.1 ± 0.1	16.0	22.9	85
EMAC	20	1.2 ± 0.1	15.7	17.7	85
MRI	30	0.9 ± 0.1	15.2	19.6	97
UMSLIMCAT	20	1.0 ± 0.1	13.2	21.3	85
UMUKCA-METO	30	0.7 ± 0.1	12.8	19.6	113
UMUKCA-UCAM	30	0.8 ± 0.1	13.6	18.5	115
ERA 40	30	1.3 ± 0.1	16.2	14.2	97
ERA-Interim	30	1.3 ± 0.1	16.8	14.9	89
MERRA	30	1.3 ± 0.1	16.8	14.8	88
JRA25	30	1.1 ± 0.2	15.8	17.4	89
JRA55	30	1.3 ± 0.1	16.9	13.7	88
CFSR	20	1.2 ± 0.1	17.4	15.2	85
NCEP1	30	0.8 ± 0.1	15.3	-	85
NCEP2	20	0.8 ± 0.1	27.7	-	87

Table IV.2.: Characteristic QBO parameters calculated from the zonal mean temperature. Values for models and reanalyses are listed; there is no comparable observational dataset. Values and their errors as described in section IV.2 are given. Amplitudes in temperature are consistent within models and reanalyses, however, the height of the maximum is too high for half of the models. Section IV.3 discusses the common strengths and biases in models compared to reanalyses.

* The error of these parameters is mainly determined by the grid spacing (refer to Tables III.2 and III.3).

the same length as the original dataset are taken as QBO and subjected to the same analysis procedure as the original. The standard deviation of the metrics calculated from the surrogates is taken as error for the minimum/maximum period and the maximum Fourier amplitude. Further away from the equator and outside of the QBO dominated region, the construction of surrogate timeseries is not possible; therefore, error calculation of the lowest QBO level, latitudinal and vertical extent, are not possible using either of the described methods. These metrics will largely depend on the resolution of the model/reanalysis listed in Tables III.2 and III.3.

IV.3. Assessment of model performance

Having defined quantitative characteristic parameters, it is possible to compare the model performance to reanalyses. Table IV.3 gives an overview of model and reanalysis ensemble means (from ERA-Interim, MERRA and JRA55) and their standard deviations. No weighting was

applied when calculating the means, so an outlier might contribute to a strong bias or high spread. Across the reanalyses, the standard deviation is small in all parameters, which means that they agree well on the QBO representation. The QBO in models, however, is not necessarily consistent with reanalysis/observations. While models usually get the asymmetry in East and West amplitudes correct, as well as the overall Fourier amplitude, the whole QBO seems to be shifted upwards, resulting in the height of the maximum and the depth (i.e. the lowermost level affected by the QBO) being too high, even with an adequate total vertical extent. This bias in height can be linked to different factors: one is the gravity wave parametrisation in models. Its amplitude spectrum (as function of vertical wave number) has to be adapted to realistically simulate a number of different phenomena and thus can not only be optimised for adequate QBO morphology. Secondly, the diffusion around the tropopause might be too strong for a QBO pattern to develop at these levels. Further, the momentum deposition can only happen at levels represented in the models, so will depend on the horizontal resolution.

In latitude, the modelled QBO is slightly too narrow in both the wind and temperature field. In the wind field, the models' widths also decay too fast with decreasing altitude as seen in Figure IV.8, left panel. Again, this might be linked to shortcomings in the gravity wave parametrisations or coarse resolution.

The Fourier amplitude of the temperature is realistic (Table IV.3); however half of the models peak much too high (5-7hPa), whereas the others peak at about the correct height (20-30hPa). This leads to a huge spread in peak height in the models and an elongated structure in height when looking at the model mean spectrum. Additionally, models exhibit a too high value in QBO temperature amplitude around $\pm 20^\circ$ latitude. This bias in height is likely to have an influence on the meridional circulation and thus the temperatures in the subtropical latitudes. These structural biases are shown in Figure IV.8, right panel.

The mean period also seems to be too long in models, however, the high standard deviation of the multi-model mean as noted in table IV.1 shows that some models have very realistic periods, whereas a few (UMUKCA-METO, UMUKCA-UCAM and CMCC-CMS) are biased much too long. Excluding these, the period distribution of the multi-model mean agrees well with the reanalysis mean (Table IV.3). In the case of the UMUKCA models, that were part of the CCMVal-2 project, the excessively long periods might be due to chemistry feedbacks (e.g. ozone heating) and an experiment objective that did not necessarily focus on stratospheric dynamics.

Excluding the models with obvious shortcomings regarding the QBO period, some other

<i>ZM Zonal Wind</i>	Model mean	Model mean (ex)	Reanalysis mean
Height of maximum (hPa)	12.0 ± 3.5	13.3 ± 4.1	20.0 ± 0.0
Fourier amplitude (m/s)	15.1 ± 3.2	16.3 ± 3.5	14.8 ± 0.3
Latitudinal extent ($^{\circ}$)	19.2 ± 1.1	19.7 ± 1.2	21.0 ± 0.3
Vertical extent (km)	18.5 ± 2.2	18.5 ± 1.6	16.8 ± 1.5
Lowest Level (hPa)	79.1 ± 5.2	79.8 ± 3.6	87.6 ± 1.7
Mean Period (months)	35.9 ± 11.2	28.9 ± 2.8	28.0 ± 0.0
Min Period (months)	28.3 ± 9.7	23.8 ± 1.9	22.7 ± 0.2
Max Period (months)	55.2 ± 36.9	39.1 ± 9.6	35.0 ± 0.5
Amplitude Easterly	-33.3 ± 3.7	-33.3 ± 4.1	-34.0 ± 1.9
Amplitude Westerly	18.9 ± 5.0	17.5 ± 5.2	15.8 ± 0.8
Descent rate Easterly	0.6 ± 0.2	0.7 ± 0.1	0.7 ± 0.0
Descent rate Westerly	0.8 ± 0.2	0.9 ± 0.1	1.3 ± 0.1
<i>ZM Temperature</i>			
Height of maximum (hPa)	16.8 ± 10.7	13.8 ± 9.9	30.0 ± 0.0
Fourier amplitude (m/s)	1.1 ± 0.3	1.2 ± 0.4	1.3 ± 0.0
Latitudinal extent ($^{\circ}$)	14.3 ± 1.2	14.5 ± 1.0	16.9 ± 0.0
Vertical extent (km)	19.6 ± 1.9	18.9 ± 1.6	14.5 ± 0.7
Lowest Level (hPa)	91.5 ± 14.0	86.3 ± 10.2	88.0 ± 0.7

Table IV.3.: Overview of the characteristic QBO parameters in reanalyses and models. Values are means and standard deviations of the parameters in tables IV.1 and IV.2. The model mean was calculated from all models (“ex” excluding CMCC-CMS and both UMUKCA models for obvious shortcomings in QBO modelling (Figure IV.1)), the reanalysis mean from the most recent datasets, namely ERA-Interim, MERRA and JRA55.

biases are alleviated slightly, such as lowering the height of the QBO maximum and widening the latitudinal extent and lowering the QBO influence down to nearly 80hPa. However, the bias in descent rates, which are not asymmetric enough, remains, and the Fourier amplitude is even higher. Apart from the mean period, none of the parameters is significantly improved compared to the reanalysis ensemble mean.

Another important QBO feature that is not captured well in models is the asymmetry in the descent rates of easterly and westerly shear zones. Figure IV.9 shows the mean descent rates for both shear zones in the models, reanalyses and observations as well as their respective standard deviation. All but one model show slower descending easterlies than Westerlies, however, the difference between the shear zones is not as prominent as in reanalyses and observations, where Westerlies descend about twice as fast. The standard deviation of the descent rates in the models is comparable for both shear zones, whereas in reanalyses and observations it is higher for descending easterlies. The prominent stalling features of the observations (e.g. in 1964), where the descent of the easterly shear zone can be halted for up to 12 months, are sometimes not even represented in models (Figure IV.1). In the multi-model/-reanalysis mean, the standard deviation is much higher for the westerly descent rate; the easterly descent rate is fairly realistic.

The discrepancies in the westerly descent rates, which occur in reanalyses as well as in models, is likely to be caused by the forcing of sub-grid scale waves, which have to be parametrized. This is also indicated by the differences in the timing of phase changes (Figure IV.5). Kawatani *et al.* [2016] note the delay in the East-West transition in reanalyses as well. They explain it by weak westerly forcing from resolved waves (that is not accounted for by parametrized waves, either).

One other reason why the timings of phase changes differ between models and the observations could be the lack of coupling of the parametrized gravity wave to actual orographic and nonorographic sources. The parametrisation schemes that were used in the models are “globally spectral” ones which do not take different local wave sources into account. However, with improvement in modelling it is possible to link gravity wave drag to different sources (convection, fronts, planetary wave breaking), which may vary locally and over the year (see de la Cámara *et al.* [2016] and references therein). Assessing these improved models within the framework presented here could resolve on whether the shortcomings so far are tied to the gravity wave parametrizations.

For future analysis in the scope of the next model intercomparison (CMIP6), it would be desirable to store more fields for QBO analysis. These include the transformed Eulerian mean of the

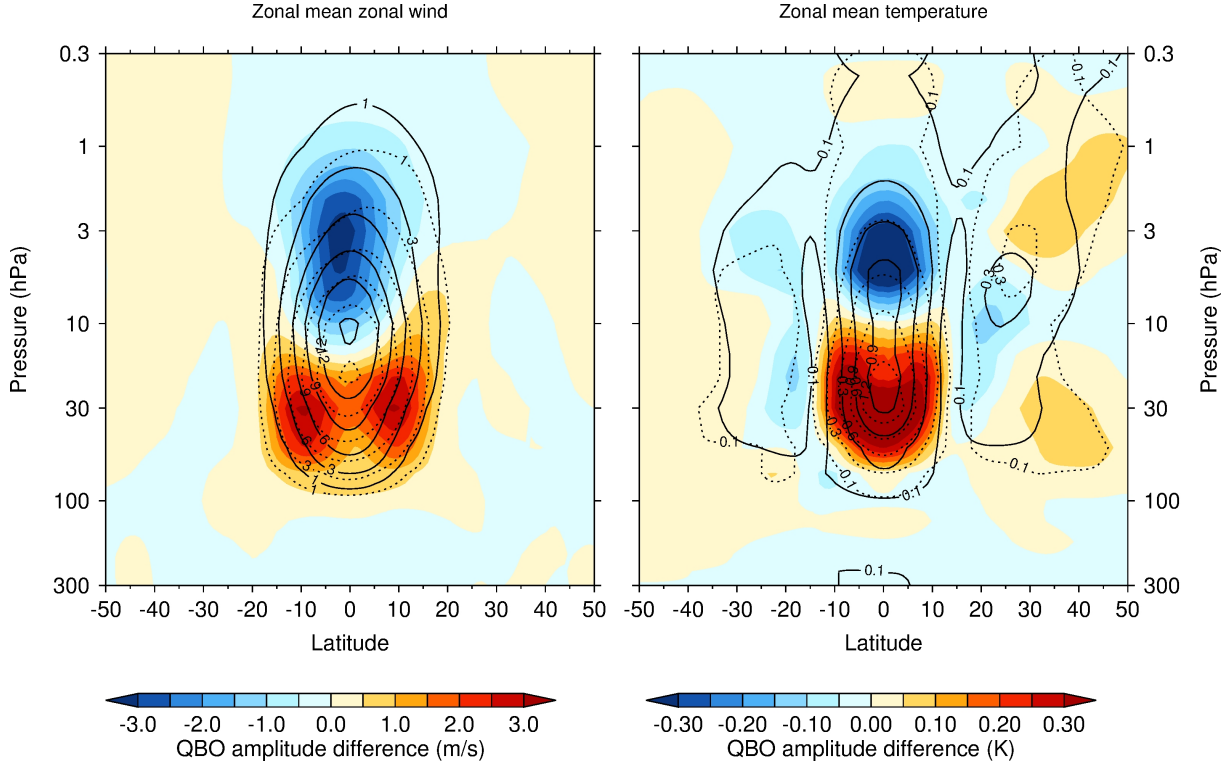
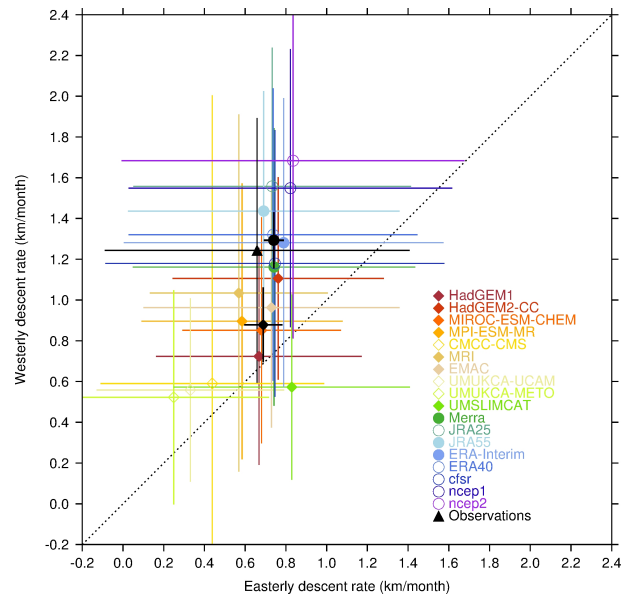


Figure IV.8.: Latitude-Height structure of the QBO in zonal mean $\bar{u}$ (left) and T (right). Solid contours indicate the model mean, dashed contours the reanalysis mean, colours show the difference of reanalysis and model mean, with positive values indicating the models being too low in amplitude and negative values models being too high.

Figure IV.9: Westerly vs. easterly descent rates in all analyzed datasets. The symbols (diamonds for models, circles for reanalyses, triangle for observations) show the respective mean descent rates, the coloured lines the corresponding standard deviation within one dataset. The black diamond/circle shows the multi-model/-reanalysis mean and the lines the standard deviation of this mean. Filled symbols indicate models/reanalyses that are included in the calculation of the mean values. The dotted line is drawn for reference and indicates equal descent rates for both shear zones.



meridional circulation ($\bar{v}^*$, $\bar{w}^*$) as well as the Eliassen-Palm flux and parametrised gravity wave drag. This would make a comparison of the interaction between the QBO and the meridional circulation possible, especially regarding the height bias of the models. Further, including an analysis of the involved forcings (advection, resolved and parametrised waves) might shed some light on the causes of the misrepresentation of descent rates and stalling events. In models that exhibit a too long QBO period, the wave forcing is also very likely to be a major contribution to this bias, but without the respective data, this cannot be conclusively assessed.

Being able to quantify the quality of a QBO simulation is an important achievement; however, it can only be a first step towards a better simulation of the stratosphere and its impacts. The QBO is not the only predictable feature in the Earth’s climate system, and might not be the most important one when it comes to surface weather (e.g. ENSO certainly has a global influence on the latter). Hence, simulating a QBO for the sake of getting a “beautiful” stratosphere cannot be the goal, neither can tuning parameters in the gravity wave scheme be the way to achieve it if other patterns get unrealistic in the process. Some questions to be answered, are which parameter(s) is/are the most important ones to get right? What is it about the QBO that can influence the surface? How strong is this influence? How does it change with the seasons? An attempt to address these will be made in Chapter V.

IV.4. Possible parameter derivation from EOF analysis

Further analysis of the wind field is done using Empirical Orthogonal Functions (EOFs, section III.2.5). This method, first applied to the equatorial zonal wind by Wallace *et al.* [1993], can be used as an alternative for deriving some of the parameters described above, or to define a QBO index for regression analysis (Section IV.4). The EOFs are calculated as spatial functions of the height ($M = 7$ for the observations) and corresponding principal component timeseries are derived for each point in time (years 1956-2015, $N = 720$).

Examination of the reconstructed zonal wind field using only the first two EOFs (Figure IV.10), that account for about 71% of the field variance, shows that the main structure of alternating easterly and westerly shear zones is represented well. Even the difference in the shear zone descent rates (compare Figure IV.7 and Table IV.1) can be captured, but is not as prominent as in the full field. Some of the interesting features of the QBO, like long stalling phases, where the descent of the easterly shear zone can be halted for up to 6 months (e.g. in 2000), are not

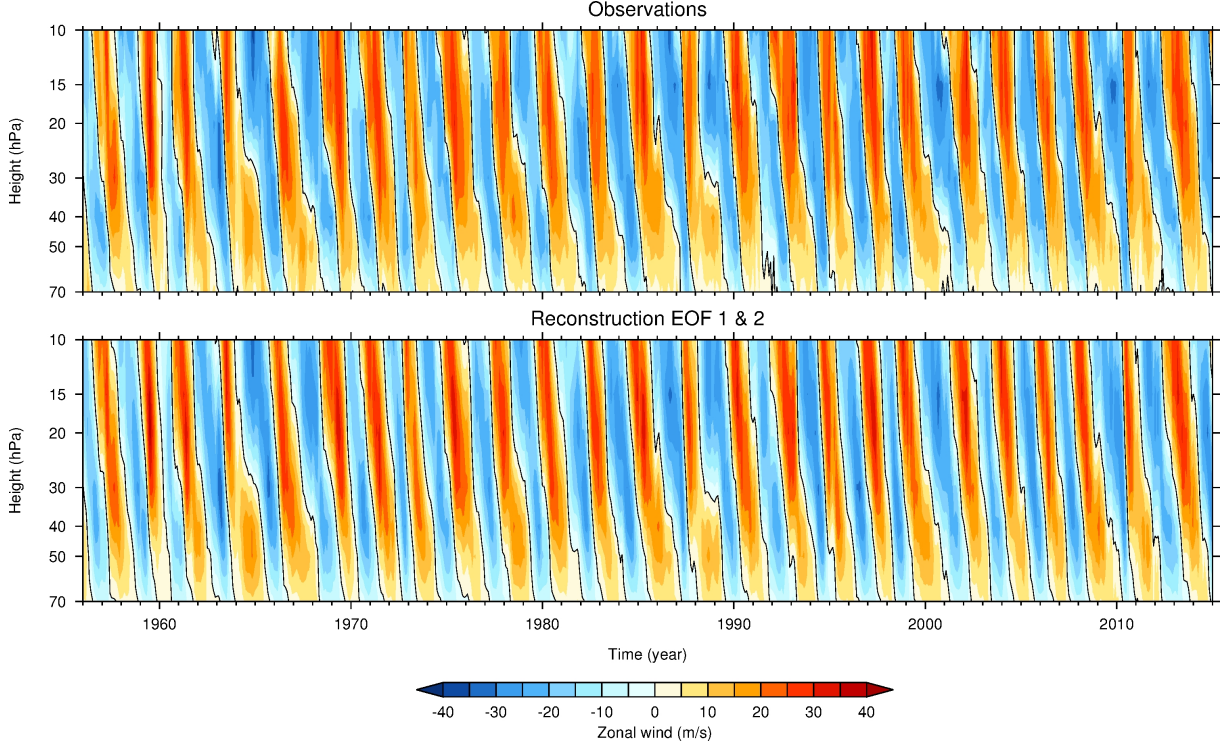


Figure IV.10.: Observation and reconstructed wind field from the first two EOFs, 1956-2014. The timeseries at each height is calculated as $\sum_{k=1,2} e_k \lambda_k p_k(t)$, with e_k indicating the value of the spatial EOF at height k , λ_k being the corresponding eigenvalue and $p_k(t)$ the corresponding principal component timeseries. The QBO structure of alternating easterly (blue) and westerly (red) shear zones is captured very well, however, some details like the very slow descent of the easterly shear zone between 2000 and 2001, are not represented exactly. A good reference for the details is the zero wind line (black) that shows more variance in the full dataset compared to the reconstruction (e.g. 1991-1992, between 50 and 70hPa).

quite realistic and get smoothed out in the reconstruction.

From the first two EOFs, the possibility of an alternative parameter derivation arises, specifically for the QBO period and descent rate. The length of a QBO cycle can be defined straightforwardly from the principal component timeseries associated with the first two EOFs as the time needed for the completion of a full 2π cycle in phase space (Figure IV.12). The derived mean period from the EOF phase angle $\psi = \arctan(\frac{p_2}{p_1})$, which is the polar coordinate angle of the point (p_1, p_2) as the average of the principal component timeseries is equal to zero, $\Pi_\psi = 28.0 \pm 3.6$ months compares very well with the period calculated directly from the wind timeseries at 20 hPa ($\Pi_{20hPa} = 28.2 \pm 4.4$ months); thus, it provides an alternative way to the definition of the parameter. In both cases, the period is defined as the mean from all periods and the error is one standard deviation of the value. Another possibility is to define a phase progression rate instead of a descent rate which has the advantage of being less complicated

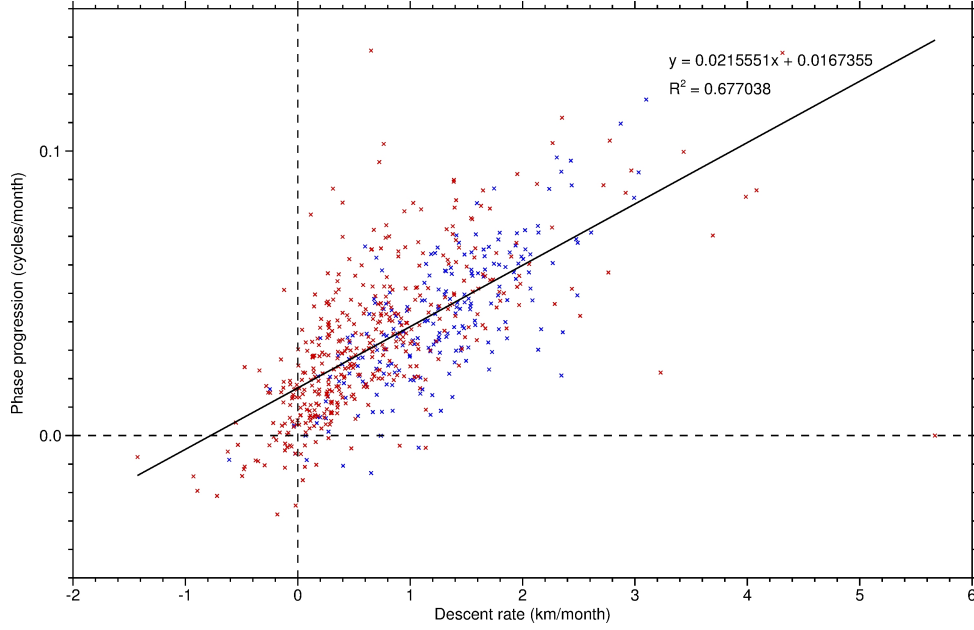


Figure IV.11.: Comparison of the phase progression rate with the descent rate from the FU Berlin observations. Red crosses: descending easterly shear zone, Blue crosses: descending westerly shear zone. The black line shows a least squares fit to the data points.

to calculate. The time derivative $\frac{\partial \psi}{\partial t} = \dot{\psi}$ can be used to assess the phase progression of the QBO, as noted by Wallace *et al.* [1993], and is well correlated with the descent rate described in section IV.1 (Figure IV.11). In contrast to the descent rate, which relies on zero crossings of the wind profile (there can be more than one, or none at all), the phase progression rate can easily be defined at each point in time. A least squares fit to descent rate vs. phase progression rate exhibits a slope of approximately 0.022/km, with an R^2 value of 0.68. According to a two-sided t -test, the coefficient is significant on the 99% level. This makes the phase progression rate a valuable characteristic parameter that is closely related to the descent rate.

Alternative ways to define a QBO phase are presented by Fraedrich *et al.* [1993] or Read & Castrejón-Pita [2012], who apply delay embedding (which is a modification of EOF analysis, but EOFs are defined as functions of time and space) and analytic phase (combining singular systems analysis and Hilbert transformation) approaches, respectively. Fraedrich *et al.* [1993] are able to extract a less noisy QBO signal from their dataset, giving a nearly perfect circle in phasespace and less variance in the QBO period. The method presented by Read & Castrejón-Pita [2012] is very useful in assessing phase synchronisation of the QBO with other variables. However, both studies focus more on QBO dynamics and interactions, in contrast to the more morphological approach of the QBO metrics presented in section IV, for which the more straightforward method based on Wallace *et al.* [1993] seems reasonable. Fraedrich *et al.* [1993] themselves acknowledge

Height (hPa)	ψ_W ($^\circ$)	ψ_E ($^\circ$)
15	-93.2	87.7
20	-66.5	114.5
30	-25.2	155.8
40	5.0	-176.0
50	25.7	-155.3

Table IV.4.: Height of lower shear zone boundary (i.e. the height of phase change) and corresponding phase angle ψ_W for the descending westerly and ψ_E for the descending easterly shear zone, using FU Berlin data.

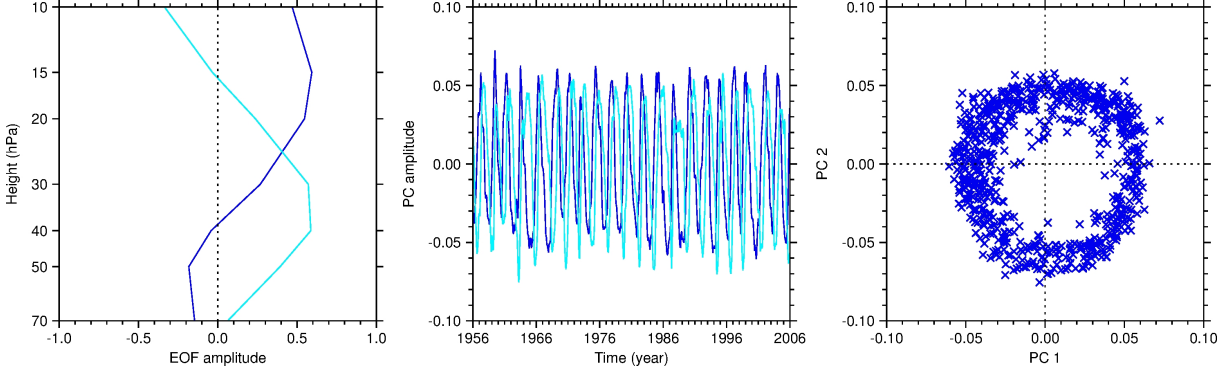


Figure IV.12.: Left: Height structure of the first two empiric orthogonal functions, derived from the FU Berlin QBO dataset, 1956-2015, 10-70hPa. EOF 1 (blue) shows the anticorrelation of the high and low levels, whereas EOF 2 (turquoise) captures the variability at the intermediate heights. Middle: associated first two principal component timeseries. Right: Phasespace of the first 2 EOFs defined by their associated principle component timeseries $p_1(t)$ and $p_2(t)$.

the usefulness of the simple EOF analysis for example for assessing the state of the atmosphere in terms of QBO phase.

The phase angle ψ can be used as a proxy for the QBO state, defined for example by the angle corresponding to the height of phase change (Table IV.4). This relationship is derived from the height of phase change in EOF profiles $\cos(\psi)\vec{e}_1 + \sin(\psi)\vec{e}_2$ with $\vec{e}_1$ and $\vec{e}_2$ being the first two EOFs as seen in Figure IV.12. It can be used to classify QBO states more comprehensively than by just using the sign of the wind at a certain level.

IV.5. Observational EOF analysis

The analysis presented here is based the technique described by Wallace *et al.* [1993] and comments on possible physical interpretations of an index derived from EOFs. The height profile of the first two EOFs as well as their associated principal component (PC) timeseries of the FU Berlin QBO dataset (1956-2014, between 70 and 10hPa) is shown in Figure IV.12. These account for 39.7% and 31.0% of the total variance respectively. Both PC timeseries are oscillations

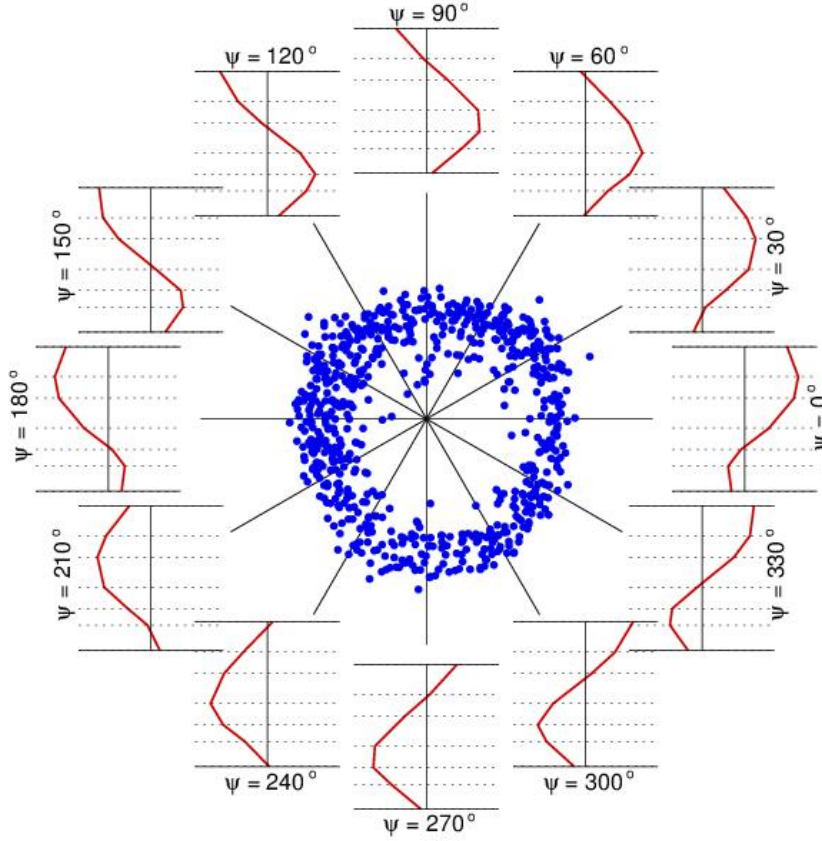


Figure IV.13: The EOF phasespace from IV.12 (blue dots) including exemplifying QBO profiles at certain angles ψ . The pressure levels 70, 50, 40, 30, 20, 15 and 10 (bottom to top) are drawn for every profile. Shifting the phase by φ leads to a projection onto new axes that are rotated *clockwise* compared to the ones shown here (Figure IV.17).

with an approximate 2 year period, being $\sim 90^\circ$ out of phase. The QBO phase space therefore is approximately a circle around the origin. Together the first two EOFs account for about 71% of the wind variance in this height region during 1956-2015.

Figure IV.13 shows the phase space spanned by EOF 1 and 2 as well as exemplifying QBO profiles spaced $\Delta\psi = 30^\circ$. These example profiles are calculated as $\cos(\psi)\vec{e}_1 + \sin(\psi)\vec{e}_2$ and are also used to derive the respective height of phase change. The QBO is evolving counterclockwise in phase space, as can be seen by the descent of the shear zones and the corresponding height of phase change (Table IV.4). For a descending easterly ($60^\circ \leq \psi \leq 240^\circ$), the phase space is more densely populated which is due to a slower descent of easterlies (example in Figure IV.7).

Figure IV.16 updates the analysis from Wallace *et al.* [1993] with the FU Berlin data to present day. The phase angle ψ can be used to identify possible annual synchronisation. If a phase locking occurred, then some phase angles would occur more often in certain months, leading to monthly means different from zero. However, the observational record does not confirm this expectation (Figure IV.16, left panel). The mean phase angle might have an annual modulation, but has a high standard deviation in each month as well, making the significance of this difference questionable. Given the observational record of about 60 years (28 cycles) statistically significant statements are difficult to make about a possible synchronisation anyways, though it has been

Figure IV.14: Length of QBO period (1 month intervals) in the observational dataset, defining the period as time needed for completion of a 2π cycle in EOF phasespace.

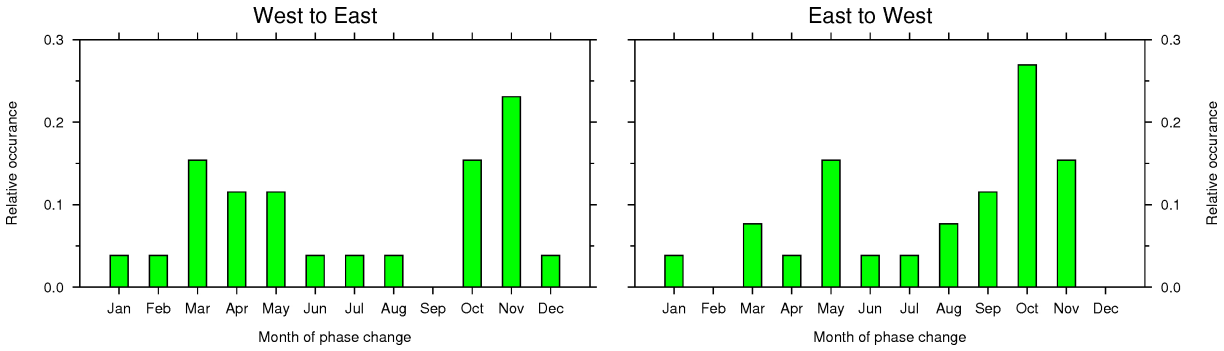
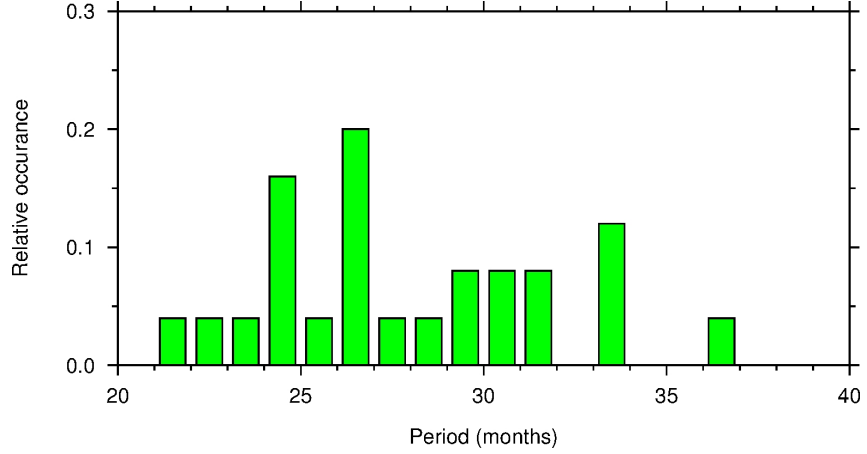


Figure IV.15.: Timing of phase change using the EOF phase angle ψ as phase reference, i.e. $\psi = -66.5^\circ$ for a West-East transition and $\psi = 114.5^\circ$ for an East-West transition.

pointed out in both climate model [Krismer *et al.*, 2013] and reanalysis studies [Pascoe *et al.*, 2005; Read & Castrejón-Pita, 2012; Taguchi & Shibata, 2013].

Figure IV.14 shows the period distribution for the FU Berlin observational dataset, using the time needed for completion of a 2π cycle as period. Similarly to the period defined by the zonal wind timeseries at level of maximum QBO (Figure IV.6), there is no clear phase locking at 24, 30 or 36 months, though there are more periods falling into the 24-25 month interval. Using the phase angle ψ to define phase transitions, however, there seems to be a clearer locking to the annual cycle. Compare Figure IV.14 to Figure IV.5: defining the phase transition via the phase angle corresponding to a phase change at 20 hPa (height of maximum QBO) by the angles listed in Table IV.4, i.e. $\psi = -66.5^\circ$ for a West-East transition and $\psi = 114.5^\circ$ for an East-West transition, leads to a clustering of transitions in boreal spring and autumn. Especially the East-West transitions show a more distinct peak in late autumn/early winter when using the phase angle definition.

Recent analysis has suggested that more synchronisation occurs with increased mean upwelling (that would be caused by warmer climate), as shown by a simplified model study [Rajendran

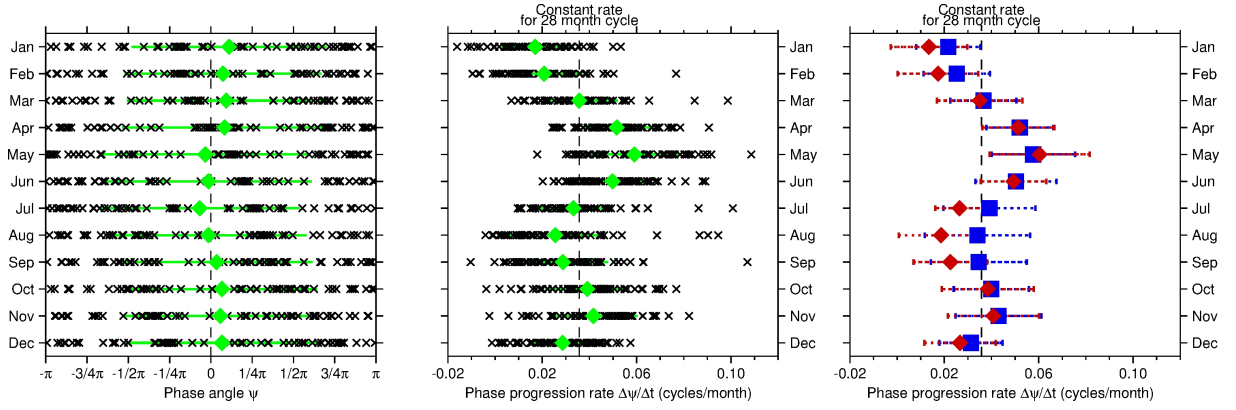


Figure IV.16.: Left: Phase angle ψ for each month and year. The monthly mean and ± 1 standard deviation is shown in green. Though the mean phase is not equal to zero in any month, this difference is not significant. Middle: Phase progression rate $\dot{\psi}$ for each month and year. The vertical dashed line indicates the rate of 1 cycle per 28 months. Right: Mean descent rate and standard deviation for a descending easterly (red) and westerly (blue).

et al., 2016]. Also, Read & Castrejón-Pita [2012] point out a synchronisation with the semi-annual cycle with a frequency ratio between 4:1 and 6:1. Schulte [2016], however, find harmonics of 14 months in the QBO timeseries as well as nonlinear interactions with modes of 10, 16 and 26 months, none of which are in agreement with an annual cycle modulation.

In contrast to the phase ψ , the phase progression rate $\dot{\psi}$ shows a clear annual cycle, which is modulated by a semiannual cycle. During January and February, the progression rate is low, which are the months where most of the stalling periods occur; some of the very slow descents also happen in boreal summer (c.f. Gabris & Troshichev [2011]), as represented by relatively small $\dot{\psi}$. In contrast, NH spring and autumn descent rates are faster than an average constant rate of $1/28^{\text{th}}$ of a cycle. These findings are consistent with Wallace *et al.* [1993] and with the parameter analysis in section IV.1. This behaviour becomes even more clear when separating the descent rates of the shear zones. Figure IV.16, right panel, shows the mean and standard deviation of easterly and westerly shear zones separately for the different months. The standard deviation in the phase progression rate is higher for descending easterlies and with the exception of May, easterlies descend slower than westerlies.

Given that the first two EOFs account for 71% of the zonal wind field variance, their respective phase space can be identified with representing QBO states. This gives a more comprehensive classification than taking the wind at only one level. By the criterion of North *et al.* [1982], eigenvalues λ_i have a sampling error of $\delta_{\lambda_i} \approx \lambda_i \sqrt{\frac{2}{N}}$, where N is the number of realizations. With eigenvalues of $\lambda_1 = 0.397$ and $\lambda_2 = 0.310$ the first two EOFs fall within the error range of each other and cannot be claimed to be distinct. This means that any linear combination of

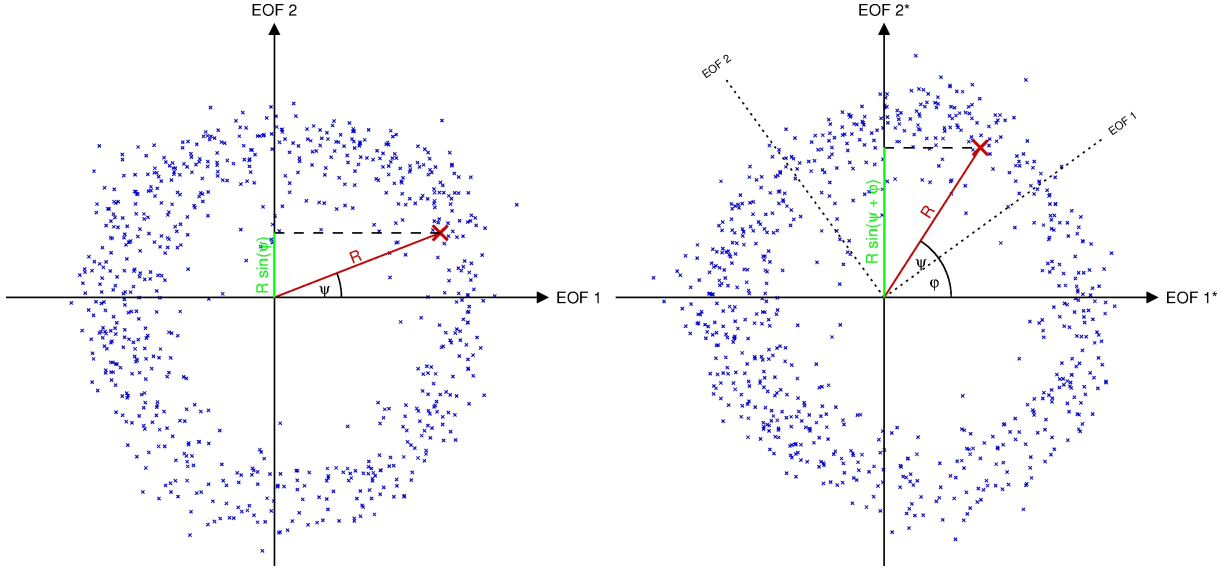


Figure IV.17.: Left: Phase space of the first two EOFs (compare Figure IV.12) from the observational zonal wind at the equator. Scaling of the axes is equal, which gives a circular trajectory in the phase space, that would be expected from two oscillations that are 90° out of phase. One point in time is highlighted in red and serves as an example of the construction of the QBO timeseries u^* in the unrotated phase space. Right: Addition of an (arbitrary) phase angle $\varphi = 36^\circ$ to construct $u^* = R \sin(\psi + \varphi)$, projecting onto the new axis EOF2*. For orientation, the original axes EOF 1 and EOF 2 are drawn dashed.

the first two EOFs will be an adequate representation of the respective subspace (i.e. the QBO phase space). The degeneracy of EOF 1 and 2 is used when constructing a QBO index from the PCs, following Baldwin & Dunkerton [1998], as $u^* = R \sin(\psi + \varphi)$ with $R = \sqrt{p_1^2 + p_2^2}$ and $\psi = \arctan(\frac{p_2}{p_1})$ where p_1 and p_2 denote the PC of the first and second EOF, respectively. φ is an arbitrary phase shift that can be interpreted as a rotation in phase space, thus a projection on a coordinate system that is rotated by $-\varphi$ to the one spanned by the EOFs. For $\varphi = 0^\circ$, $u^* = R \sin(\psi)$ is the projection of a point (p_1, p_2) onto EOF 2. Addition of φ to the phase angle ψ gives $u^* = R \sin(\psi + \varphi)$, which is the projection onto a new axis, EOF 2*. This axis itself can be interpreted as a combination of EOF 1 and 2; specifically $\text{EOF } 2^* = (\sin(\varphi), \cos(\varphi))$ where the coordinate system is spanned by $\text{EOF } 1 = (1, 0)$ and $\text{EOF } 2 = (0, 1)$. Figure IV.17 explains this further.

The reconstructed QBO index u^* is a useful single timeseries that can serve as a forcing parameter in regression analysis. Further, the phase angle can be used to classify QBO states for composite analysis. This makes the EOF analysis a helpful tool in QBO impact assessment. However, the interpretation of the phase shift φ is not trivial. Baldwin & Dunkerton [1998] relate it back to one most influential pressure level, looking at the height that shows the highest correlation with a certain phase angle (Figure IV.18). This suggestion seems a bit too simple,

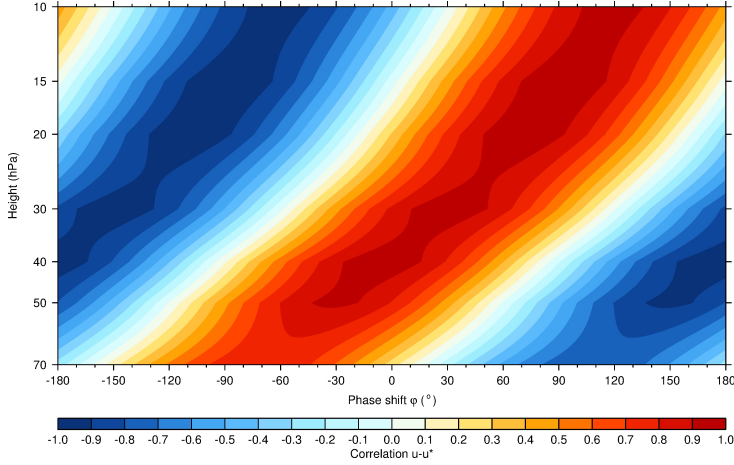


Figure IV.18: Correlation between the reconstructed index $u^* = R \sin(\psi + \varphi)$ with the measured equatorial wind u at different heights. The indicated height levels are the ones where there is data available. A positive φ means a negative shift in time. The maximum correlation between $\bar{u}$ and u^* at each height and the corresponding phase angle is listed in Table IV.5.

comparing surface regression analysis with a QBO index from the wind on a single level with the one derived from EOFs as will be presented in Section V.2.2. The reason to construct a new index in the first place was to take into account the QBO state in the whole tropical stratosphere, therefore having a measure of the shape of the profile at all heights; cutting it back to one most important level is contradictory to this aim. Here the phase shift is interpreted as a projection onto new EOF axes as described above and thus to a most influential QBO configuration, or vertical profile, in the stratosphere.

Another interpretation for φ is as a shift in time. Given that the average cycle (2π) lasts for about 28 months, a step of 10° is equivalent to a timeshift of $\frac{28}{36}^{\text{th}}$ months (~ 0.8 months). Figure IV.18 shows the correlation between the measured wind at the indicated height levels (1956-2015) and the reconstructed EOF wind index u^* . The highest values exceed 0.9 for all but the lowermost (70hPa) level, so that the identification of the $\bar{u}$ at a certain level with u^* at a specific phase shift φ is certainly justified. Table IV.5 lists the maximum values of the correlation $\text{Corr}(u, u^*)$ as well as the corresponding phase angle. Further, one can identify the descending shear zones as the phase shift can be equated to elapsing time - it is important to note that a positive φ actually relates to a negative shift in time (clockwise rotation in phase space). Taking for example $\varphi = -30^\circ$: the timeseries u^* is highly correlated with the wind at 50hPa, so one could identify e.g. a strong surface response at that phaseshift (section V.2.2) with 50hPa being the most influential level, like Baldwin & Dunkerton [1998]. However, following the high correlation upwards, it could as well be a lagged response by 110° (~ 8.5) months to the wind at 20hPa.

Height (hPa)	φ_{max} ($^{\circ}$)	$\max(\text{Corr}(u, u^*))$
70	-72	0.79
50	-31	0.92
40	-5	0.96
30	31	0.96
20	71	0.97
15	93	0.98
10	119	0.94

Table IV.5.: Phase angle φ at which the correlation between equatorial zonal wind $\bar{u}$ and the reconstructed timeseries $u^*(\varphi)$ is maximised, as well as the maximum correlation value.

IV.6. Chapter summary

A set of metrics to characterise the morphology of the QBO was developed (Section IV.1). The equatorial stratosphere from models from the CMIP5 and CCMVal-2 intercomparison projects that internally produce a QBO were analysed using these metrics and their representation of the QBO compared to reanalyses and the FU Berlin zonal wind observational dataset (Section IV.3).

Models are generally good in representing the differences in amplitudes of easterlies and westerlies, as well as the period range and variability of the QBO. Common weaknesses include the upward shift of the QBO, westerlies descending too fast and a too narrow structure in latitude. Some of the discrepancies between models and reanalyses/observations are due to the vertical resolution of the models, which is coarser than the observations (the peak level in the observations, 15hPa, is not a common model level). Another issue is the representation of the gravity waves and their dissipation in the atmosphere, which drives the QBO; these processes have to be parametrised and their effect thus depends on parameter choices and resolution.

Methods to define some of the metrics using the first two EOFs and their associated principal component timeseries were presented and the FU Berlin dataset analysed (IV.5). Defining a QBO timeseries u^* from the first two EOFs is an alternative to identifying a state using one single pressure level.

V. Surface impacts

An important reason to study the QBO, its driving mechanisms and interactions is not only for scientific curiosity, but the hope for improving forecasts, especially long term predictions [Scaife *et al.*, 2014]. One example is the possible connection between the QBO and the Northern Hemisphere (NH) winter polar vortex and NAO, and subsequently the wintertime weather in Europe (warm/wet or cold/dry). Another potentially useful teleconnection is thought to exist between the tropical ocean, troposphere and stratosphere, making it possible to improve predictions of the MJO and the Indian Monsoon when taking the QBO phase into account (e.g. Claud & Terray [2006] and Yoo & Son [2016]), though Read & Castrejón-Pita [2012] suggest that this connection is spurious and due to separate synchronisation of the tropospheric biennial oscillation and the QBO with the semi-annual cycle.

V.1. History

Holton & Tan [1980] were among the first to suggest that the QBO might modulate wave propagation from the tropics to the poles. Their study diagnosed a strong influence on the 50hPa geopotential height anomalies and a weaker one on the 1000hPa (near-surface) anomalies. Numerous studies have lent further support of a QBO influence on the polar vortex [Naito & Hirota, 1997; Hu & Tung, 2002], sudden stratospheric warmings [Holton & Tan, 1982] and surface pressure, but some have challenged the exact nature of the Holton-Tan effect [Naoe & Shibata, 2010; Garfinkel *et al.*, 2012]. Anstey & Shepherd [2014] give an overview of the impacts of the equatorial stratosphere on the extratropics. Figure II.11 shows the original analysis by Holton & Tan [1980] and the updated version by Anstey & Shepherd [2014], examining the difference in geopotential height at 1000 hPa between QBO West and East years.

This study updates some of the previous QBO teleconnection studies and assesses the impacts using new methods. In the first part of this chapter, the analysis of an observational dataset will be presented, the second part compares the findings to a 505 year Control run analysis. The focus

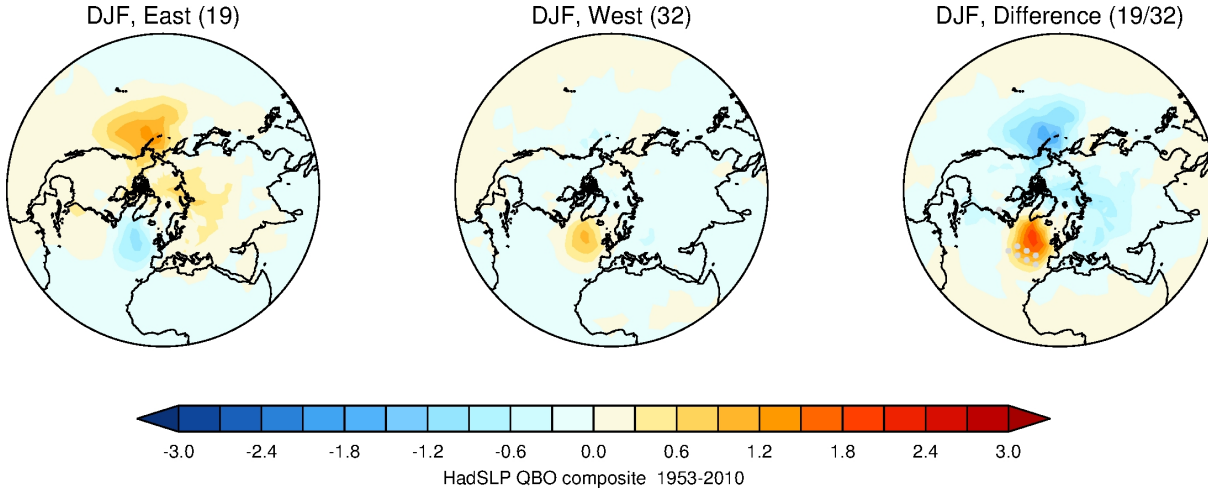


Figure V.1.: MSLP anomaly (in hPa) composite analysis of the Northern Hemisphere winter season. As indicator of the QBO phase, the equatorial $\bar{u}$ at 50hPa is taken. Grey dots in the difference plot indicate significance at the 95% level as obtained by a bootstrapping test. Note that these composites show anomalies, not absolute values (in contrast to Figure II.11). The mean climatology and standard deviation in DJF is shown as a comparison in Figure V.2. Numbers indicate the number of years within each composite.

will be on mean sea level pressure in the Northern Hemisphere winter (December-February).

V.2. Impacts in data

V.2.1. Composite surface analysis

The classical QBO teleconnection assessment [Holton & Tan, 1980; Anstey & Shepherd, 2014; Scaife *et al.*, 2014] defines a QBO “west” and “east” phase, usually via the wind at 30hPa or 50hPa before looking at the mean differences between the two states during winter (December, January, February), using a t -test to attribute significances. For completeness and for subsequent comparison with new approaches, this method was used in an initial assessment and results are shown in Figure V.1. The DJF climatology of observed mean sea level pressure is shown in Figure V.2 as reference.

However, this traditional method has a caveat; it arbitrarily uses a single pressure level to define the QBO phase. This can be overcome by using an index derived from the first two empirical orthogonal functions for $\bar{u}$, as described in section IV.4, which takes the vertical structure of the QBO into account as well. Another issue involves seasonal means. The polar vortex, which is thought to be influenced by the QBO, and itself is thought to influence the surface, is most variable during wintertime; these monthly variations will be obscured by seasonal averaging. Furthermore, the timing of the final warming has an impact on the surface [Hardiman

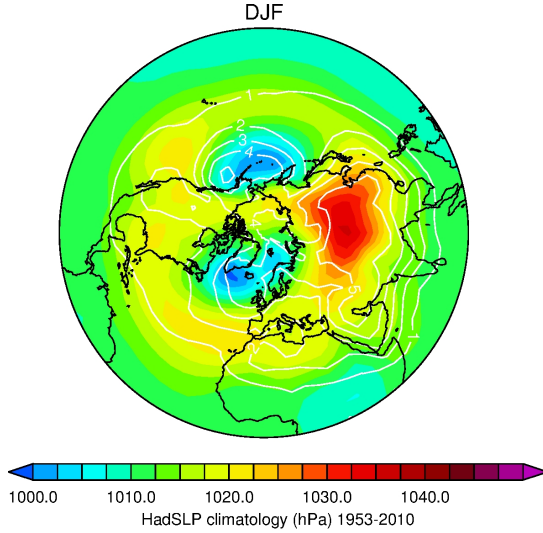


Figure V.2: Mean sea level pressure climatology for the winter seasonal mean. Lines indicate standard deviations of 1, 2, 3, 4, 5 hPa.

et al., 2011], and will also get smoothed out.

This problem can easily be dealt with by looking at single months (Figure V.3). In the composites, the QBO phase is defined by the 50hPa wind u (FU Berlin dataset) being $\leq -5\text{m/s}$ (East) or $\geq 5\text{m/s}$ (West) in the respective month. The significance of the values being different from a randomly chosen sample within the single composites is tested using a bootstrap test, while the significance of the difference is assessed with a t -test. In December, there is a strong Siberian signal, whereas in January the pressure difference shows a significant NAO signal. In February, the sign of this is reversed and it becomes less significant, however, a significant pressure anomaly over the North Pacific is evident. These strong features get smoothed out when taking the seasonal mean (Figure V.1), especially as the sign of the NAO pattern (Figure III.7) reverses from January to February. These differences between the single months make it obvious that seasonal averaging is not adequate for analysing the QBO-surface influence. As the North Atlantic and Pacific regions have high subseasonal variability (Figure V.4), significance in the composites indicates that the QBO does have an influence on the surface pressure pattern.

Sorting the QBO state by the wind at one pressure level and only into two different states has some shortcomings. Figure V.5 shows the vertical profiles of equatorial zonal wind for the years that get classified as “East” and “West” when using the 50hPa timeseries as an index. The profiles show a large variance regarding the height of the phase change as well as the maximum amplitude. As the position of the zero wind line, which is thought to influence wave propagation into the extratropics [Holton & Tan, 1980], differs according to the height of the phase change, a further distinction of QBO phases is made into “pure” QBO states and (much more common) “sheared” states. Pure states refer to when the equatorial winds are of the same sign in a

QBO state	Phase angle (°)
Sheared East over West	$\psi > 88$ and $\psi < -155$
Pure East	$-155 \leq \psi \leq -93$
Sheared West over East	$-93 < \psi < 26$
Pure West	$26 \leq \psi \leq 88$

Table V.1.: The range of phase angles for identifying four different QBO states. In the “pure” states, the respective shear zone is dominant in the height range between 10 and 70hPa, whereas in the sheared states the upper levels are dominated by winds of one sign and the lower levels by winds of the opposite sign.

deep layer, defined by a phase change occurring above 15hPa or below 50hPa; sheared states are defined by a phase change between 15hPa and 50hPa. These states are summarised in Table V.1, together with the corresponding phase angle according to Table IV.4.

Figure V.6 shows the composite mean state based on this selection, Figure V.7 the corresponding January equatorial zonal wind profiles. The analysis suggests that much of the composite difference in January comes from the difference between years with “pure” equatorial East or West states. Thus, a further distinction of the QBO states should be made as normally the East over West and West (West over East and East) states are combined into one composite when taking only the 50hPa level into account (Figure V.5).

V.2.2. Regression surface analysis

In Earth’s very interactive climate system, the QBO is not the only factor influencing surface pressure. However, in simple composite analysis, other covarying factors are not taken into account and the signal might actually be obscured by other forcings. One possible way of circumventing these factors is to use regression analysis, with different explanatory variables x_i used to reconstruct the observed MSLP. Here, apart from a QBO index, indices accounting for the 11-year solar cycle influence, volcanic eruptions, El Nino and rising CO₂ levels are used in a multilinear regression.

Figure V.8 shows the scaled regression coefficients for MSLP derived during NH winter, using the 50hPa observed equatorial u as the QBO index. The patterns should be comparable to Figure V.3, but depicting the QBO influence more accurately (as other possible influences are accounted for as separate explanatory variables). Indeed, the Siberian low in December, the NAO signal in January and the more Annular Mode like pattern (compare to the first EOF, Figure III.7) in February are significant features in the regression analysis as well, and even more significant in late winter compared to the composite analysis. This indicates that the QBO does

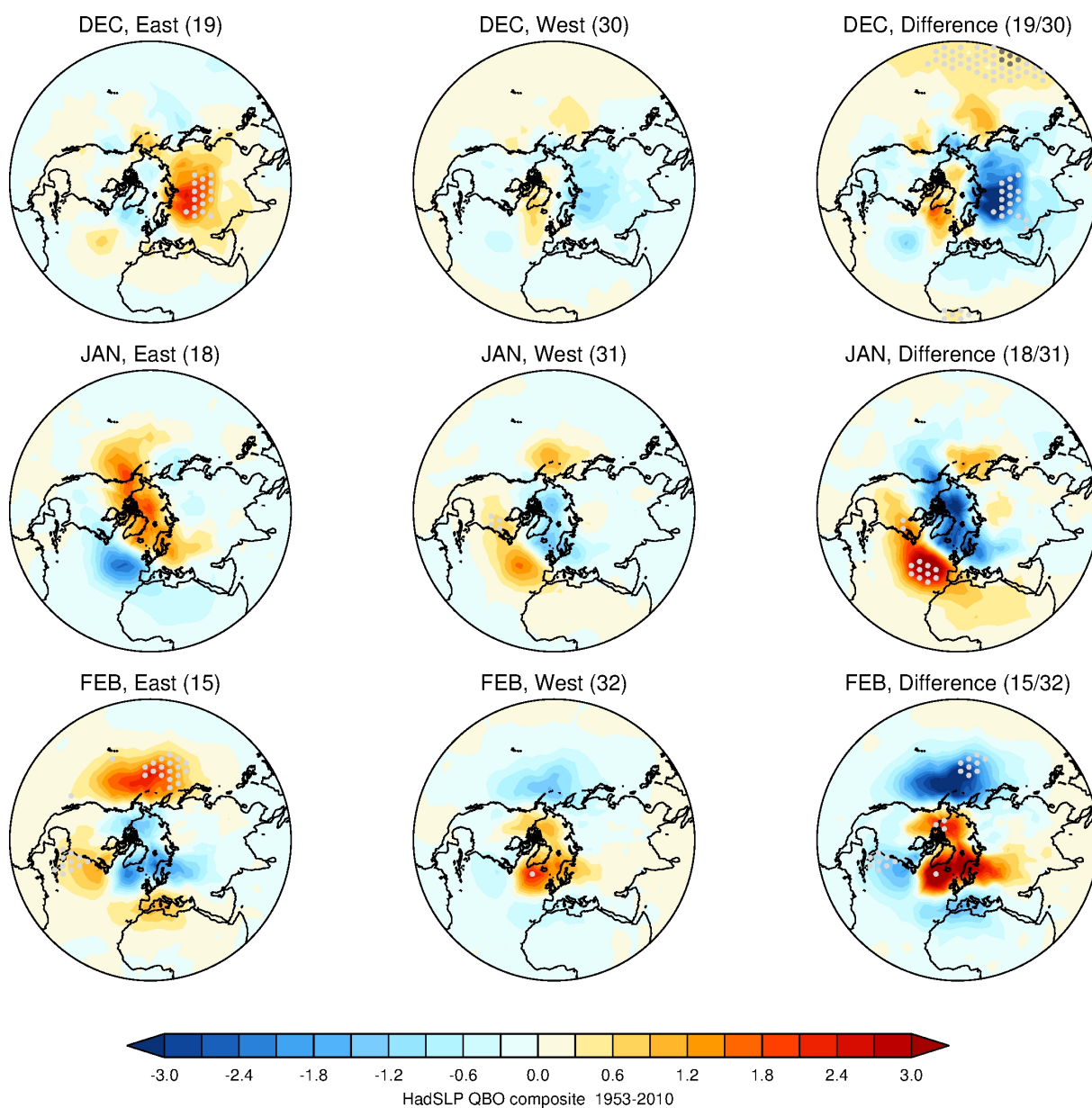


Figure V.3.: Composite analysis of single winter months of the Northern Hemisphere mean sea level pressure anomalies (in hPa). Stippling indicates significance at the 95% (light grey) and 99% level (dark grey).

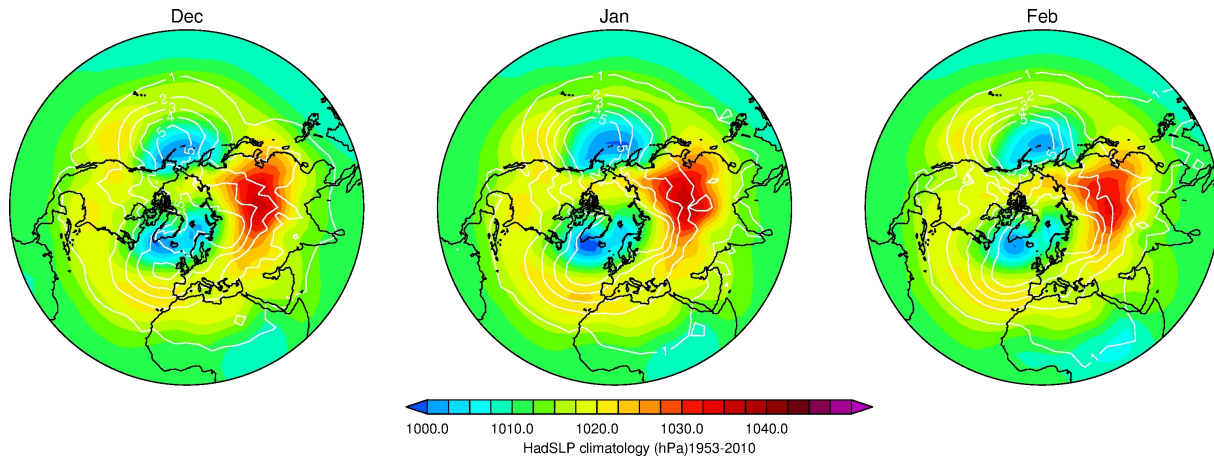


Figure V.4.: MSLP climatology in the winter months. Lines indicate standard deviations of 1, 2, 3, 4, 5 hPa. For comparison with the seasonal mean in Figure V.2.

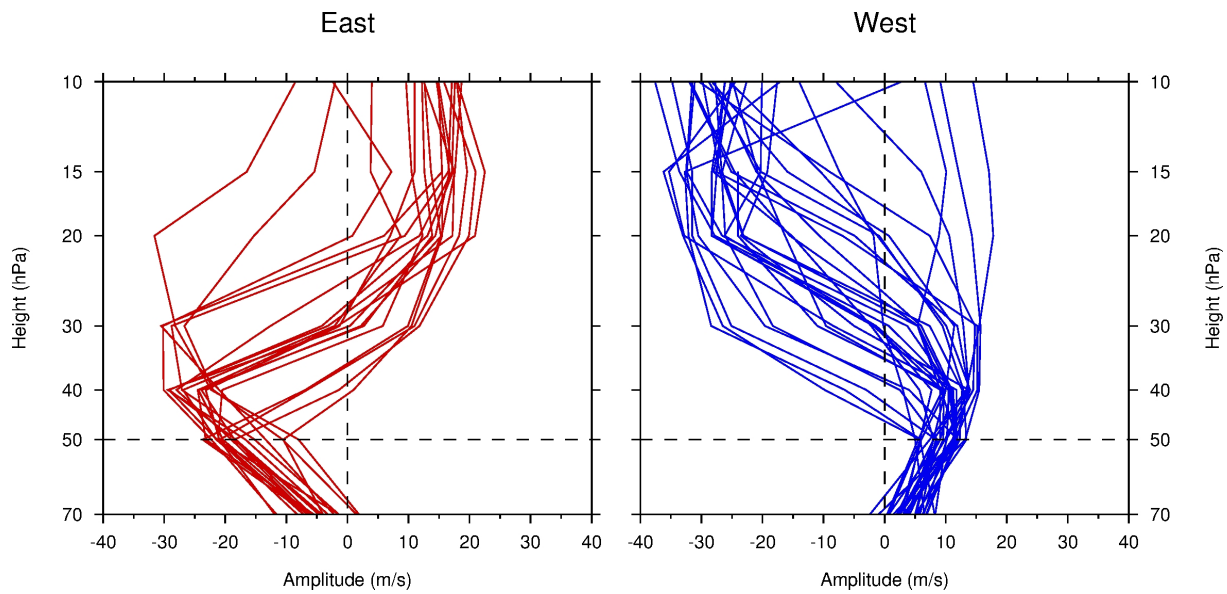


Figure V.5.: January QBO profiles for years that are classified as “East” and “West” according to the amplitude of the wind at 50hPa. These are the years selected for the composite in Figure V.3.

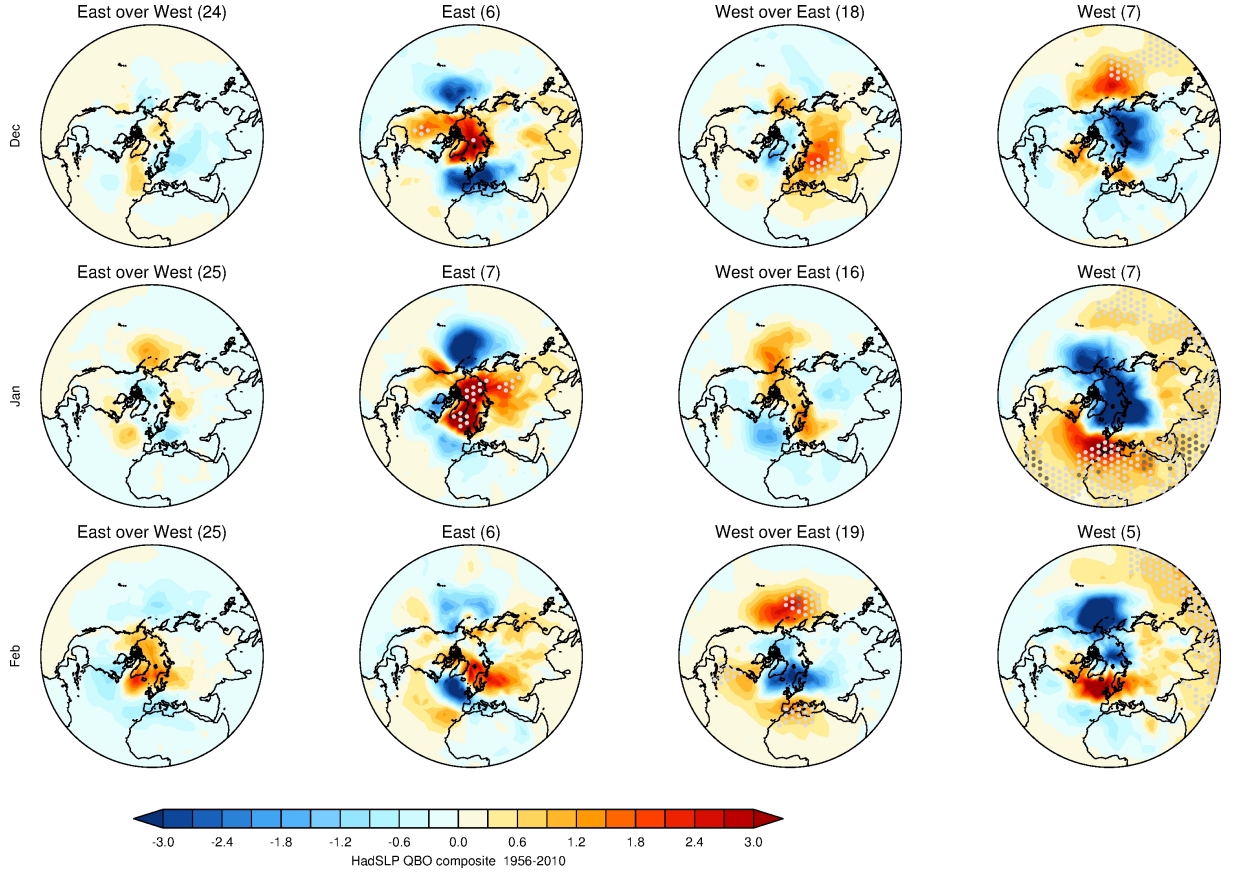


Figure V.6.: NH MSLP anomaly (hPa) composites in December, January and February. Stippling indicates significance at the 95% (light grey) and 99% significance as obtained by a bootstrap test. Numbers indicate the number of years within each composite. Compare to Figure V.3. The usual approach of defining the QBO at 50hPa combines the East over West and West state (first and fourth column) into one composite, as well as the West over East and East state (third and second column) into a second composite.

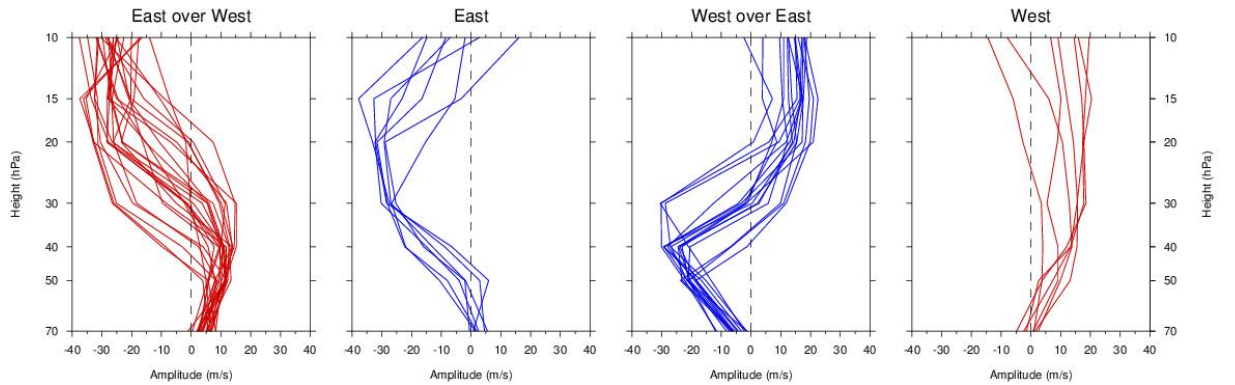


Figure V.7.: January QBO profiles for years that are classified as “pure” or “sheared” states according to the EOF phase angle (Table V.1). Profiles correspond to the composites in Figure V.6.

have an influence on the surface, especially in January and February. It is worth noting that the sign of the NAM/NAO pattern in January and February is reversed, as in the composite analysis.

Figure V.8 also shows results of using different pressure levels as the QBO regression index. Baldwin & Dunkerton [1998] suggested that the 40hPa equatorial wind is the optimum level to characterise the QBO influence on the Northern Hemisphere polar vortex. In the surface analysis, the coefficients are higher and show larger significant regions when using the equatorial wind at 30hPa or 40hPa compared to 50hPa, especially in January; 30hPa has the larger response in January, whereas the 40hPa response is largest in February. Employing a QBO index level at 50hPa in the composite analysis follows previous studies, but these results do not necessarily justify this choice. Furthermore, it is unlikely that only one pressure level should be most influential in causing a surface response since the QBO dominates the equatorial stratosphere at all levels between 100 and 10hPa. Thus, the derived QBO index from the first two EOFs, $u^* = R \sin(\psi + \varphi)$ (definitions see Section IV.4) is implemented in the regression analysis instead of the wind at a single level. Figure V.9 shows the regression coefficients in January for phase shifts in the range of $-90^\circ \leq \varphi \leq 90^\circ$. Equivalent plots for December and February can be found in the Appendix (Figures B.2 and B.3).

The NAO signal maximises between $10^\circ \leq \varphi \leq 30^\circ$. This is slightly higher than the maximum polar vortex response found by Baldwin & Dunkerton [1998]. Referring back to Figure IV.18, the highest correlation for these φ values can be found at around 30hPa, in accordance with the surface response seen in Figure V.8. As mentioned in section IV.4, this can as well be interpreted by a certain QBO profile being most influential on the surface response. Given that a phase shift of $\varphi = 30^\circ$ is equal to a projection onto a basis function $\vec{e}_2^*$ rotated by 30° clockwise compared to the original second EOF axis (Figure IV.17), this corresponds to the profile seen in Figure IV.13 at $\psi = 60^\circ$. This profile, being one consistent phase throughout the stratosphere, also explains why the regression coefficient patterns for the wind at 50, 40 and 30hPa are very similar and just differ in amplitude. It also backs up the suggestion in section V.2.1 to sort the QBO into 4 instead of only 2 different phases with the “pure” phases being the most influential ones; it stresses the importance of looking at the QBO state over a greater range of heights instead of just one level.

To explore this four state analysis further, another multilinear regression is performed on the mean sea level pressure anomalies. The same explanatory indices $x_i(t)$ as before (solar, volcanic,

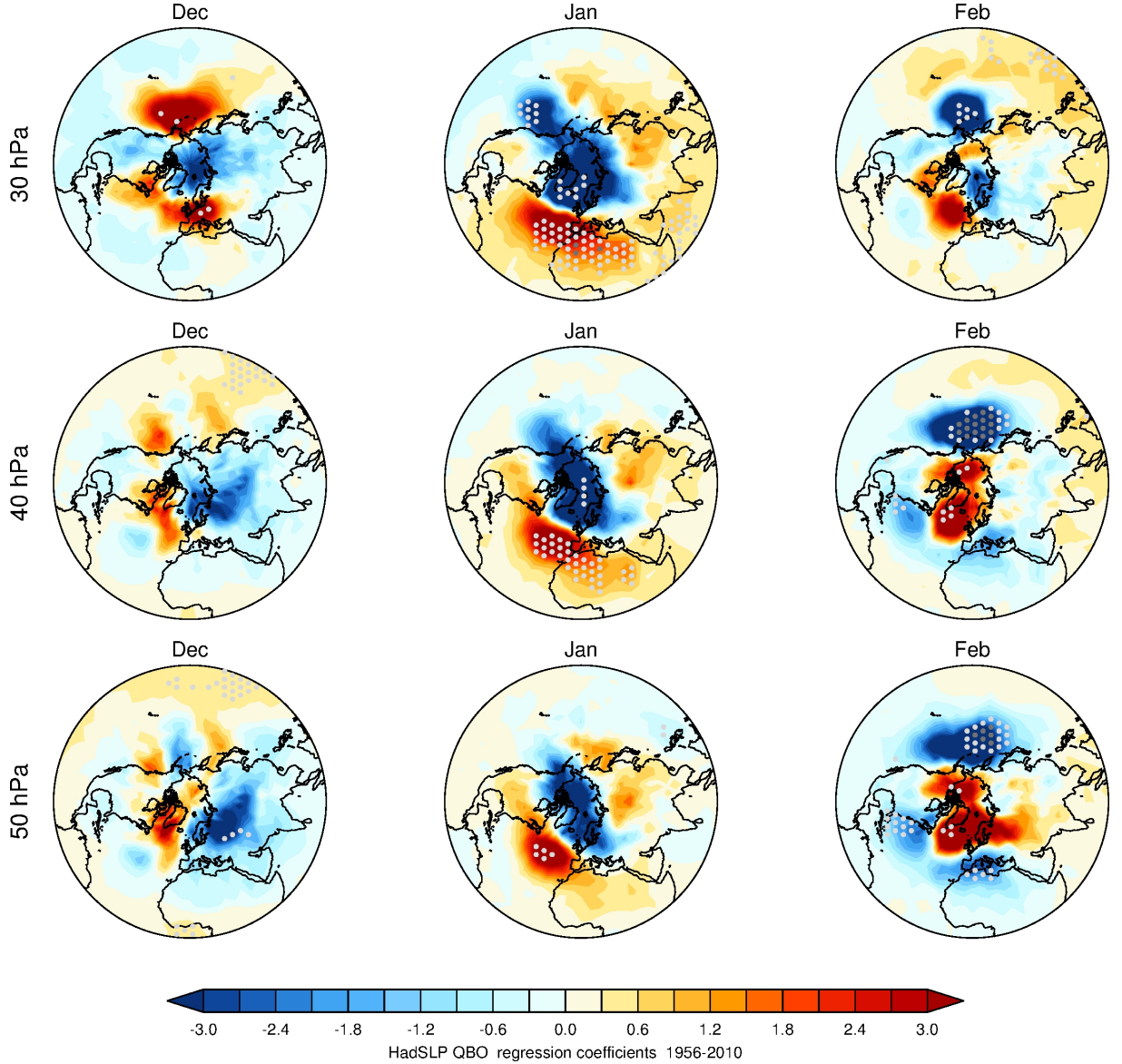


Figure V.8.: Scaled coefficients β_{QBO} (hPa) from the regression analysis $y(t) = \sum_i \beta_i x_i(t) + \epsilon(t)$ of MSLP (y) and predictors $i \in \{ 11\text{-yr Solar Cycle, Volcanos, ENSO, CO}_2, \text{QBO} \}$. Observed equatorial u at 30hPa (upper row), 40hPa (middle row) and 50hPa (lower row) is used as QBO index x_{QBO} . The patterns are comparable to the composite analysis, but other influences like the solar cycle and ENSO are regressed out. Significance at the 95% and 99% level as obtained by a t -test is indicated with light and dark grey dots.

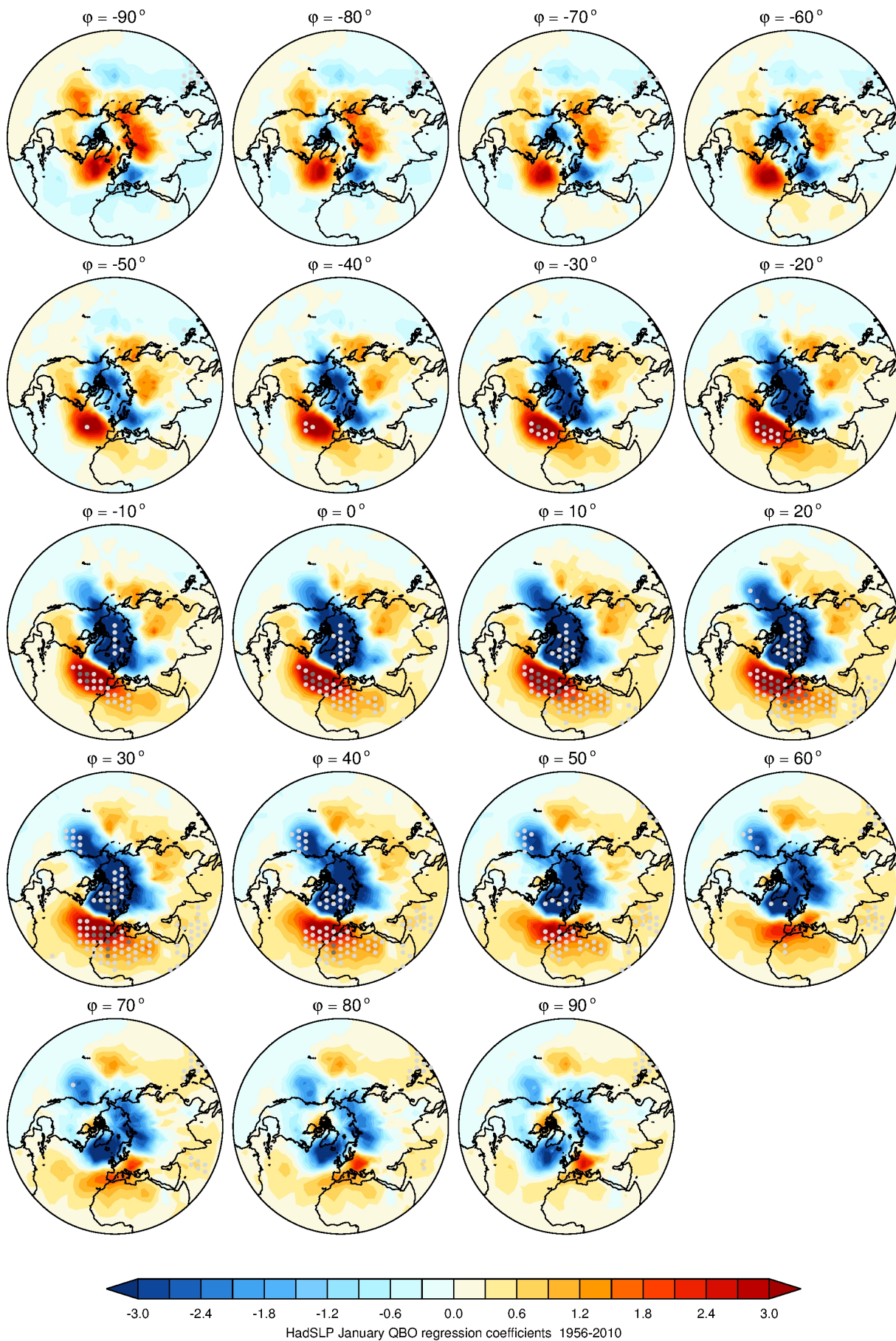


Figure V.9.: QBO regression coefficients in January, using u^* at various phase shifts φ as QBO index. By construction, the coefficients at $\varphi = -90^\circ$ and $\varphi = 90^\circ$ are equal and of opposite sign as the timeseries u^* used as regression index is equal and of opposite sign for a shift of 180° .

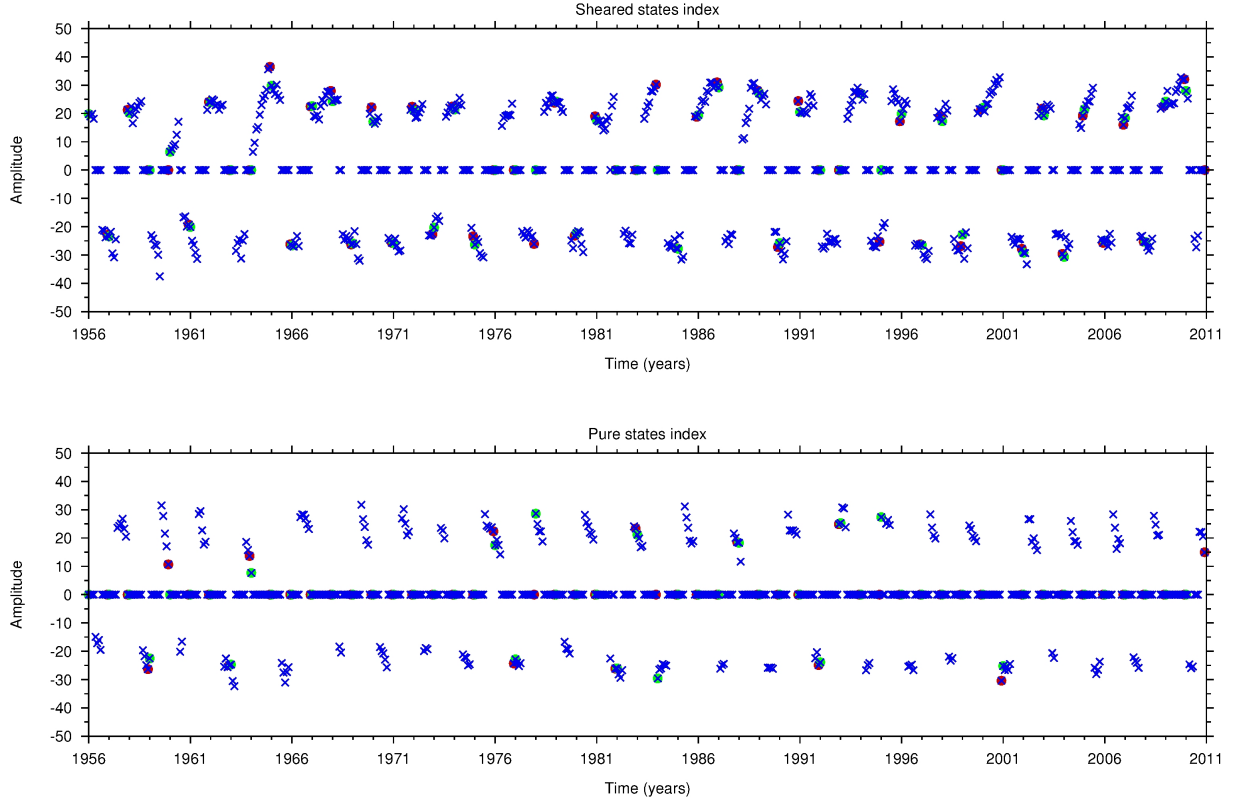


Figure V.10.: QBO timeseries for the regression analysis in Figure V.11. The upper series represents the “sheared” QBO states, the lower one the “pure” states. The red dots indicate December, the green dots January of each year. The latter are the years selected for the composite in Figure V.6.

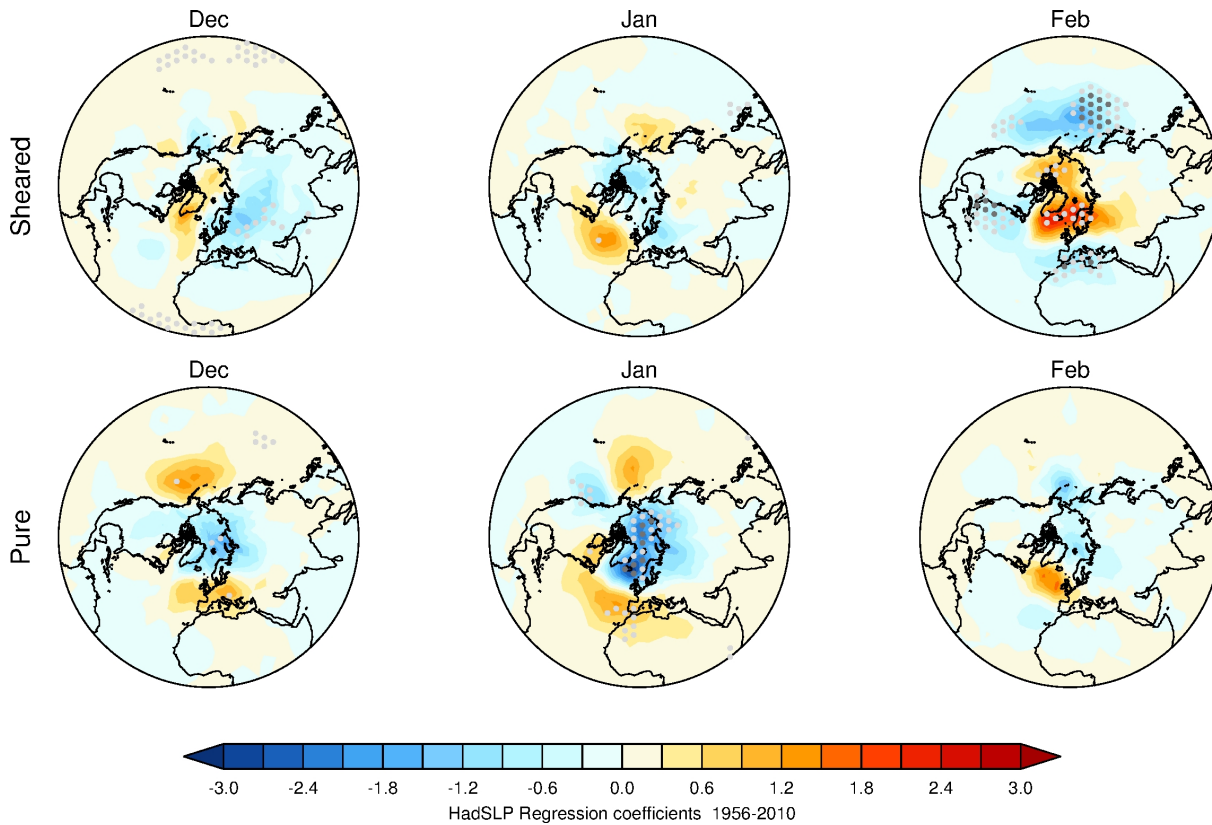


Figure V.11.: Same as Figure V.8, using two different indices based on the QBO phase angle to imitate the composite analysis of four different QBO states (Figure V.6).

CO₂, ENSO) are used; additionally, two QBO indices are defined from the phase angle based on Table V.1. The “pure” index is set to -1 for an East state and 1 for a West state. The “sheared” index is set to -1 for a West over East and 1 for an East over West state. To account for different amplitudes of the QBO, these indices are multiplied by the maximum absolute amplitude of the QBO between 10 and 70hPa at every point in time. Timeseries are shown in Figure V.10.

Defining two different indices to describe the four states leads to two patterns of regression coefficients that are expected to roughly represent the difference of the composited states in Figure V.6. Figure V.11 shows the scaled regression coefficients for the two indices in December, January and February. For the pure states, a clear NAO signal can be seen, which is strongest in January, whereas in the mixed states the contrast between the North Pole and the North Pacific in February is most prominent. The patterns in both cases are consistent with the composite analysis. Given that other known influential factors are regressed out, these findings support the hypothesis of the QBO having an influence on the Northern Hemisphere extratropics in winter.

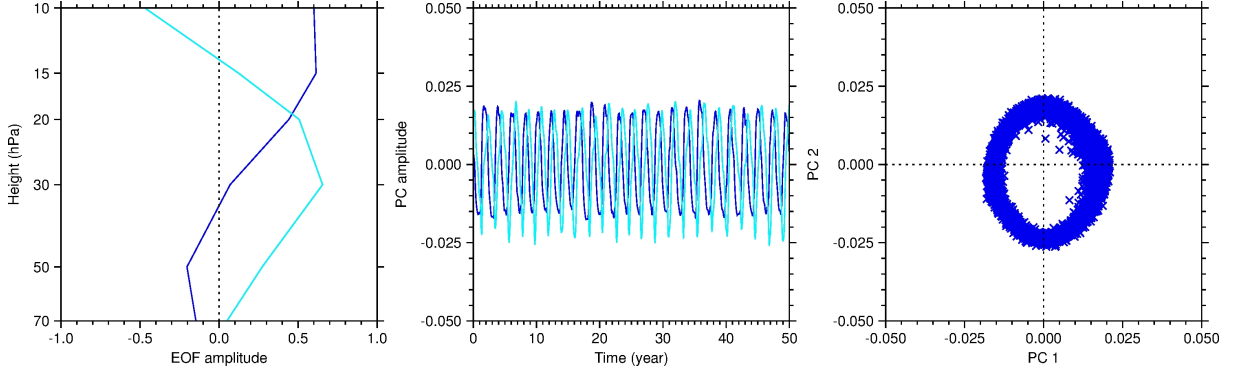


Figure V.12.: Left: The first two empirical orthogonal functions (blue EOF 1, turquoise EOF 2) of $\bar{u}$ in the equatorial mid-stratosphere from 505 years HadGEM2-CC Control run. The corresponding principal component timeseries are shown in the middle and their phasespace on the right. Together the first two EOFs explain 82% of the variance in zonal mean zonal wind.

V.3. Control run

One problem faced with the observational analysis is the limited availability of data. There have been regular QBO observations from 1953 onwards only, and the HadSLP dataset ends in 2012¹. This gives only about 60 years to work with, which is very little, especially in composite analysis where only a subset of the available years is utilised. This is one of the reasons the analysis above is repeated with 505 years of HadGEM2-CC’s Control run data. Another is that due to the setup of the Control run, there are no “external” influence factors such as the sun, ozone, CO₂ or volcanoes, so the natural variability and atmospheric dynamics can be assessed more reliably. However, only processes that are known can be represented in a model and some of these processes have to be parametrised due to their short temporal or spatial scale (e.g. gravity waves, section II.2).

To assess the structure of the QBO in the Control run, an empirical orthogonal function analysis analogous to the one the observational dataset was subjected to, is carried out. The first two EOFs account for 82% of the variance of the equatorial $\bar{u}$ between 10 and 70hPa, which makes the QBO by far the dominant mode of variability, even more than in the observations. The height structure of the first two EOFs is comparable to the one derived from the observations, as are the two principal component timeseries which exhibit a regular, roughly 2 year long oscillation, and are about 90° out of phase (Figure V.12). In contrast to the observations, the model Control dataset does not include a 40hPa level, which explains some of the difference in the EOF profiles. The QBO characteristics of the zonal mean zonal wind (as described for

¹There is a near-real time product available from <http://www.metoffice.gov.uk/hadobs/hadslp2/data/download.html> as well. However, the HadSLP2r dataset is not homogenous with the HadSLP2 dataset (years 1850-2004); the homogenous continuation of HadSLP is available only until 2012.

<i>ZM Zonal Wind</i>	Observations	
	Reanalysis	Control run
Height of maximum (hPa)	15.0	10.0
Fourier amplitude (m/s)	14.6	19.7
Latitudinal extent (°)	21.0	18.5
Vertical extent (km)	16.8	22.5
Depth (hPa)	87.6	84.8
Mean Period (months)	27.9	26.4
Amplitude Easterly	-23.2	-35.0
Amplitude Westerly	26.1	22.1
Descent rate Easterly	0.7	0.8
Descent rate Westerly	1.2	1.1

Table V.2.: HadGEM2-CC Control run characteristic parameters as described in Section IV.1. Observational values are given for comparison, if possible; otherwise (latitudinal extent, depth, vertical extent), the mean reanalysis value is given. The Control run lies well within the model range (Table IV.3) and is comparable to the observations.

the other models in Section IV.1) are summarized in table V.2 and lie well within the model range, not deviating significantly from the observations (Table IV.1), apart from the QBO peak being too high in altitude, which is a common model bias, probably related to gravity wave parametrisation (see Section IV.3).

The monthly means of MSLP in December, January and February are shown in Figure V.13 and can be compared to the observations in Figure V.4. The patterns are very similar, with lower pressure over Iceland and the North Pacific, a ring of higher pressure surrounding the Icelandic low and a prominent high pressure feature over Siberia. The Control run exhibits slightly too high pressure values, with higher variance as well. However, the difference from the observations is not significant, being consistent within one standard deviation of the observations in the relevant areas.

Given the similarities to the observations in equatorial zonal wind and mean sea level pressure, the Control run seems to be a valid dataset to examine the QBO-surface interaction. Different analyses, as described for the observations (sections V.2.1, V.2.2) are carried out to quantify the influence of the equatorial stratosphere at the surface. The classic composite analysis, shown in Figure V.14, shows some similarities with the observations, e.g. lower pressure over the North Pole in the West-East composite; however, there are quite a few differences, starting with the amplitude being much weaker. This is probably partly due to incorporating 505 compared to 65 years. However, given the similar climatology and the higher variance in the Control run, this

cannot fully explain the difference in the composites. Furthermore, the pattern in the Control run in February shows a pressure system over Scandinavia and the Aleutian region of opposite sign compared to the observations (Figure V.3). Nonetheless, there are significant regions of pressure anomalies in the different QBO states in the Control run, indicating that there is a connection between the tropical stratosphere and the extratropical surface pressure.

Regression analysis is used to explore the QBO teleconnection further. Apart from a QBO index, an ENSO index is calculated from the sea surface temperature anomalies, averaged over the Niño 3.4 region. Figure V.15 shows the Control run analogue to Figure V.8, using equatorial $\bar{u}$ at 20hPa, 30hPa and 50hPa as the QBO index². The regression shows a stronger response than the composite analysis, but reduced areas of significance compared to the observations. Again, the patterns differ as well, with no strong NAO in the Control run; instead, the coefficients in February are significant over Scandinavia when using 50hPa as QBO index. Using pressure levels in this comparison, however, is a bit ambiguous, as the height structure of the QBO in observations and the Control run differ (Table IV.3).

Using the reconstructed index u^* (Figure V.16) is consistent with employing a single level regression analysis. As with the composites, there is no strong NAO signal, but rather a contrast between the Siberian Sea and the Aleutian region, especially in the φ range that gave the strongest signal in the observation analysis. Equivalent plots for December and February can be found in the appendix (Figures B.4 and B.5). The differences in January between the Control run and the observations can also be seen when using the “4 state” regression analysis that distinguishes between QBO “pure” and “sheared” states (Figure V.17). The patterns are not identical with the ones seen in observations. However, the finding of the signal being most significant in January for pure states and in February for the sheared states still holds. In observations (Figure V.11), the area exhibiting significant QBO regression coefficients for pure states in January is the NAO region, whereas in the control run, the contrast between the North Pole and the Aleutian region is emphasized. For the sheared states in February, the significant region in the control run is shifted from the North Atlantic over to Siberia compared to the observations.

Given the different patterns seen between the Control run and the observations when assessing QBO influence, an EOF analysis of the MSLP anomalies in the winter months, identical to the one shown in Figure III.7 is done. Figure V.18 shows the pattern of the first EOF in the

²This is due to the lack of the 40hPa level in the model dataset and also helps to account for the QBO being shifted upwards.

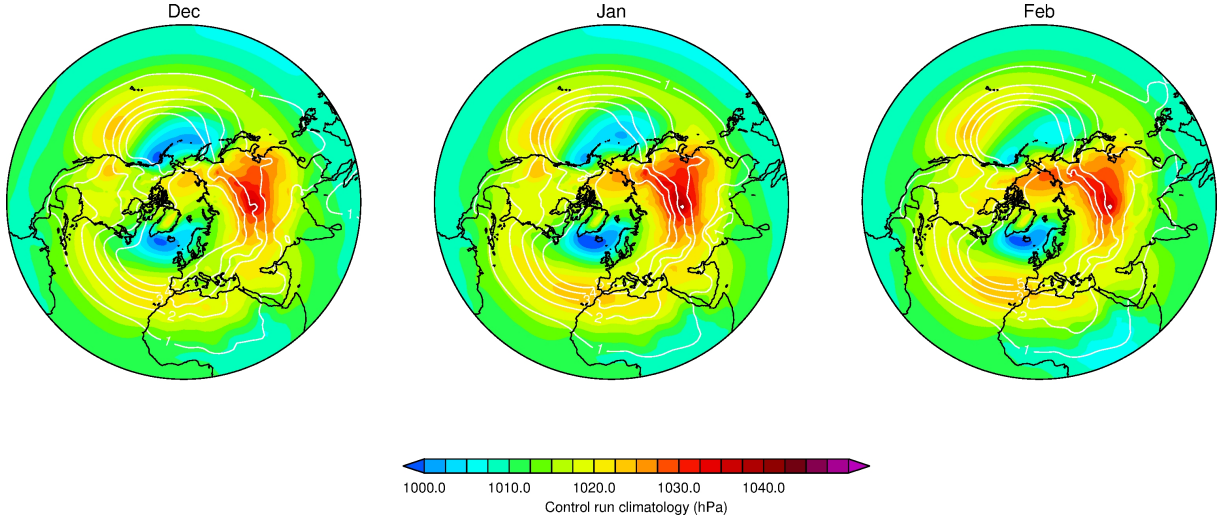


Figure V.13.: Monthly mean sea level pressure and standard deviation (white lines) for the HadGEM2-CC 505 years Control run.

Northern Hemisphere and its associated timeseries. Compared to the first EOF in observations, the amplitude is weaker in the Control run and it does only account for 10% of the total variance, which might explain the smaller amplitude in the composite analysis. The pattern itself is similar to the one found in the HadSLP dataset and shows a typical NAO structure and the pressure system over the Aleutian region (Figure III.7).

V.4. Chapter summary

The impact of the stratospheric QBO on NH mean sea level pressure in January was investigated using composite and multilinear regression analyses. Similarly to previous studies [Holton & Tan, 1980; Anstey & Shepherd, 2014], the QBO state (easterly or westerly at 50hPa) was found to have a significant influence in the NAO region (Figure V.3). The NAO pattern becomes even more pronounced when applying regression analysis, accounting for other factors like the 11-year solar cycle, ENSO, Volcanoes and rising CO₂ levels (Figure V.8).

The regression analysis was taken one step further by using an index $u^* = R \sin(\psi + \varphi)$, derived from the first two EOFs as QBO index (Figure V.9), enhancing the QBO signal in the NAO region even more. The angle φ for which the NAO pattern maximised in amplitude and significance is most closely correlated with the wind between 30hPa and 40hPa (Table IV.4).

Both composite and regression analyses suggest that the common approach of characterising the QBO state by one wind at one single level is not sufficient; rather, a classification into four states (Figures V.6, V.11 and Table V.1), taking into account the wind profile in the stratosphere,

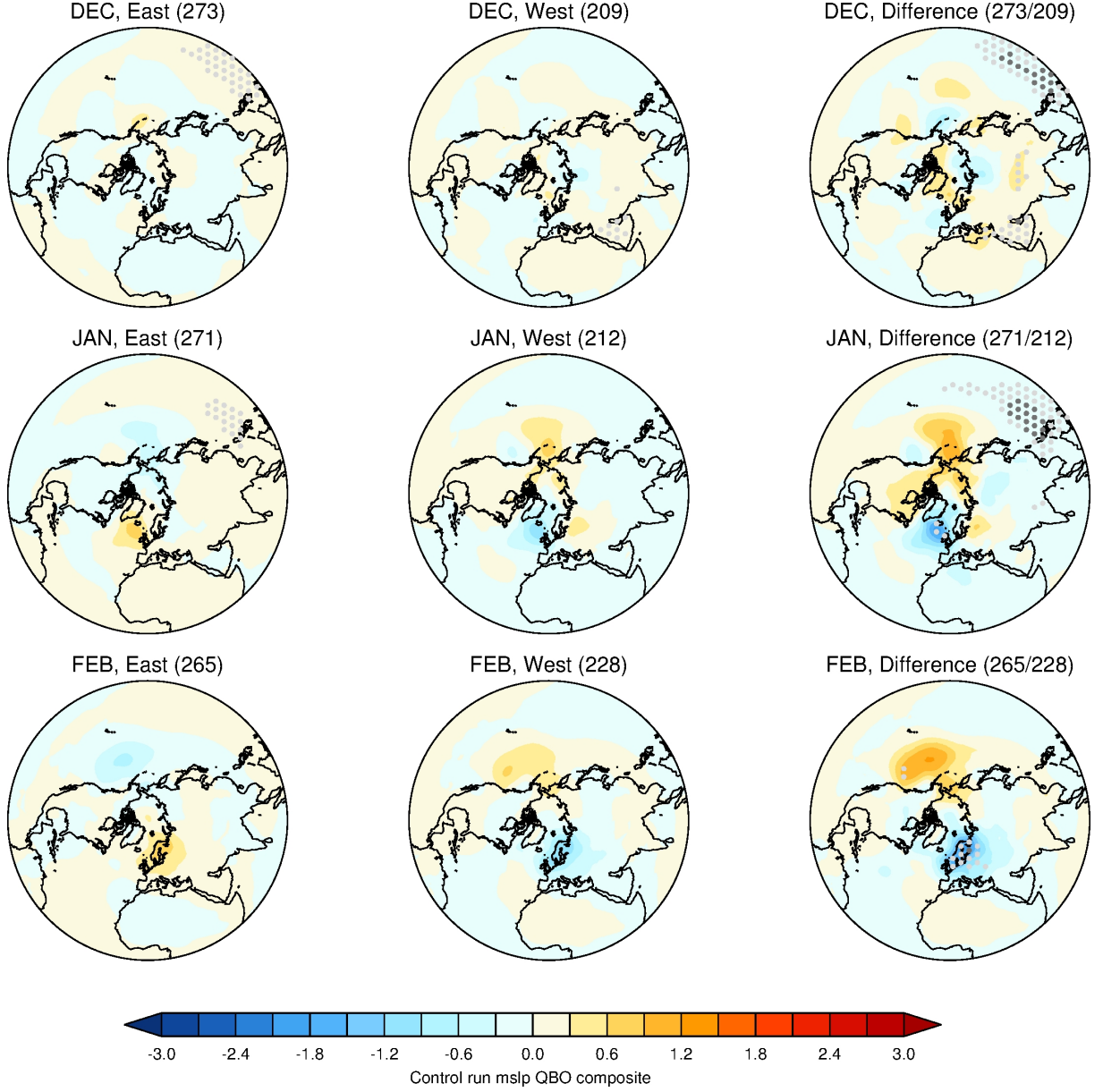


Figure V.14.: Classical composite analysis of the 505 years Control run mean sea level pressure. The two QBO states are defined using the zonal wind at 50hPa, with “East” months requiring $\bar{u} \leq -5\text{m/s}$ and “West” months $\bar{u} \geq 5\text{m/s}$. Dotting indicates significance at the 95% and 99% level; in the single composites, significance was tested with a bootstrapping test, in the composite with a t -test.

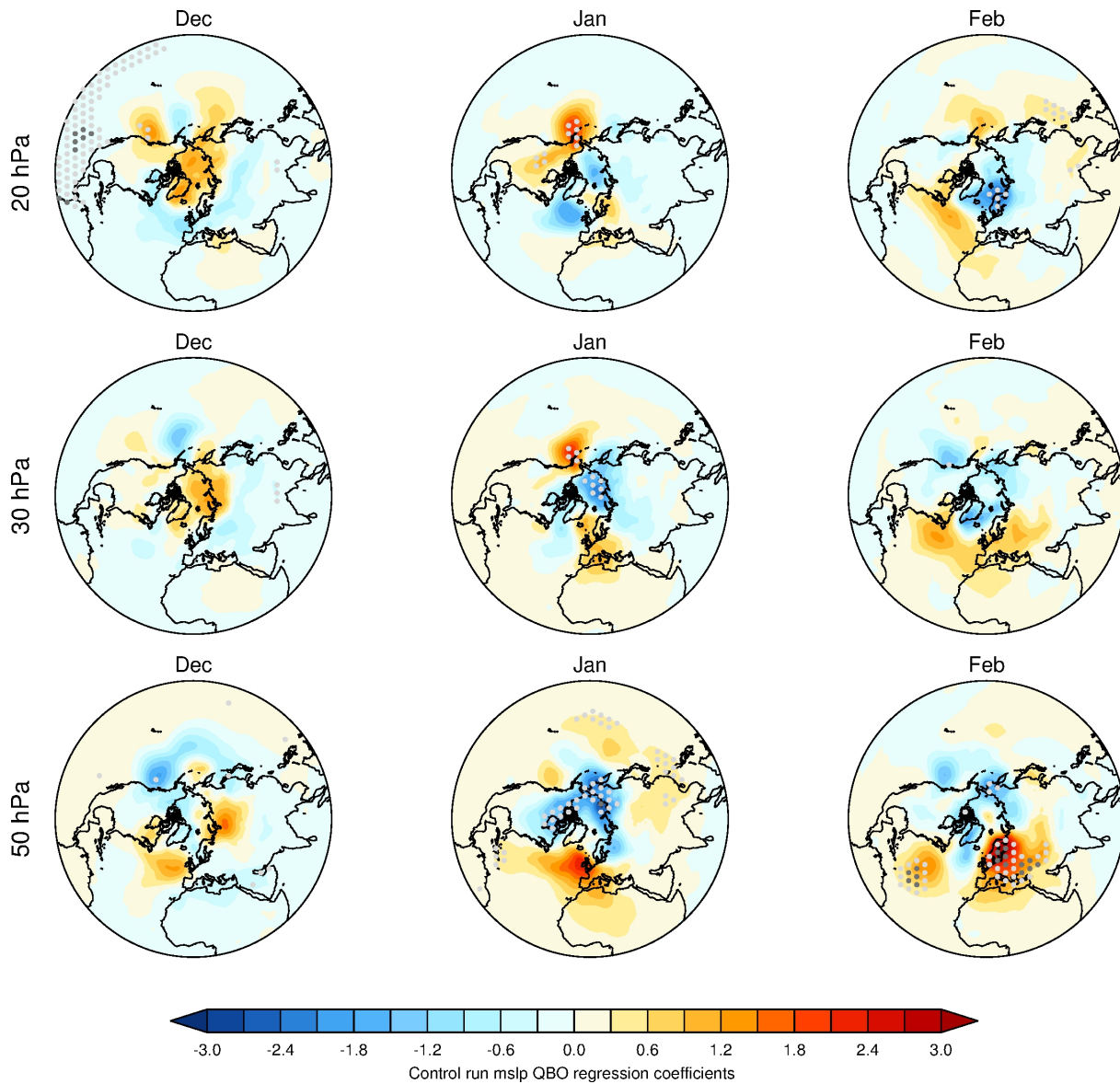


Figure V.15.: Scaled QBO regression coefficients of HadGEM2-CC 505 year Control run mean sea level pressure. The equatorial $\bar{u}$ at 20hPa (upper), 30hPa (middle) and 50hPa (lower) was used as QBO index. Dotting indicates significance at the 95% and 99% level as obtained by a *t*-test. This figure is analogous to Figure V.8, but using only a QBO and an ENSO index in the regression analysis.

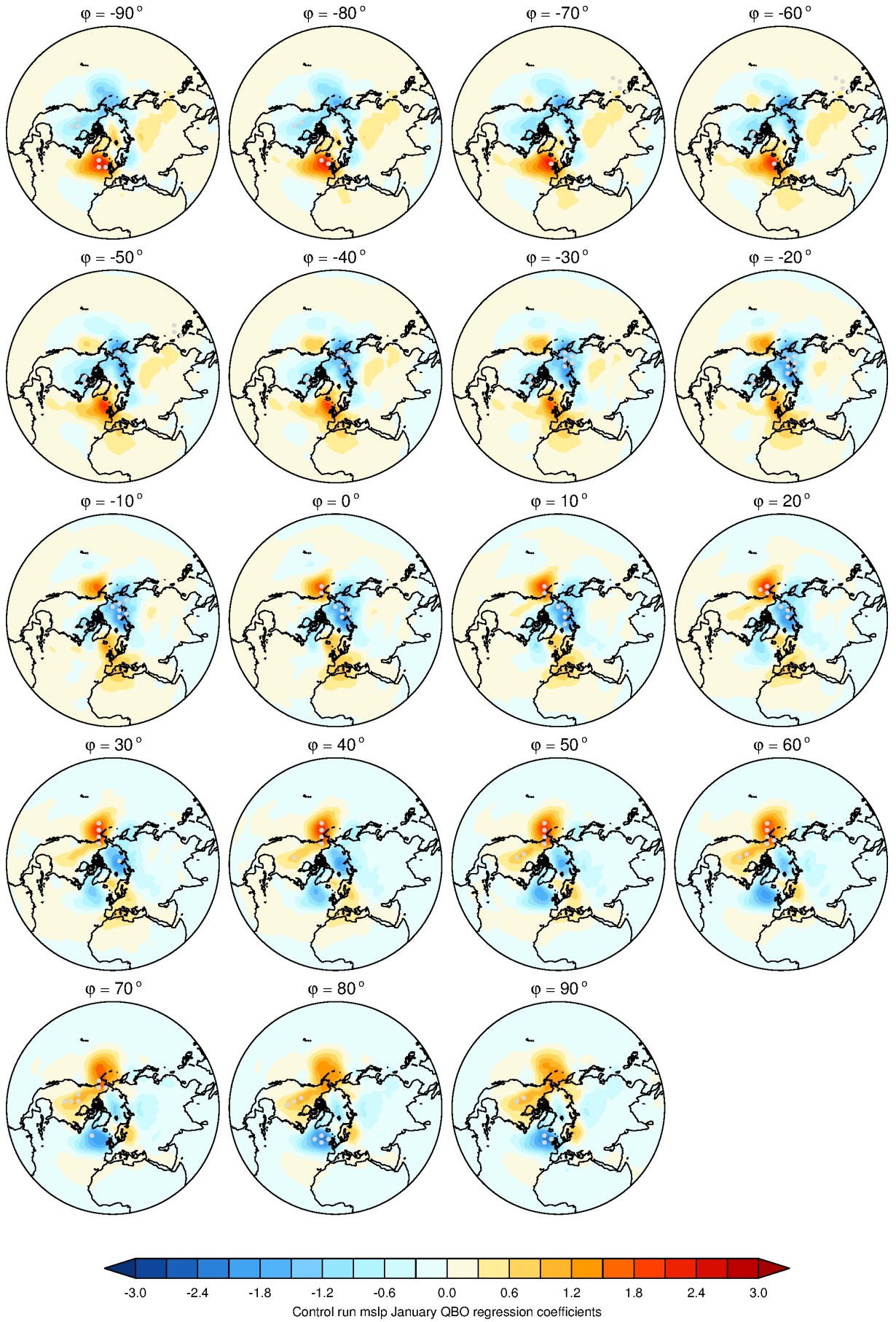


Figure V.16.: Same as Figure V.15, but using the reconstructed index u^* and for January only. Compare to observational analysis in Figure V.9.

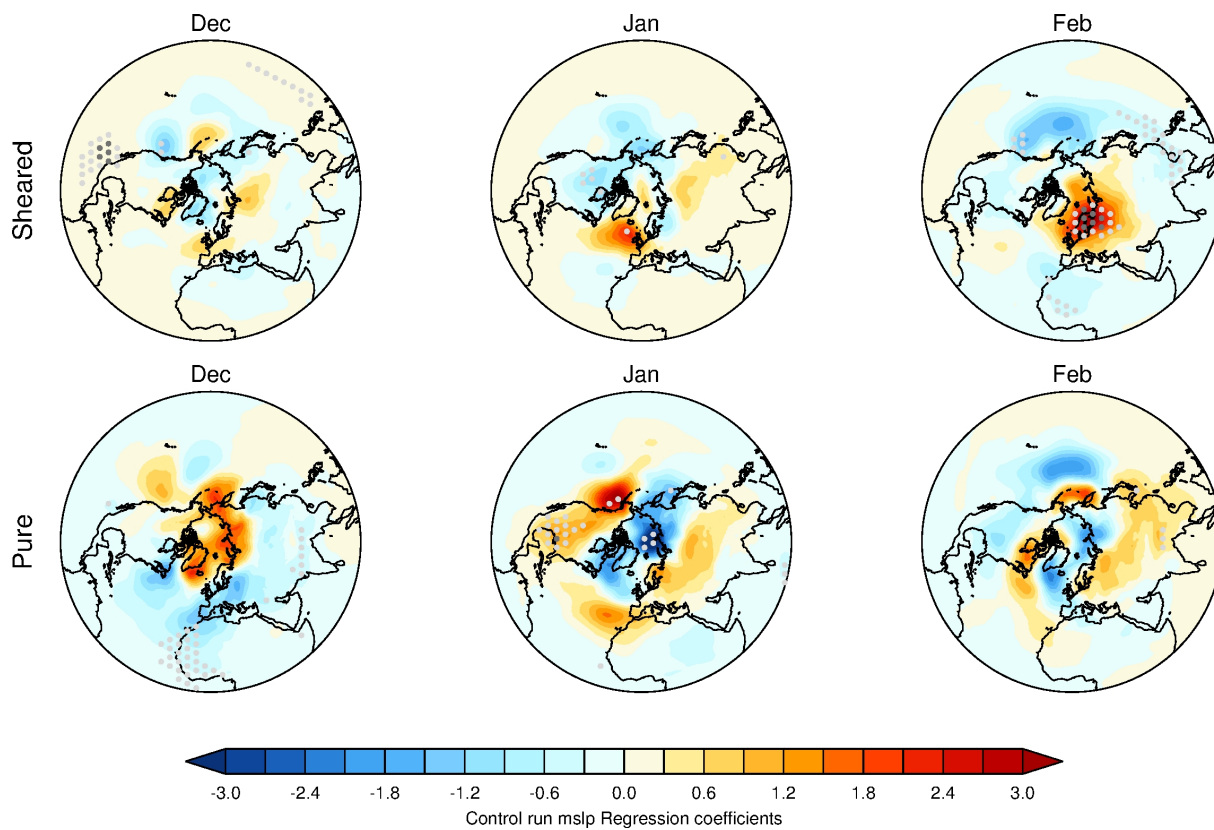


Figure V.17.: Same as Figure V.15, using two different QBO indices of “sheared” and “pure” states as for Figure V.11.

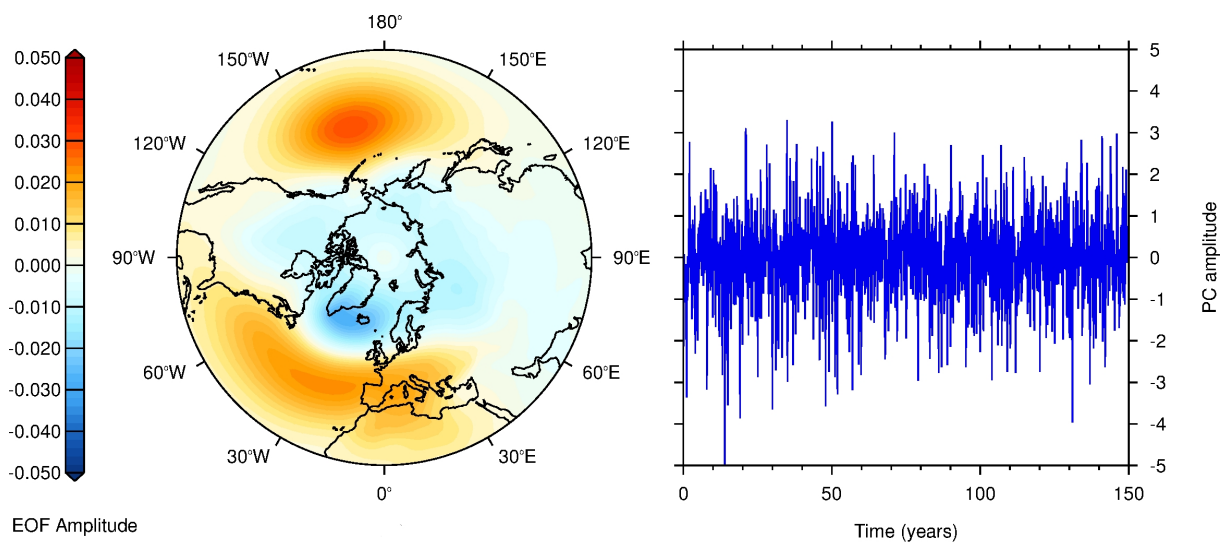


Figure V.18.: The first empirical orthogonal function (left) and its associated principal component timeseries (right) from the mean sea level anomalies in the Control run.

is found to be more useful when assessing surface impacts.

As the observations are limited by the short QBO record, similar analyses were carried out for the 505 year HadGEM2-CC control run. The main findings, such as the QBO having an impact on NH winter MSLP, as well as the usefulness of a 4-state QBO classification, are confirmed in the model run. However, the QBO signals are weaker and the patterns shifted compared to the observations. The reasons could be general shortcomings of the model regarding the variability of MSLP, the representation of the QBO or the relevant interaction mechanism. It is also possible that the recent 60-year period is not representative for the QBO-NAO interaction and that a longer dataset would not show a significance alteration of MSLP by the state of the equatorial stratosphere.

VI. Influence pathways

VI.1. Background

Given that regression and composite analyses of observations show a connection between the QBO state and the mean sea level pressure, how exactly might the QBO influence the extratropical surface? Holton & Tan [1980] proposed that it is the QBO state changing the background wind, resulting in altered propagation of planetary scale waves, which then influence the polar vortex. However, the correlation between QBO and vortex state is not consistent over time [Lu *et al.*, 2008] and the interaction mechanism has been questioned, e.g. by Garfinkel *et al.* [2012]. The latter suggest that the QBO surface influence happens through the change of the mean meridional circulation rather than a zero wind line modulation. Another interaction mechanism, suggested by Simpson *et al.* [2009], is that tropospheric eddy momentum fluxes change in response to lower stratosphere temperature anomalies. This second mechanism can be explored via an analysis of the zonal and meridional wind field, similar to the surface analysis.

Figure VI.1 shows the monthly mean zonal mean states of zonal wind, temperature and circulation in the NH winter months from the ERA-Interim dataset, 1979-2009. The equatorial stratosphere exhibits a climatological mean easterly state. In the troposphere, the westerly subtropical jets in each hemisphere can be seen. The polar night jet is strong in the Northern Hemisphere. The middle row shows the residual mean circulation ($\bar{v}^*, \bar{w}^*$), the lower row the Eliassen-Palm-Flux (F_ϕ, F_z), commonly used diagnostics in the assessment of atmospheric phenomena (e.g. Dunkerton *et al.* [1981]; Gray *et al.* [2003]). In the winter, there is a strong tropospheric meridional (Hadley) circulation in the Northern Hemisphere and strong downwelling in the NH polar vortex. Wave momentum is deposited in the Northern Hemisphere subtropical and polar jets. These background states are used as reference in the further analysis.

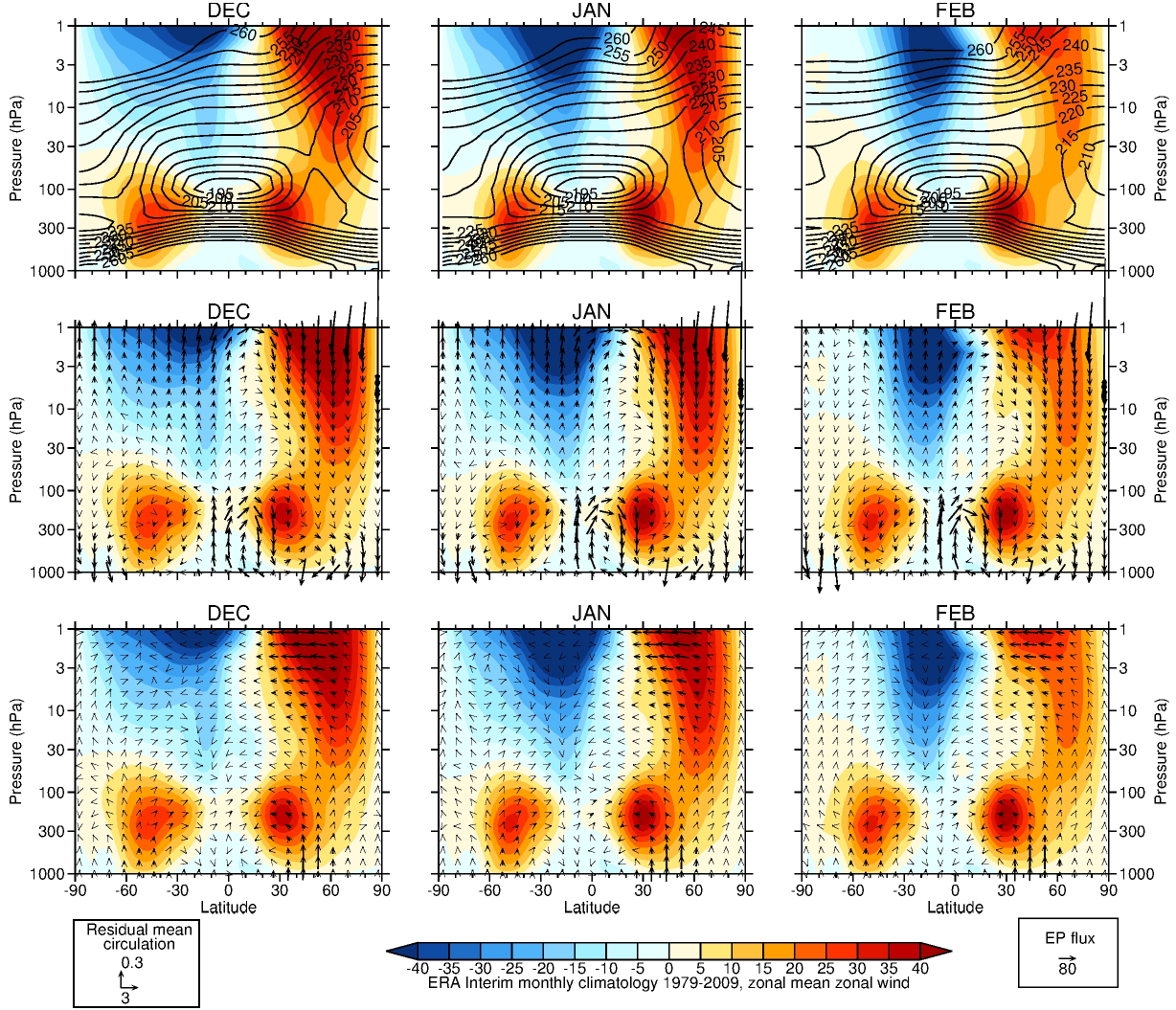


Figure VI.1.: Monthly mean zonal mean zonal wind (colours, all panels) and temperature (upper), residual meridional circulation (middle) and Eliassen-Palm-Flux (lower). The vertical EP-flux was scaled by the square root of pressure for better visibility. Reference vectors for $(\bar{v}^*, \bar{w}^*)$ and (F_ϕ, F_z) (in $\frac{m}{s}$ and $\frac{kg}{s^2}$ respectively) are given in the bottom boxes left and right. ERA-Interim dataset, 1979-2009.

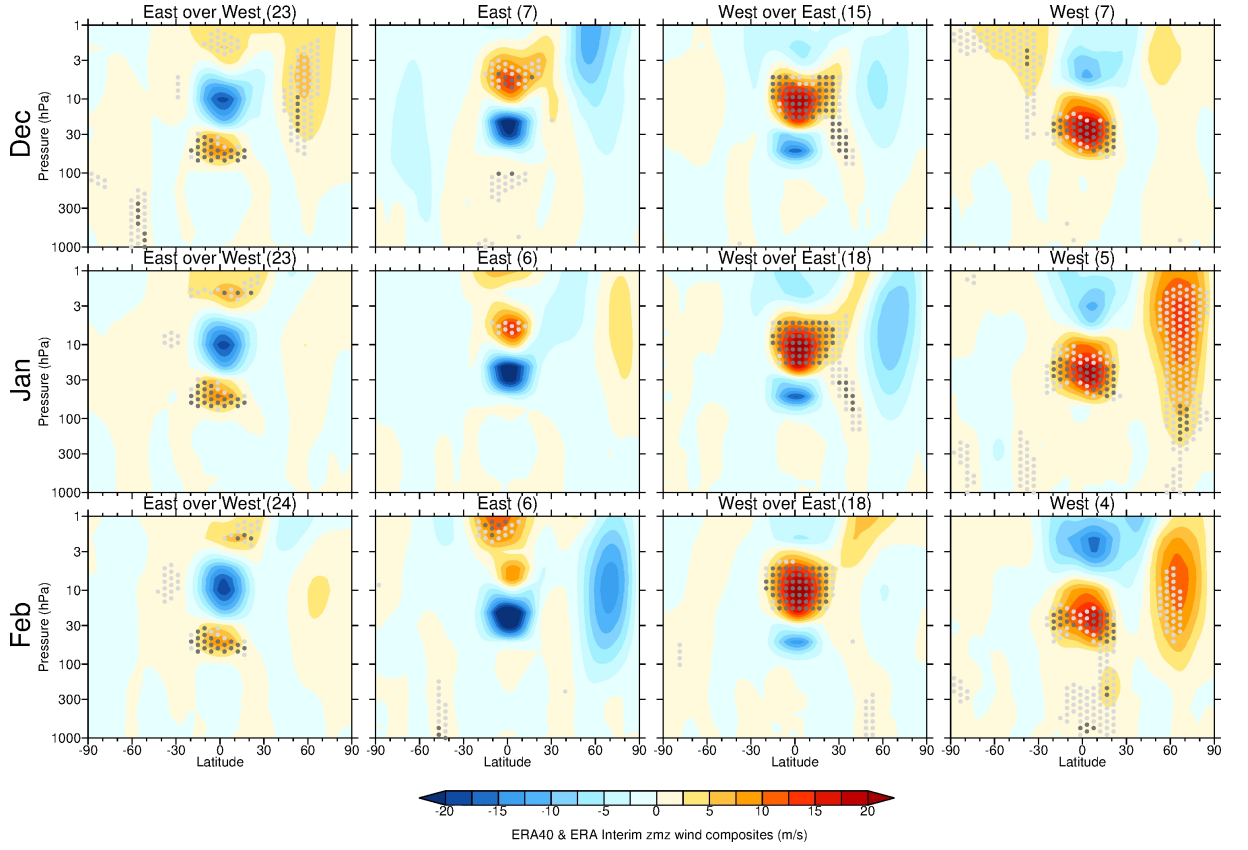


Figure VI.2.: Zonal mean zonal wind winter month composites using ERA40 & ERA Interim data (1958-2010) and the four different QBO states defined as in section V.2. Numbers in brackets show the numbers of years contributing to each composite. Stippling indicates significance (bootstrap test) at the 95% and 99% level (light and dark grey).

VI.2. The Northern Hemisphere teleconnection

A first assessment uses the composite analysis as in the previous section, defining four different QBO states. Data from a combination of ERA40 (1958-1978) and ERA-Interim (1979-2010) is used. The composites of zonal mean zonal wind are shown in Figure VI.2 for the NH winter months. Significance in the equatorial stratosphere is due to the state of the QBO and appears mainly in the westerly shear zone as the mean state is easterly (Figure VI.1). In the polar vortex region (stratosphere north of 60hPa), the composites show hardly any significance, though the sign of the winds in the region are mostly consistent with the Holton-Tan effect.

Most significance in the extratropics can be found in the pure West states in January and February. Despite only a few years contributing to the composite, the connection to the troposphere is not just by chance, as the bootstrap test indicates. The presence of a westerly anomaly during the QBO pure west state in the subtropics might influence wave propagation between the equator and the Northern Hemisphere. The composites from the pure states showing a strong Holton-Tan effect are consistent with Anstey & Shepherd [2008], despite their study applying other selection criteria. As also proposed in this work, Anstey & Shepherd [2008] suggest that there is not one single pressure level that is important for the polar vortex response to the QBO, but rather certain vertical structures.

Figure VI.3 shows the QBO regression coefficients for zonal wind, temperature and residual mean circulation during NH winter months. In the temperature field, significance can be found at the equator, in agreement with the thermal wind relationship. Additionally, the lower stratosphere at high Northern latitudes is significantly colder in January. This analysis supports the theory of the relation of the QBO to the secondary circulation: At the equator, there is a positive/negative temperature anomaly below a descending westerly/easterly caused by the anomalous downwelling/upwelling in the residual circulation. In the winter months this QBO induced secondary circulation is more focussed in the Northern Hemisphere, which is in agreement with Jones *et al.* [1998]. The EP flux vectors show momentum flow from the equator to the higher latitudes in the upper stratosphere as well as a strong upward flux in the mid-latitudes in early winter. The wave forcing in the lower equatorial stratosphere and the anomalous descent both help the westerly QBO phase coming down. The tropospheric meridional circulation is strengthened in this QBO configuration (compare Figure VI.1) and an additional residual circulation cell in the stratosphere is introduced.

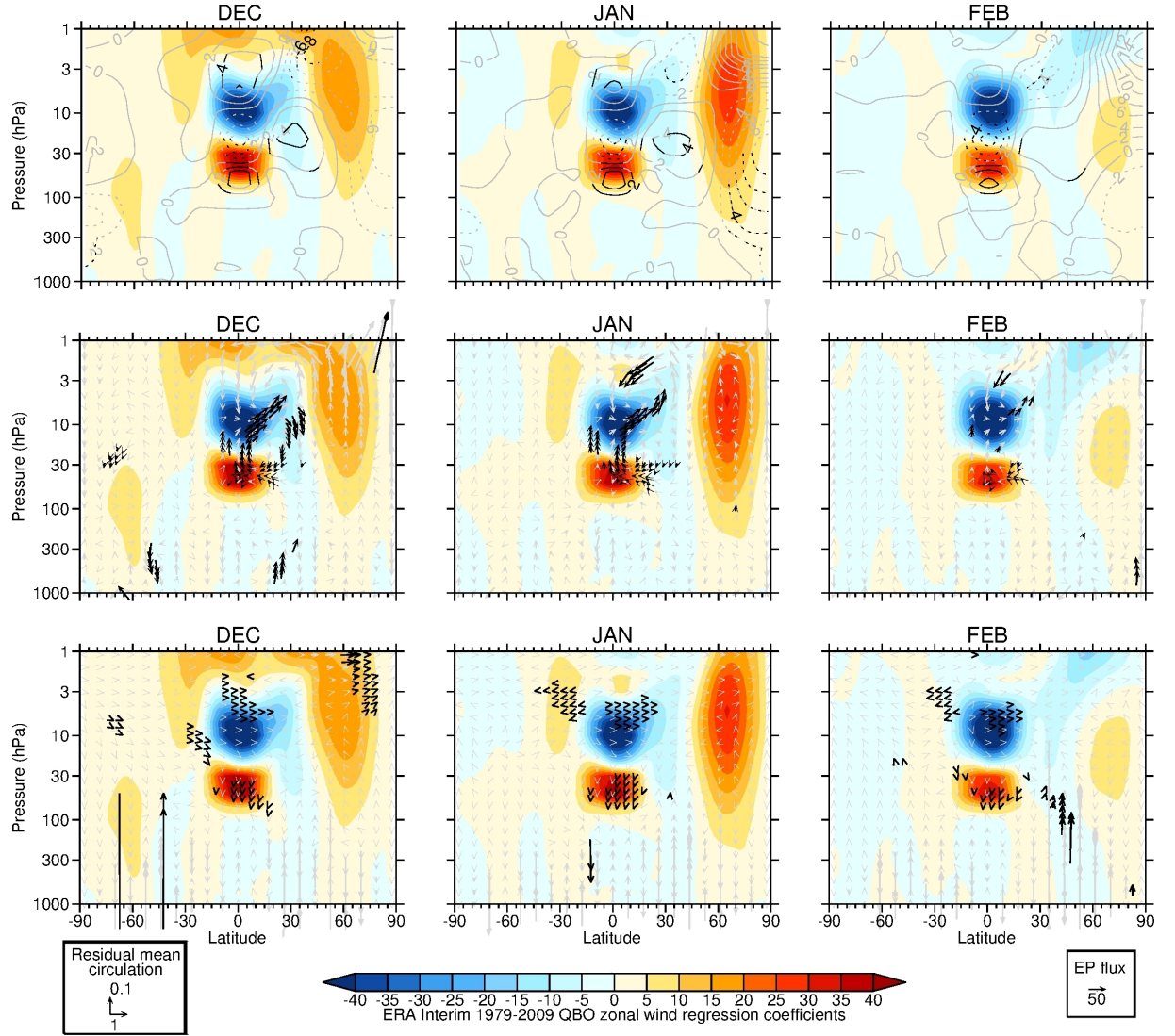


Figure VI.3.: ERA Interim zonal mean QBO scaled regression coefficients in winter months. The QBO index regressed against is the reconstructed u^* at $\varphi = 0^\circ$. The colours show the coefficients of the zonal wind field, the contours in the upper row the coefficients of the temperature field (spacing of 2K), negative values are dotted. Middle row vectors show the residual mean circulation, lower row the EP-flux (scaled by square root of pressure in the vertical). Grey lines/vectors depict the whole field, black ones where the respective field is significant at the 95% level with $\sigma_{(v^*, \bar{w}^*)} = \sqrt{\sigma_{v^*}^2 + \sigma_{\bar{w}^*}^2}$.

Looking at the QBO regression coefficients of these fields in the winter months, it is important to note that the significance in any field is mostly limited to the stratosphere. However, the mean sea level pressure regression analysis (e.g. Figures V.8, V.9) does show a connection between the QBO and the high-latitude surface. While it is only barely evident in the zonal mean zonal wind and temperature field, the QBO-surface influence shows up much more clearly in geopotential height (Figure VI.4), where it maximises in January for phase shifts between $10^\circ \leq \varphi \leq 30^\circ$ (compare Figure V.9). Additionally, QBO regression coefficients of the tropospheric polar vortex (50°N - 70°N , 100-300hPa) are significant for these phase shifts. As an example of QBO-Northern Hemisphere interaction, the regression analysis for the wind field (u, v, w) is shown in Figure VI.5. The modulation of the vortex, that is associated with a change in Annular Mode pattern, can be clearly seen. Most notable are the strong vortices over the North Pacific and Atlantic present at all heights in the stratosphere.

However, the QBO is not the only factor having a strong influence on geopotential height or the residual mean circulation - ENSO regression coefficients β_{ENSO} in the tropical and Northern Hemisphere troposphere and 11-year solar cycle regression coefficients β_{Solar} in the stratosphere are of similar value and significance. Figure VI.6 shows the regression coefficients of the Solar cycle, ENSO and the QBO for January in four different zonally averaged variables: zonal wind, temperature, geopotential height and divergence of Eliassen-Palm flux. The QBO index at $\varphi = 0^\circ$ was chosen as this maximises the QBO response (taken as product of coefficients and significance) of the stratospheric polar vortex (50°N - 70°N , 30-1hPa). The solar and QBO influence on the polar vortex $\bar{u}$ are of the same sign, though only the latter is significant. ENSO, however has a weakening effect, with significance mostly in the troposphere; in zonal wind, temperature and geopotential height the regression coefficients of ENSO are most significant in the troposphere in the tropics.

Significance in geopotential height can also be seen in the extratropical troposphere, where it is again of opposite sign to the Solar and QBO influence. Regarding EP-flux, the ENSO signal in the upper stratosphere resembles the solar signal and opposes the QBO signal, however, significance is rather low in all three coefficients. This might be due to $\vec{F}$ being a rather noisy quantity.

High significance of the QBO regression coefficients is noticeable in all four variables in the equatorial stratosphere, but also in the mid-latitude stratosphere, reaching the surface for temperature and geopotential height at NH high latitudes. This indicates that the QBO has an

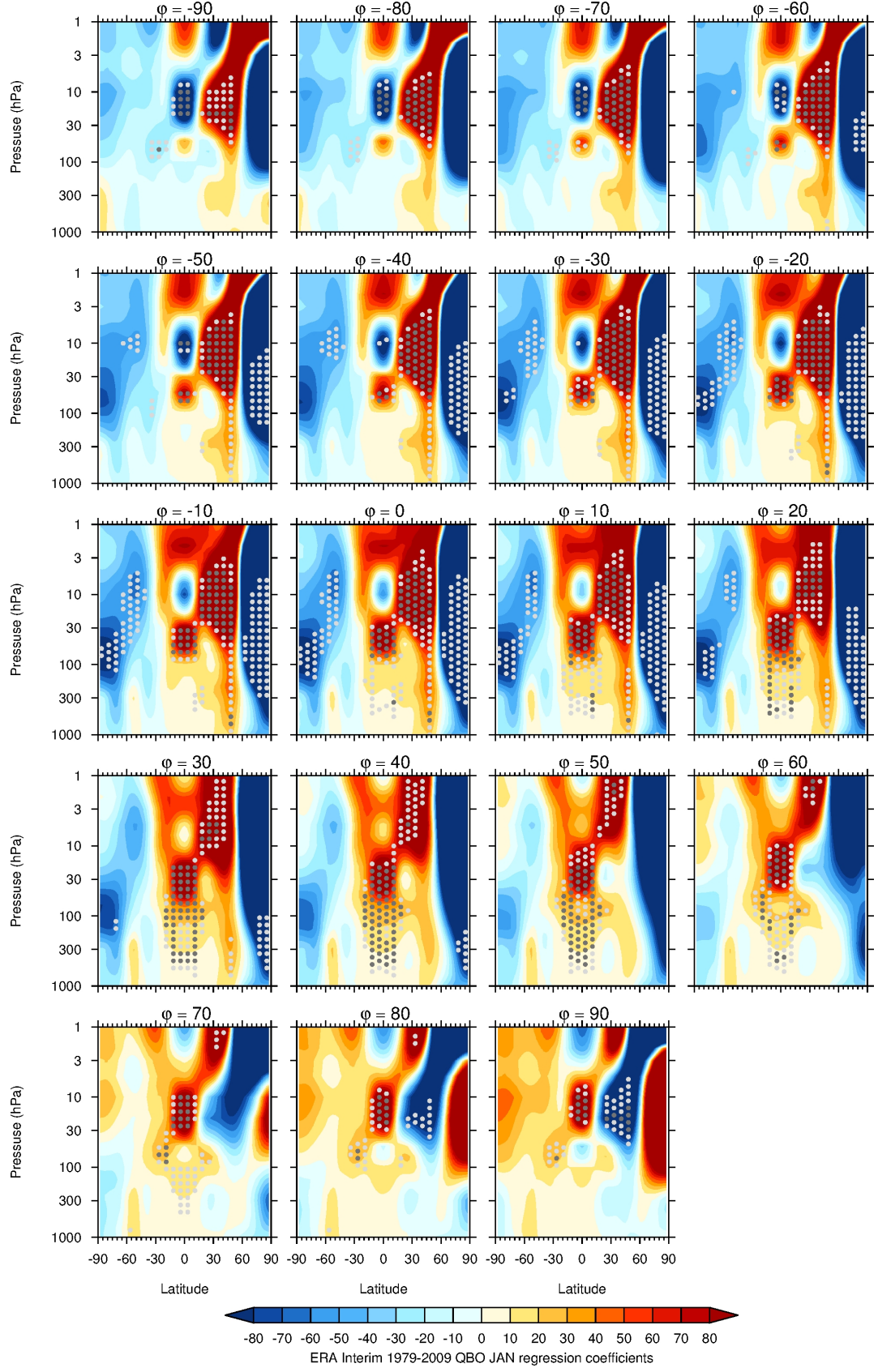


Figure VI.4.: Scaled QBO regression coefficients in January of geopotential height, using $u^*(\varphi)$ as QBO index. Dotting indicates significance at the 95% (light grey) and 99% (dark grey) level.

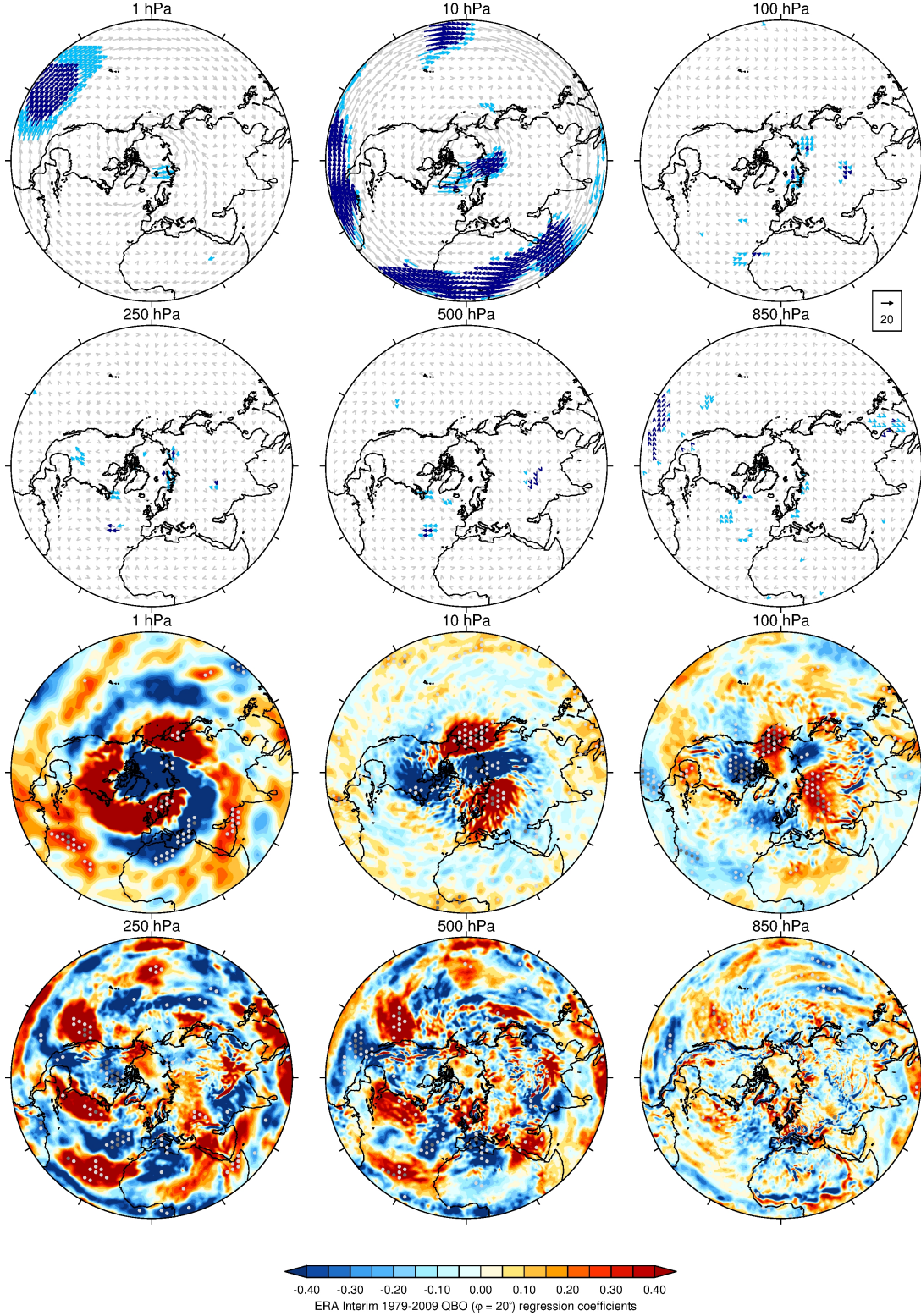


Figure VI.5.: QBO regression coefficients of the circulation anomalies at several pressure levels in the stratosphere and troposphere. Upper two rows: Light grey vectors show the full field of (u, v) regression coefficients, light blue ones where the 90% significance level is and dark blue ones indicate 95% significance, which was calculated as $\sigma_{(u,v)} = \sqrt{\sigma_u^2 + \sigma_v^2}$. A reference vector for the circulation (in m/s) is given. Lower two rows: Colours show QBO regression coefficients of w (in m/s), stippling indicates 95% and 99% significance. The reconstructed u^* at $\phi = 20^\circ$ was used as QBO index.

influence on the NH troposphere (as also shown by the surface analysis, Figure V.9); these effects on geopotential height, zonal wind and temperature are similar to the solar signal, as recently pointed out by Yamashita *et al.* [2015]. The influence on EP-flux, however, is less similar, with significance of the solar signal being around 60°N and 100hPa, whereas the QBO signal is strongest at the equator and mid-latitudes (up to 50°N) above 100hPa. Looking at the significant features, the QBO signal seems to be transmitted via the stratosphere, though tropospheric processes as presented by Yamashita *et al.* [2015] cannot be excluded here.

The effects of these three forcings (Solar, QBO, ENSO) on the residual circulation ($\bar{v}^*, \bar{w}^*$) is shown in figure VI.7; included also is the December stratospheric polar vortex, as Yamashita *et al.* [2015] suggest that the latter might play a role in Sudden Stratospheric Warmings later in the year. Inclusion of this coefficient does not substantially change the patterns of the other three regression coefficients, so it is valid to compare figures VI.7 and VI.6. The solar term actually has very low amplitude and significance in the meridional circulation, which is expected given the same for the temperature field. The solar influence on EP flux is similarly low. At northern polar latitudes solar and QBO signal in wind and temperature again resemble each other, with ENSO coefficients being of opposite sign. Polar vortex strength in December clearly has an influence on $\nabla \cdot \vec{F}$ in January, and a large impact on temperature, which might be a precursor of SSWs.

In the circulation, a clear QBO influence can be seen, which is in accordance with the QBO theory (Figure II.7), and mostly confined to equatorial and mid-latitudes. The coefficients of EP flux also seem consistent with the theory of wave propagation (divergence in the westerly, convergence in the easterly shear zone) and indicate stronger poleward wave fluxes in the upper stratosphere. These are opposed by the equatorward fluxes in the other forcing terms. In contrast to Yamashita *et al.* [2015], there is little significance in the tropospheric wave flux and solar/QBO influences are of opposite sign in the upper stratosphere. One reason might be that in their ERA40 composite analysis, they do not take ENSO or Volcanic influences into account, which are thought to influence the large scale circulation and the polar vortex. Comparing the sign and pattern of the ENSO and QBO coefficients also backs up the modulation of El Niño teleconnections via the stratosphere as pointed out by Richter *et al.* [2015].

Looking at the timing of the connections, one can refer back to the regression analysis using different phase shifts in $u^* = R \sin(\psi + \varphi)$. In geopotential height (Figure VI.4), for example, the QBO regression coefficients in December and January become significant in the equatorial

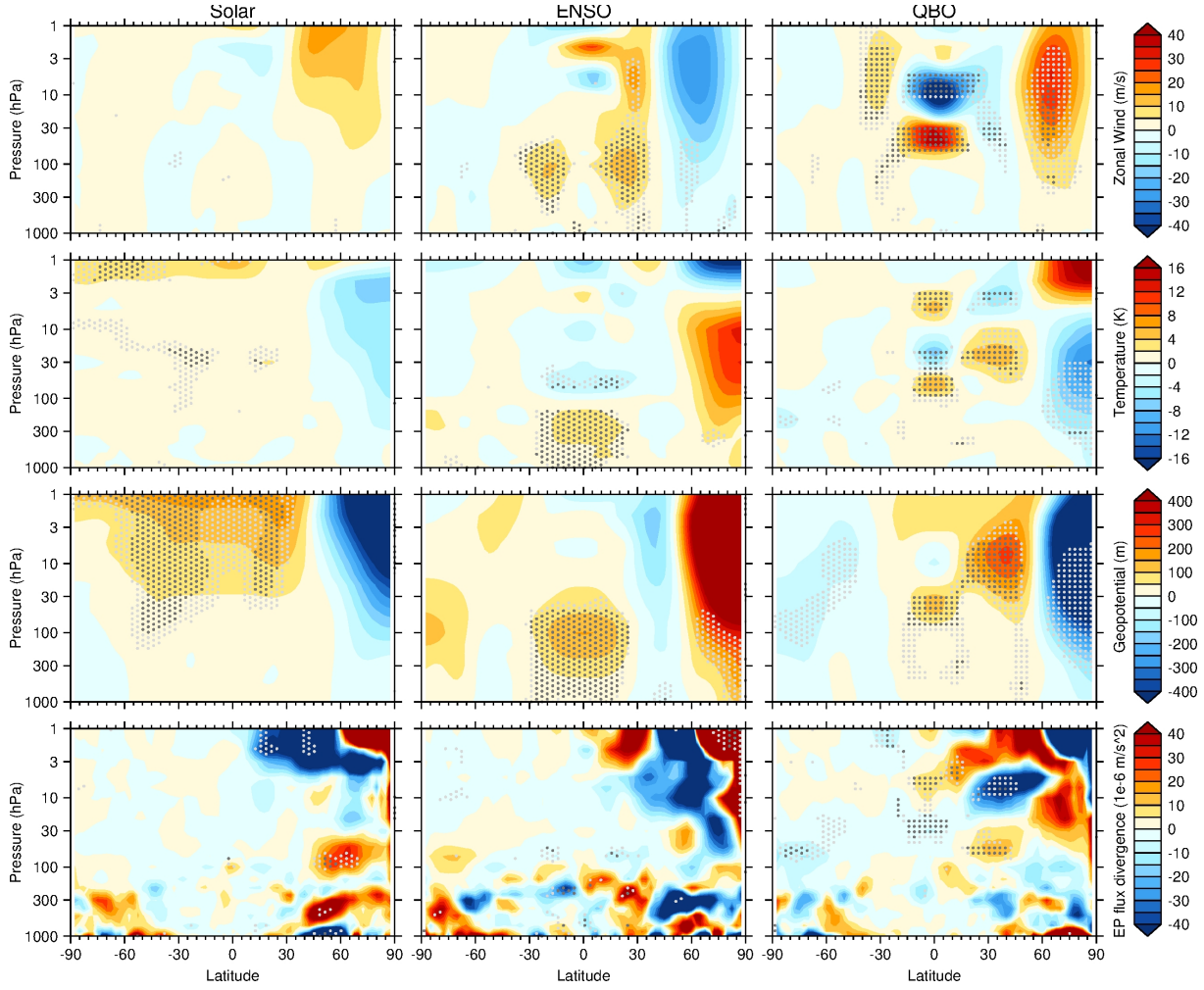


Figure VI.6.: Scaled Solar, ENSO and QBO regression coefficients of 1979-2009 ERA Interim January zonal means of zonal wind (1st row), temperature (2nd row), geopotential height (3rd row) and Eliassen-Palm-Flux divergence (4th row). Note the similarities in the Solar and QBO signal in polar zonal wind, temperature and geopotential height and the opposing coefficients of ENSO. Colours denote the amplitude of the regression coefficients, dotting the 95% (light grey) and 99% (dark grey) significance level.

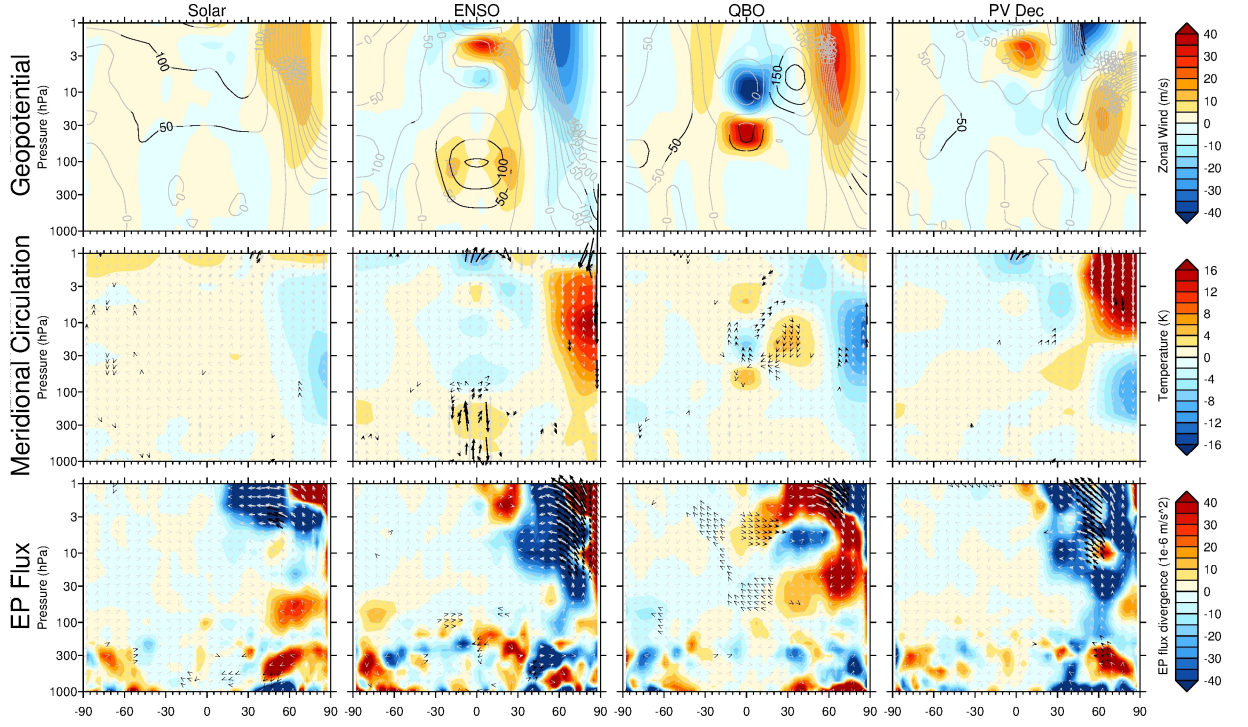


Figure VI.7.: Extension of the analysis shown in VI.6 with a term including the December stratospheric Polar Vortex strength (50°N - 70°N , 30 - 1hPa). Colours show regression coefficients of zonal wind (upper row), temperature (middle row) and EP-flux divergence (lower row), grey lines/vectors the geopotential height, residual mean circulation and EP-flux, respectively. Black lines/vectors indicate where the respective field is significant at the 90% level.

troposphere at phase shifts $50^{\circ} \leq \varphi \leq 70^{\circ}$, coincident with stronger upwelling south of the equator which means a shift of the background circulation to lower latitudes. As the phase shift φ can be interpreted as a shift in time as well, there seems to be a modulation of upwelling first, followed by the change in geopotential height and even later the stratospheric induced meridional circulation and wave propagation into the higher latitudes. As seen from the QBO regression coefficients of the EP flux, the extratropical modulation takes place mainly via the stratosphere.

Taking latitude-longitude cross sections of the circulation field and applying the same regression procedure (using QBO u^* , Solar, ENSO, CO_2 , Volcanic) leads to the QBO coefficients shown in Figure VI.5. Three stratospheric levels (1hPa , 10hPa and 100hPa) are shown as well as three tropospheric ones (250hPa , 500hPa , 850hPa). At 1hPa , significance is limited to a sector in the equatorial Pacific, which gets smoothed out in the zonal mean analysis. At 10hPa , the equatorial winds are dominated by the QBO, which leads to large and significant regression coefficients. Additionally, the influence on the stratospheric polar vortex can be seen, with significance located mostly over Greenland and Siberia. In accordance with Figure VI.6, the

direction of the Polar circulation is opposite to the equatorial circulation. At lower levels in the atmosphere, there is little significance, although an (anti-)cyclonic circulation response to QBO forcing appears in the North Atlantic (which would be expected from a QBO-NAO connection) and the region of the Aleutian Low (mainly at 250hPa).

The pattern in the upwelling w (Figure VI.5) is very interesting as well. While there is little significance in the zonal mean, the latitude-longitude cross sections show distinct patterns. At 1hPa, it resembles a wave with zonal wavenumber 1, at 10hPa and 100hPa a clear wave number 2 pattern emerges at the North Pole. This wave pattern shifts more to the mid-latitudes and higher wave numbers at 250hPa and 500hPa; at 850hPa the QBO signal in upwelling is weak and hardly significant apart from a region near the equator in the Pacific, which can also be seen in the (u, v) circulation response and hints towards a QBO-ENSO link.

VI.3. Equatorial QBO interactions

It has been further investigated whether the QBO modulation of equatorial upwelling is confined to a certain region. Such a relation may not be seen in a zonal mean plot. Figure VI.8 shows the QBO regression coefficients of w at 850 hPa in January for different phase shifts. Anomalous upwelling at high phase shifts ($80^\circ < \varphi < 90^\circ$) can be seen at mainly in the Pacific and Indian Ocean up to a latitude of about 20°S . Corresponding anomalous downwelling happens in a thinner band in the Northern Hemisphere around 5° . For $50^\circ \leq \varphi \leq 70^\circ$, which is where the zonal mean regression coefficients show significance in the troposphere, there is a band of stronger upwelling just off the equator in the Pacific and the Indian Ocean. These two regions are especially interesting in the light of the connection between the QBO and the Madden-Julian-Oscillation [Madden & Julian, 1971, 1972], that has attracted the attention of recent studies [Liu *et al.*, 2014; Yoo & Son, 2016].

The connection between the MJO and the QBO emerges even clearer when looking at cloud cover at 250hPa (Figure VI.9). As the MJO is characterised by zones of deep convection, it will manifest as cloud cover, which is why this makes a good parameter to assess its strength. The QBO seems to influence cloud cover in the sea west of Australia and to the North, indicated by high and significant regression coefficients for phase shifts $-30^\circ \leq \varphi \leq 30^\circ$; this corresponds to a westerly in the lower stratosphere (Table V.1). For a typical MJO pattern, however, this is too far South and does not extend far enough into the Indian Ocean (compare Figure II.12). At

high phase shifts ($\varphi \sim \pm 90^\circ$), however, the fields look similar to phases 4 and 8 of the winter MJO.

With the effects of the MJO on convection one would expect a stronger circulation towards the MJO centre at 850 hPa and flow away at 250hPa. The regression coefficients of (u, v) however, do not exactly show this pattern or much significance in the MJO region in January (Figure VI.10). There still are interesting features such as the significant wind anomalies over the Southern Indian Ocean that appear at tropospheric heights, as well as the strong and significant anomalies over Australia at 250hPa and 100hPa. The latter could correspond to MJO-related circulation changes (compare with Figure VI.9 at $\varphi = -40^\circ$). In the tropical stratosphere at 10hPa, the zonal wind pattern of the QBO can be seen, with subtropical winds being opposite to the equatorial wind. Based on this analysis, it is unclear what the effects of the QBO on the MJO are; however, Yoo & Son [2016] find a connection using correlation and composite analysis. Their ideas will be further addressed in Section VII.4.

VI.4. Chapter summary

Regression analysis, similar to the one in the previous chapter, was applied to zonal mean fields to address possible mechanisms by which the QBO influences the NH winter mean sea level pressure. Examined variables included zonal wind, temperature, EP flux, residual meridional circulation (Figure VI.3) and geopotential height (Figure VI.4). The QBO influences were compared to the 11-year Solar cycle and ENSO (Figure VI.6).

The QBO is found to have significant impact on NH polar tropospheric temperature and geopotential height, whereas the significance in the meridional circulation and EP fluxes is limited to the mid-latitudes. ENSO has the opposite effect on polar winter temperatures and geopotential height than the QBO, whereas the regression coefficients of the solar term are not found to be significant in the extratropics.

Additionally, the regression analysis is applied to latitude-longitude wind fields (Figure VI.5); the QBO influence on the Polar Vortex (zonal and meridional wind at 10hPa) appeared to be significant over the Atlantic and Siberia. In upwelling, a wave-1 pattern emerges at 1hPa over the North Pole, which turns into a wave-2 pattern at 10hPa. This hints towards the QBO influencing planetary wave propagation into the extratropics.

Similar regression analysis of cloud cover (Figure VI.9) and upwelling (Figure VI.8) was applied

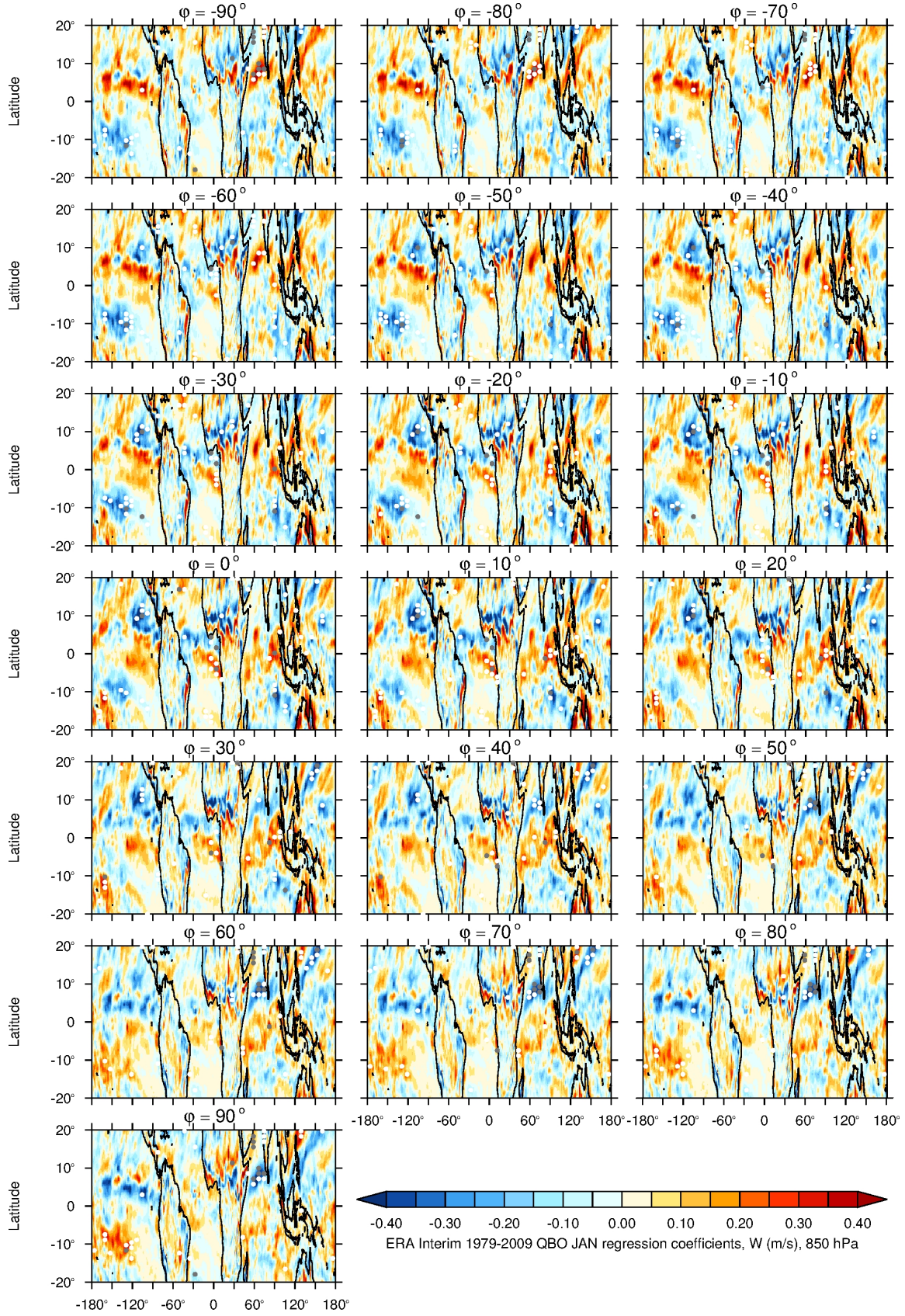


Figure VI.8.: QBO regression coefficients of upwelling w at 850hPa in January in the equatorial region. Stippling indicates significance at the 95% (white) and 99% (grey) level.

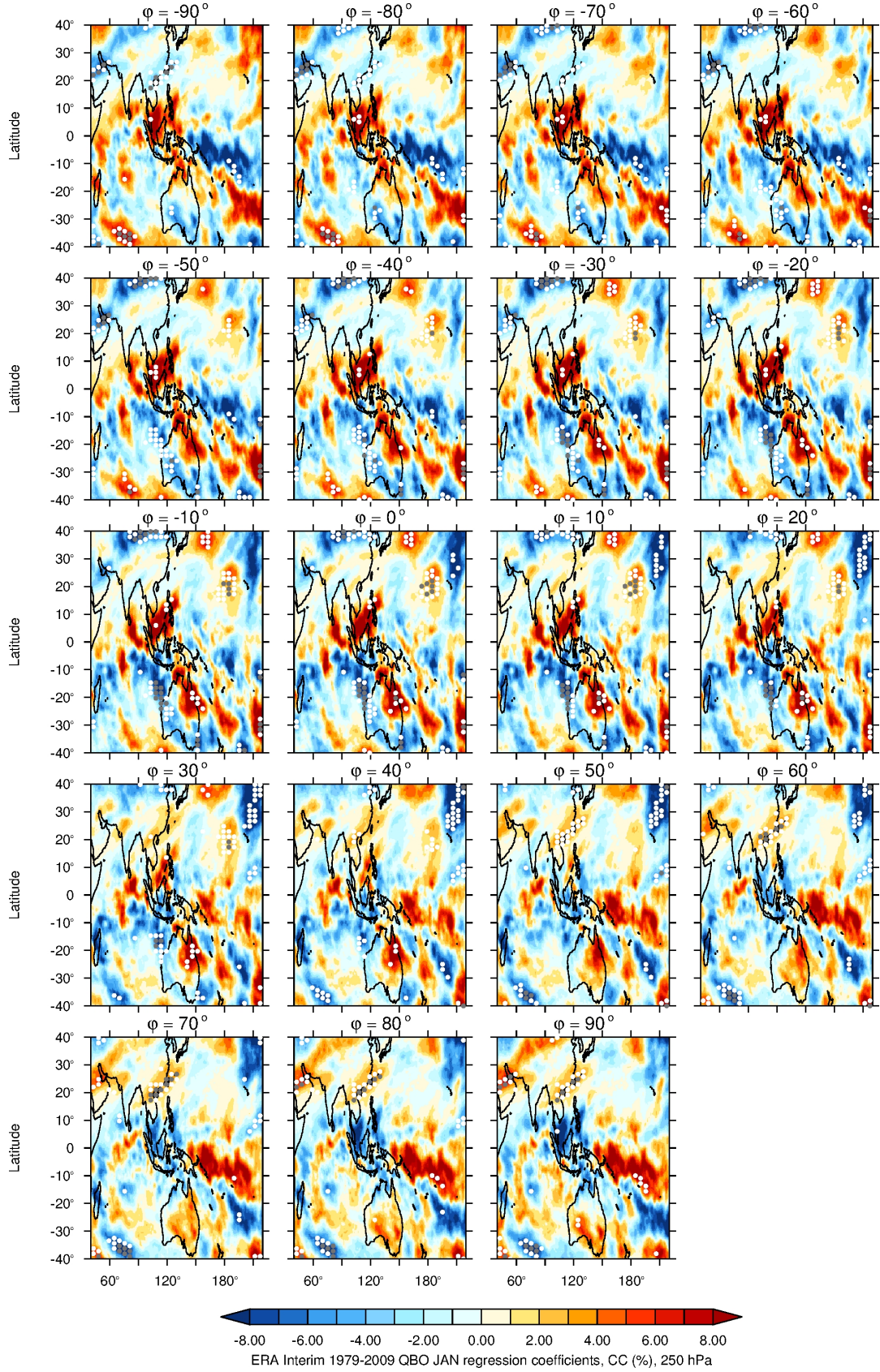


Figure VI.9.: QBO regression coefficients of cloud cover at 250hPa. This is a measure for deep convection and thus for the activity of the Madden-Julian Oscillation. The chosen sector is identical to the one in II.12, so that the patterns can be compared. Higher cloud cover (red) means less outgoing long wave radiation, whereas lower cloud cover (blue) indicates more OLR.

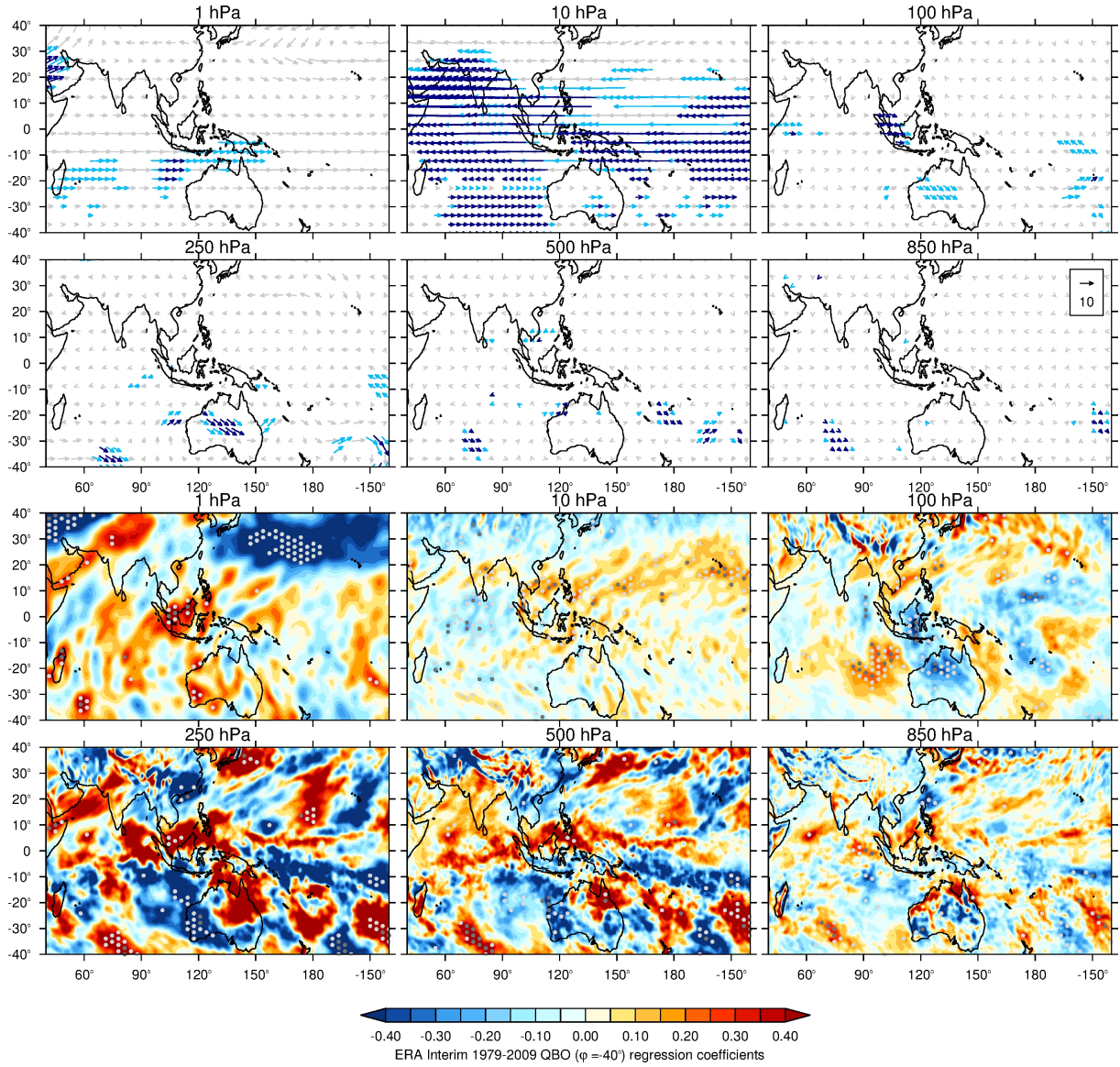


Figure VI.10.: Same as VI.5, but for the MJO region and $\varphi = -40^\circ$.

in the equatorial region to address a recently claimed link between the QBO and the MJO. The results of this analysis (Figure VI.10) are inconclusive; links between the QBO and the MJO can neither be confirmed nor disregarded. In contrast to previous studies, who used composite and correlation analysis, the regression analysis applied here accounts for other influences, such as ENSO, which might be one reason for the discrepancy with Yoo & Son [2016].

VII. Causal effect network analysis

While the methods so far give hints on the possible connection between the QBO and the extratropics and backs the theoretical understanding of the QBO, the main question of the interaction mechanism could not be solved. Therefore, a new method is applied that can distinguish between a correlation of variables and a causal relationship; the latter can be termed “forcing” when taking time lags into account: Causal Effect Networks, short CEN [Runge *et al.*, 2012a,b, 2014, 2015]. It is possible to distinguish whether a correlation between two variables X and Y is due to a direct link between the two, mediated by a third variable Z , has a common driver Z or is arising from a strong autocorrelation of one or both of the variables.

The method also allows to assess links $X \xrightarrow{Z} Y$ not taking selected datapoints in one of the variables into account, e.g. for the difference between QBO East and West phases in forcing, months with wind over/under a certain threshold could be masked out when the QBO is the driver of the link in question (X). If the effect on the winter NAO is estimated, only the relevant months will be included when the NAO is in the position of the driven variable (Y). Similarly, years with major volcanic eruptions can be excluded, or a distinction between ENSO years can be made. So far, the code only allows masking variables in one of the positions in the equation, as driver X , driven variable Y or condition Z .

Kretschmer *et al.* [2016] have already implemented this method to test an observational dataset for different arctic drivers (sea ice and snow cover in different regions) of the Northern Hemisphere winter (DJF) circulation, characterised by the Arctic Oscillation. They limit their analysis to several mid to high latitude surface influence factors that vary on shorter timescales than possible tropical drivers. However, as the previous analysis shows, the connection between the equatorial stratosphere and the extratropical surface cannot be neglected. Thus, the CEN method is applied to the timeseries used in the regression analysis to shed some light on possible links from the QBO to the extratropical troposphere.

Figure VII.1 depicts the response of the NAO, stratospheric polar vortex (PV) and MJO to QBO, ENSO and Solar forcing derived from regression analysis (Section III.2.2). Timeseries

from observations (Table III.1) are deseasonalized and detrended, the Volcanic term is regressed on as well, but is not drawn. Shown are the scaled regression coefficients in single winter months at different QBO phaseshifts and their significance. ENSO and Solar coefficients are fairly constant, whereas the QBO term leads to coefficients varying in strength and significance, which is expected. The size of the circles indicates significance (see first panel). December and January (first and second column) show similar patterns in all three timeseries, whereas the February pattern (third column) differs for NAO and polar vortex, suggesting that the QBO-extratropical interaction happens mostly in early winter. Using these, the QBO timeseries to put into the CEN analysis framework, is chosen based on the strongest and most significant coefficient in January:

- $\varphi = 20^\circ$ for QBO-NAO influence
- $\varphi = -30^\circ$ for QBO-Polar vortex connection
- $\varphi = -40^\circ$ for QBO-MJO interaction

At these phaseshifts, the QBO regression coefficients have the biggest amplitude of the three forcing terms and are highly significant.

Based on this regression analysis, one might expect a connection between the QBO and the MJO in all the winter months, as claimed by Yoo & Son [2016]; the interaction between QBO and polar vortex/NAO in Figure VII.1, however, seems to be limited to early winter, which is consistent with single month regression analysis of zonal mean wind and surface pressure. The regression using the all month timeseries (Figure VII.1, fourth column) shows no QBO-NAO connection, a weak positive QBO-PV connection ($-30 \leq \varphi \leq 30$) and a year-round QBO-MJO connection whose pattern resembles that of the winter months, but is smaller in amplitude. From this preliminary analysis, one would expect to see a link between the QBO, ENSO and the respective other timeseries (NAO, PV, MJO), for the extratropical connection especially in the early winter months. The solar impact on the polar vortex appears to maximise in January and a month later on the surface, as reflected in the NAO coefficients. This agrees well with Gray *et al.* [2013, 2016]. However, the focus of this thesis is on the QBO; therefore, the analysis is limited to early winter months (DJ), which is likely to obscure solar influences.

The CEN analysis is divided into several parts, establishing connections and influence pathways separately for the stratosphere and troposphere, testing at different significance levels (90%, 95% and 99%). With exception of the analysis that include a Solar term, the maximum lag is

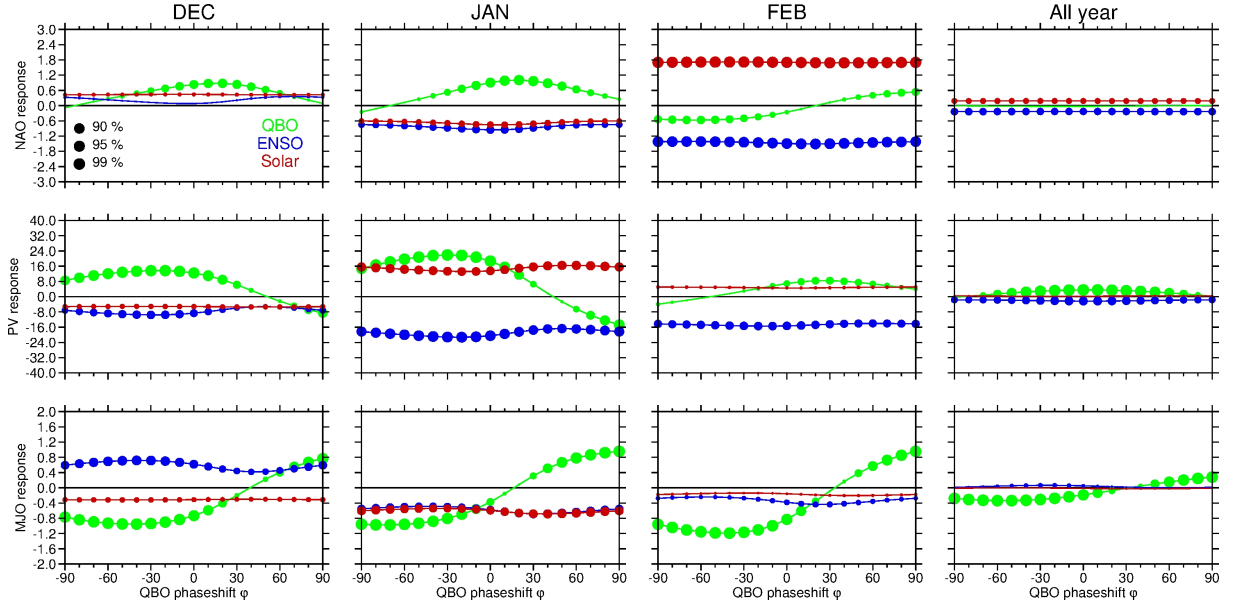


Figure VII.1.: Scaled QBO (green), Solar (red) and ENSO (blue) regression coefficients for NAO (upper row), polar vortex (middle row) and MJO (lower row), depending on the phaseshift φ ($^{\circ}$) of the reconstructed QBO timeseries. Shown are winter months as well as all year analysis. The QBO timeseries with the largest and most significant regression coefficient in January is chosen as QBO index for the CEN analysis.

set to $\tau_{max} = 12$ months, which is about half a QBO cycle ($\tau_{max} = 60$ months for inclusion of Solar timeseries). Apart from the jet indices (upper and lower polar vortex, subtropical jet) that are derived from ERA Interim, the data comes from direct observations, timespan from 1979-2010 (Table III.1). The timeseries of all factors that are included in the various network analyses are shown in Figure III.1.

VII.1. Bivariate analysis

In initial setups, only two indices are included in the analysis (“bivariate”). Two sorts of tests for links are done, one including all month data, the other one only December and January (“winter”) in the forced variable. A sensitivity test regarding the solar influence on the winter NAO (DJ) shows a weak link at lags between 3 and 4 years, which is significant at 90%, but not at 95%. For simplicity and due to higher lag times in the Solar-NAO interaction, the solar term is not taken into account in further establishing the QBO-NAO network (which is in accordance with the regression analysis shown in Figure VII.1). Further, links between the QBO and the NAO cannot happen via a pathway through the sun and the sun as a common driver is excluded as there is no link found between the solar and the equatorial QBO term in the bivariate analysis. The connection between the QBO and various other patterns is tested in order to confirm results

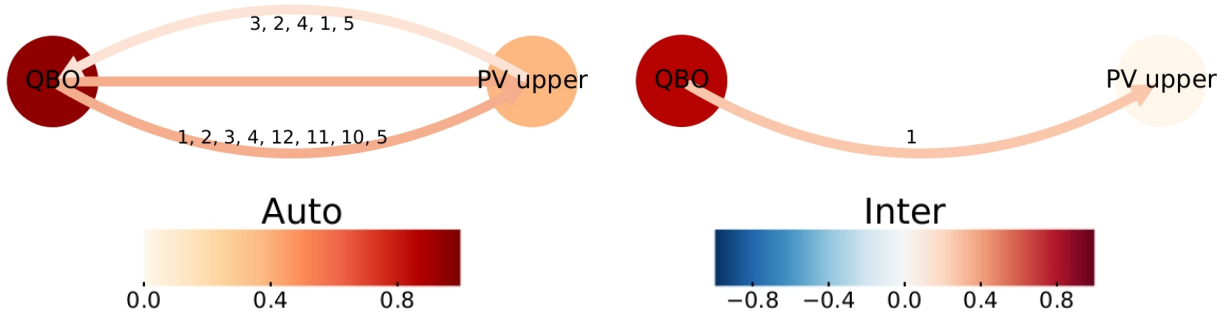


Figure VII.2.: December-January process graphs before (left) and after (right) conditioning on parents, at 95% significance. Numbers on links denote the corresponding lag time (in months), with the colour showing the strongest link as in the lower scale. Straight links without an arrowhead correspond to a lag of $\tau = 0$ and do not have a direction. The circles are depicting the autocorrelation of the processes as shown in the upper scale. Timeseries of PV upper as defined in Table III.1, QBO index is $u^*(\varphi = -30^\circ)$.

of regression analyses (Figure VII.1) with this new framework. Figure VII.2 shows the QBO-PV links as an example before and after the conditioning process (section III.2.3); monthly data are used, with masking the PV data so that only influences on December and January are taken into account. This shows the classic Holton-Tan effect of a positive relationship between QBO and vortex winds. The several significant lags in Figure VII.2 (left) are consistent with the regression coefficients being significant at various phaseshifts. Additionally, there is a link detected from the polar vortex to the QBO, which cannot be distinguished in correlation or regression analysis. These “unconditional” links are tested for independence of each other, starting with the strongest one. The link(s) remaining after these tests is shown as a graph in Figure VII.2 (right): Only the QBO→PV connection at a lag of 1 month is still significant at the 95% level.

Figure VII.3 shows the conditional process graphs at the 95% significance level for QBO-NAO, QBO-PV and QBO-MJO timeseries. As expected from the preliminary regression analysis, there is no connection between the QBO and the NAO in the all year analysis, but weak links to the polar vortex and the MJO. None of the processes show a reverse influence back onto the QBO. Regarding the MJO, this is a bit surprising, given the findings of Moss *et al.* [2016], who explain that the MJO influences stratospheric gravity waves. However, there is an MJO→QBO influence with 6-7 months lag that is significant on the 90% level (not shown).

For the polar vortex, the high lags of the QBO→PV link in the annual analysis might be due to different phase shifts showing a stronger signal in the all month analysis than in the one looking at winter only (compare Figure VII.1). The connection between the QBO and the MJO that is significant at low lag times is also consistent with the regression analysis, which showed the biggest signal and significance at $\varphi = -40^\circ$. The negative sign (blue arrow) of the link is in

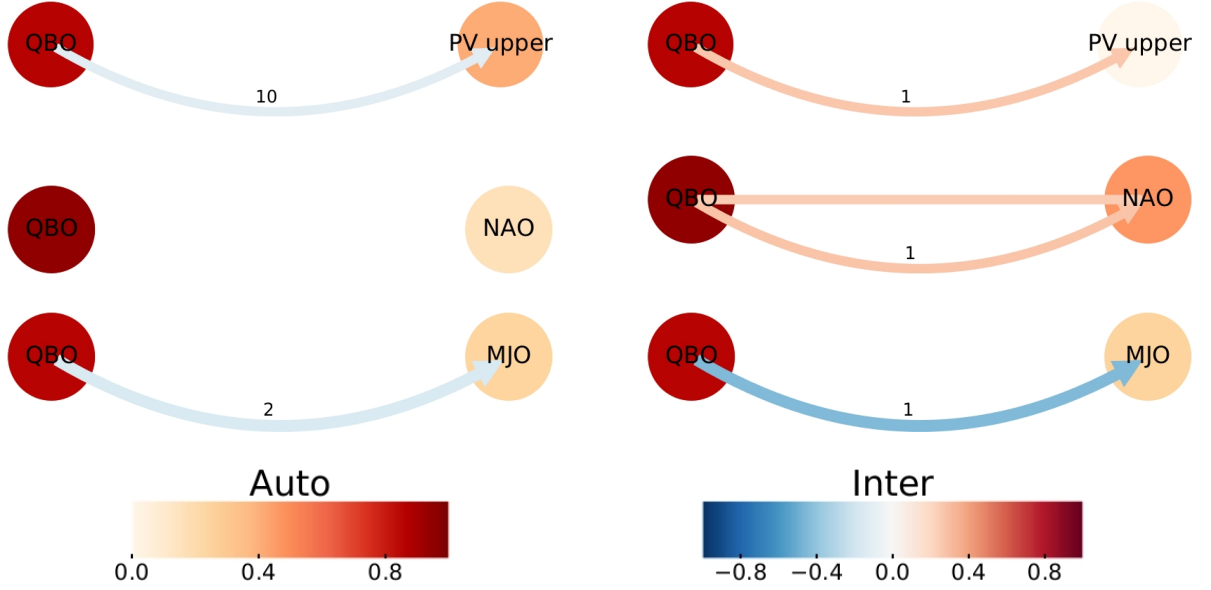


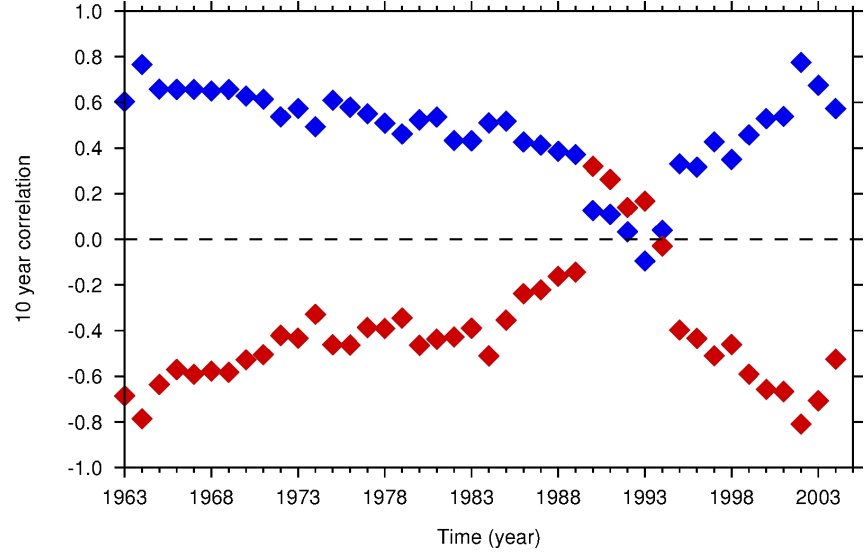
Figure VII.3.: Process graphs using all year (left column) and December-January only (right column), at 95% significance. Numbers on links denote the corresponding lag time (in months). The circles are depicting the autocorrelation of the processes as shown in the left scale, darker red indicates stronger autocorrelation. Colour of the links refers to the link strength (blue: negative, red: positive; similar to a correlation coefficient) with darker colours depicting stronger connections.

agreement with the regression analysis and the findings of Yoo & Son [2016].

The links that are identified in the all year analysis might give hints to the processes involved in the interaction mechanism: an influence through synoptic scale wave forcing would be present all year, whereas the planetary wave forcing is dominant in boreal winter. Thus, the absence of a QBO→NAO link suggests that synoptic scale waves are not the dominant interaction mechanism. The presence of a QBO-MJO connection all year is physically sensible since both are equatorial phenomena and thus should not exhibit a strong seasonal variation. Long lags in the QBO-Vortex all month interaction could also be due to the involvement of a third process that acts on these timescales.

In the winter (December-January) analysis, there are influences from the QBO on the polar vortex and the NAO at low lag $\tau = 1$, as well as a direct QBO-NAO connection. The QBO→MJO link is stronger than in the all year analysis and significant at a lag of 1 month. The QBO→NAO influence seen in the regression analysis is in good agreement with these results, as is the QBO→PV and QBO→MJO interaction.

Figure VII.4: 10 year running correlation between DJF equatorial zonal wind at 50hPa as QBO index and 60°N, 10hPa zonal wind (blue) and temperature (red) as polar vortex index. The year indicates the midpoint of the 10 year correlation. Polar vortex data from a combination of ERA40 and ERA Interim, QBO from FU Berlin dataset.



VII.2. Further exploring the Holton-Tan effect

Since the inaugural paper [Holton & Tan, 1980], that looked at composites of geopotential height depending on the sign of the tropical zonal wind at 50hPa (i.e. QBO phase), the term “Holton-Tan effect” evolved to refer to an influence of the QBO on the strength of the NH winter polar vortex. The correlation between tropical and polar stratospheric wind and temperature in the winter months, however, has varied in strength as noted by Lu *et al.* [2008]. Figure VII.4 shows the 10 year running correlation between equatorial wind at 50hPa and polar vortex wind/temperature at 60°N, 10hPa. The timeseries from ERA40/Interim 1958-2009 is deseasonalized and the winter months (December, January, February) are averaged together before calculating the correlation. For most of the record, the correlation between QBO and polar vortex wind is higher than 0.5, but this relationship breaks for a couple of years around 1990 [Lu *et al.*, 2008], which gives a correlation coefficient of 0.49 (-0.44 for QBO wind-PV temperature) for the whole timeseries. Nonetheless, the connection between the QBO and the strength of the stratospheric NH polar vortex seems robust. The spectrum of the polar vortex wind timeseries shows a peak at QBO frequencies (Figure VII.5), which has the highest amplitude of the polar vortex spectrum, followed by sub-annual variability. However, covariation does not give any information about cause and effect. Applying the CEN method, on the other hand, indeed shows a causal link between the QBO and the upper stratospheric polar vortex in boreal winter (Figure VII.2).

To account for possible Holton-Tan effect modulation by solar variability and the Pacific Decadal Oscillation (PDO), which both operate on the relevant timescale, these terms are added

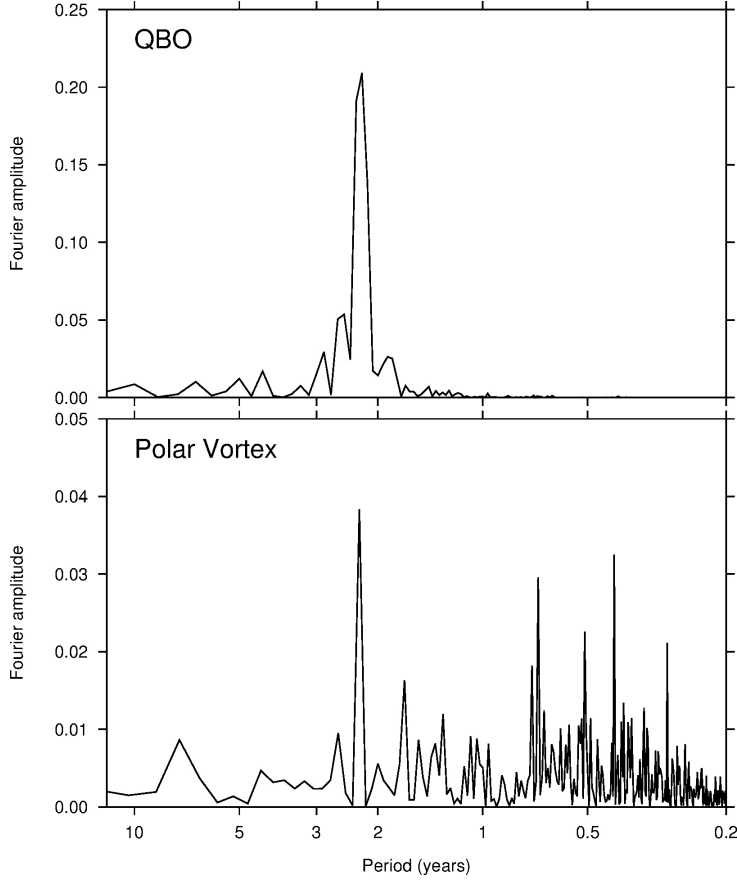


Figure VII.5: Fourier spectra of the first EOF of equatorial zonal wind between 10hPa and 70hPa (QBO, FU Berlin dataset) and at 60°N, 10hPa (Polar vortex, ERA40 and ERA Interim reanalyses). Both timeseries cover the years 1958-2010 and are deseasonalized. The amplitudes are scaled by the total field variance.

separately to the analysis. When using a solar index, the term is masked in the driven variable as the two atmospheric terms will not have an influence on the solar activity. In the Solar-QBO-PV network, the QBO→PV link (lag 1 month) is reduced; it is significant at the 90% level (Figure VII.6), but not at 95%. Figure VII.7 shows the process graphs for the QBO-PV interaction including the PDO timeseries at 95% significance. The QBO→PV link is not significant at the 95% level (it is at 90%); instead, it is found that the PDO drives both QBO and wintertime polar vortex.

As the PDO and ENSO are both ocean phenomena, it is necessary to distinguish between their interaction with each other and influence on the NH winter polar vortex. Figure VII.8 shows the network graphs using QBO, PV and ENSO or ENSO and PDO. Including ENSO only removes the QBO→PV link, whereas additionally using a PDO term reintroduces a direct connection between the QBO and the polar vortex. In the QBO-ENSO-PV network, ENSO is influencing the QBO at lag times between 3 and 6 months, as well as being influenced by the polar vortex. Adding the PDO to this network leads to ENSO indirectly influencing the polar vortex via modulation of the PDO and possibly via the direct QBO link. The PDO→PV and PDO→QBO modulation is comparable to the one established in the network without an ENSO

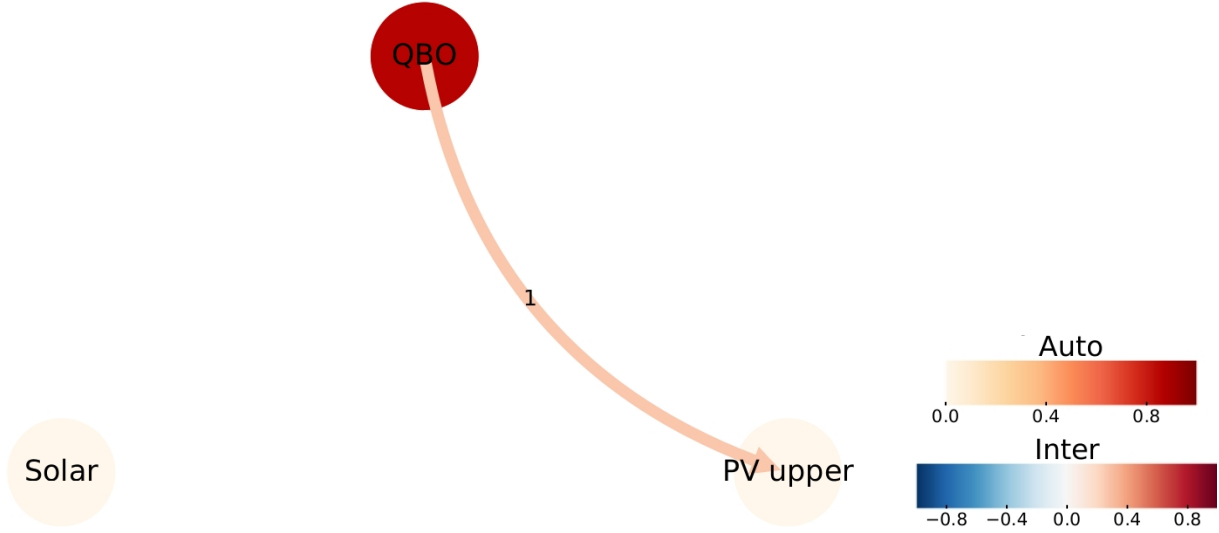


Figure VII.6.: Network graph (90% significance) with QBO, Solar and polar vortex terms. The solar timeseries is masked from the driven variable Y and only influence on PV in winter months is taken into account.

term.

These results, especially the role ENSO and the PDO might play in the Holton-Tan-effect, are very interesting. However, the exact nature of the interactions is difficult to interpret and requires further analysis.

VII.3. Establishing QBO-NAO interaction routes

Having confirmed links between the QBO and the NAO, the question remains how this interaction takes place. Limitations of the link to boreal winter can give hints, as well as the indicated lags, which are tested with monthly, fortnightly and weekly time resolution (Section VII.5). Additionally, the possible routes for the QBO→NAO links are investigated: the “stratospheric” route (via modulation of planetary scale wave propagation and polar vortex) and the “tropospheric” route (influence on meridional circulation and subtropical jet position).

To distinguish these processes, winter month analysis including QBO and NAO as well as one indicator for the possible interaction route (Polar vortex/Subtropical jet) is carried out. The QBO index at $\varphi = 20^\circ$ is used as it maximises the NAO regression coefficient (Figure VII.1).

Figure VII.9 shows the network graphs including a QBO and winter (DJ) NAO index, as well as either the upper/lower stratospheric polar vortex or the subtropical jet index. When using the polar vortex in the upper stratosphere, the QBO-NAO link from the bivariate analysis is not significant at the 95% level, whereas including the subtropical jet does show a contemporaneous

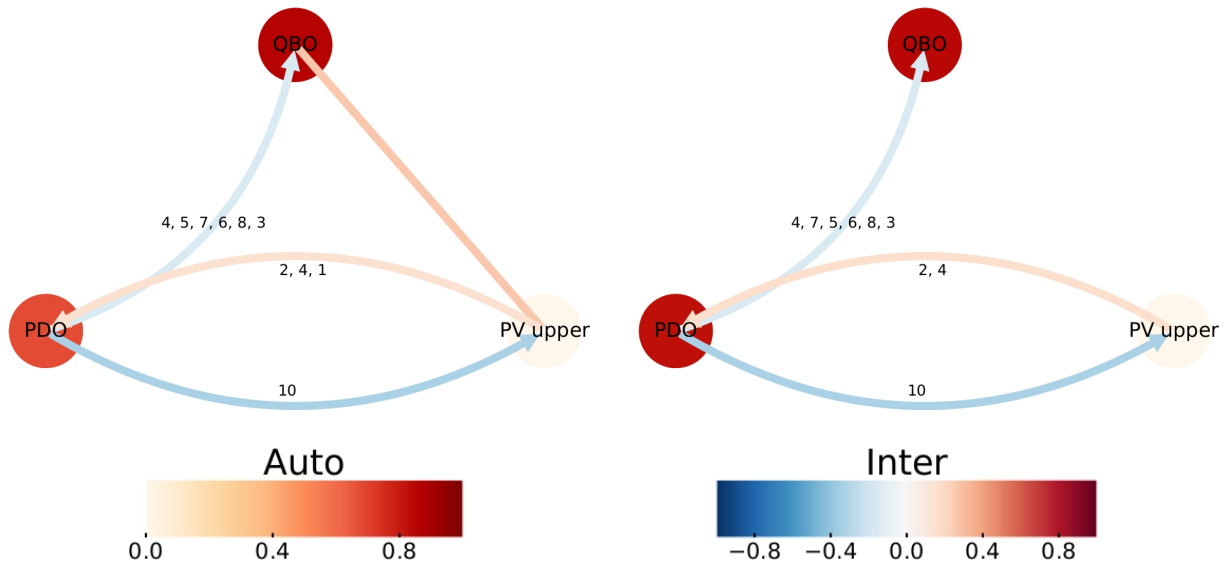


Figure VII.7.: Network graph (90% (left) and 95% significance (right)) using QBO, PDO and polar vortex terms. Only influences on PV winter months are taken into account.

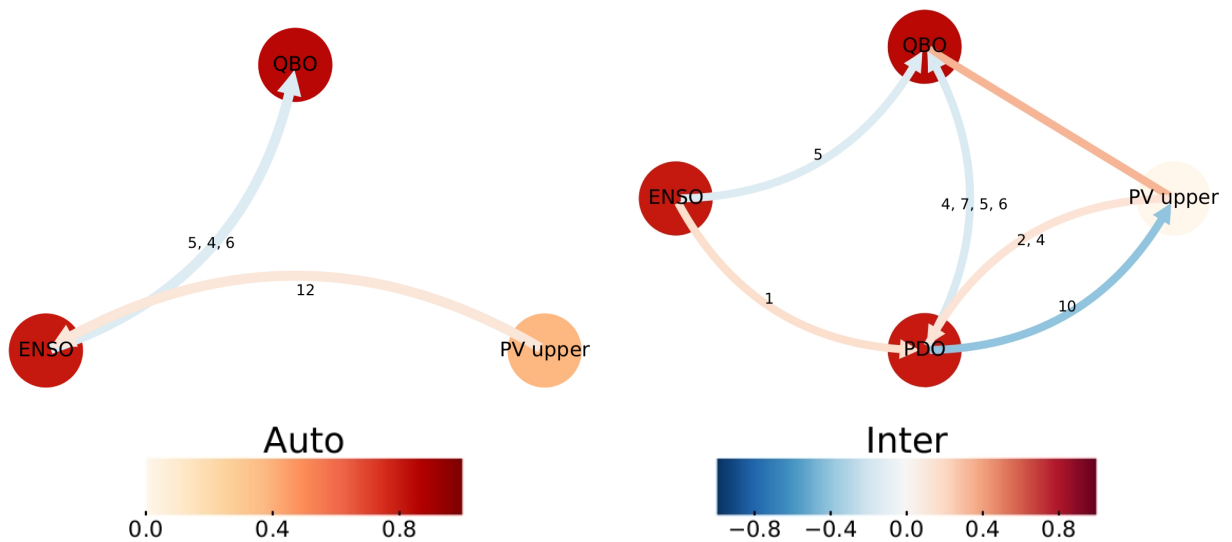


Figure VII.8.: As Figure VII.7, 95% significance, using either an ENSO term instead of the PDO, or both terms.

QBO-NAO link. In both cases, there is an interaction between the QBO and the respective jet, which indicates an indirect influence of the QBO on the NAO via both suggested pathways. Including the lower stratosphere polar vortex, however, breaks the QBO-NAO connection (top right, Figure VII.9), which is rather surprising. One might expect to see a link $\text{QBO} \rightarrow \text{PV lower} \rightarrow \text{NAO}$ exists. However, in this setup the potential link $\text{QBO} \rightarrow \text{PV lower}$ is not found to be significant at the 95% level. The polar vortex route is re-established when including the upper as well as the lower stratospheric vortex, as shown in Figure VII.10: The QBO is directly linked to the upper part of the polar vortex, which is closely connected to the lower part, both influencing each other, with the influence of the upper on the lower stratosphere being stronger than the other way round. The lower vortex then modulates the NAO. This top-down mechanism is consistent with Baldwin & Dunkerton [2001].

To elaborate further on the possible subtropical connection, an ENSO index is included as well (Figure VII.11). While the $\text{QBO} \rightarrow \text{NAO}$ influence at 1 month lag remains, there also are pathways suggesting an ENSO influence on the winter NAO; either directly with a lag of 1 year or indirectly via both the QBO and the subtropical jet. Especially the latter is strong, given both links $\text{ENSO} \rightarrow \text{Subtropical Jet}$ and $\text{Subtropical Jet} \rightarrow \text{NAO}$ having a large value and being significant at the 99% level as well (not shown). The presence of the link $\text{QBO} \rightarrow \text{NAO}$ when including the Subtropical Jet in the analysis suggests that this interaction is independent of the subtropical troposphere, in contrast to the polar stratosphere (Figure VII.10).

Figure VII.12 combines the networks from Figures VII.10 and VII.11. Most of the links found in the smaller networks remain significant at the 95% level. The QBO is connected to the NAO via contemporaneous and directional links: $\text{QBO-PV upper} \rightarrow \text{PV lower-NAO}$. The subtropical jet, being influenced by ENSO, modulates the QBO and lower stratospheric polar vortex. Additionally, it is directly linked to the NAO, which is due to the close connection of the subtropical jet and the southern node of the NAO. This analysis suggests that the QBO influence on the NAO takes part via the polar route, while subtropical influences happen separately. As the subtropical jet is found to have an influence on the QBO at 1 month lag as well as a direct connection to the NAO, this might be mistaken by different analysis methods as a $\text{QBO} \rightarrow \text{NAO}$ influence happening via subtropical jet modulation.

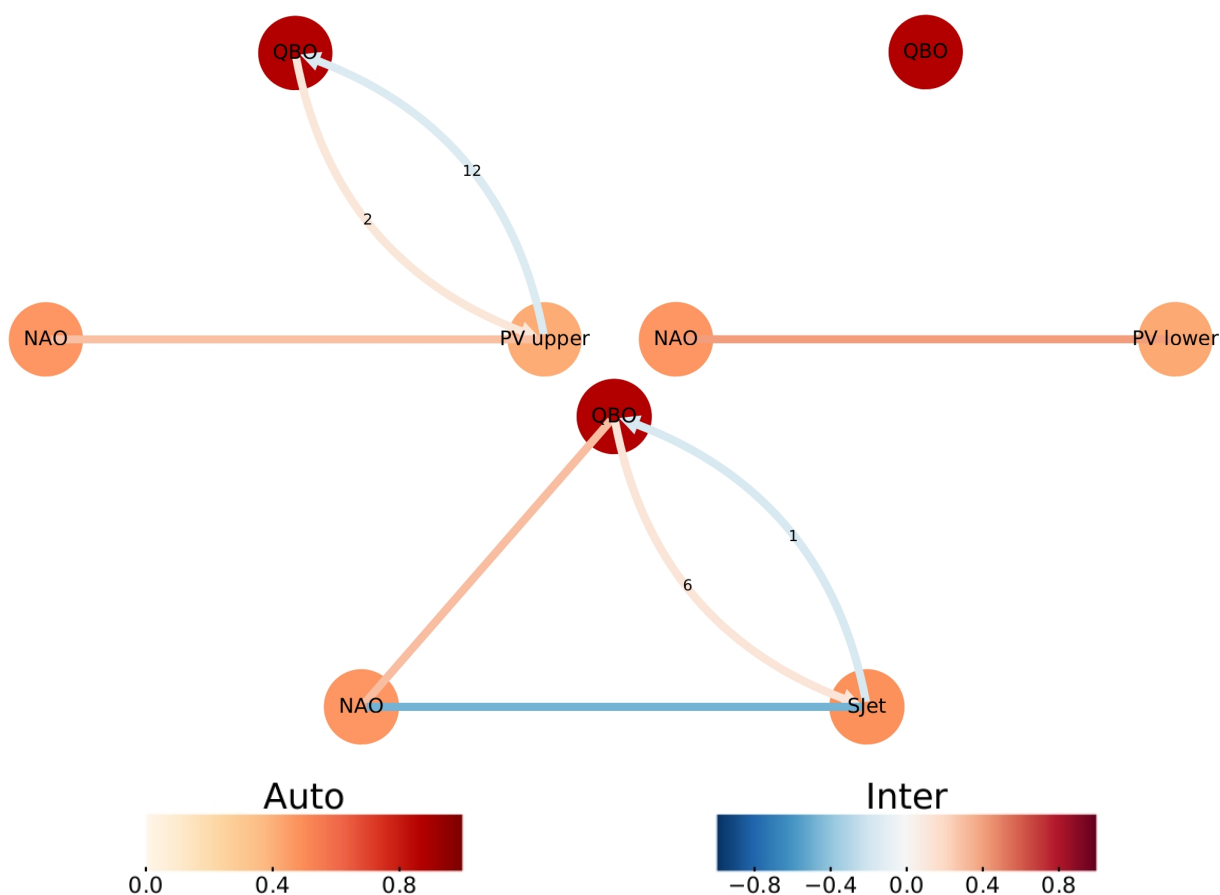


Figure VII.9.: Network graph (95% significance) establishing the QBO-NAO interaction route. Only the NAO in December and January is included. Top left: Adding the upper stratosphere polar vortex index. Top right: Adding the lower stratosphere polar vortex index. Lower: Adding a subtropical jet index. Indices as defined in Table III.1.

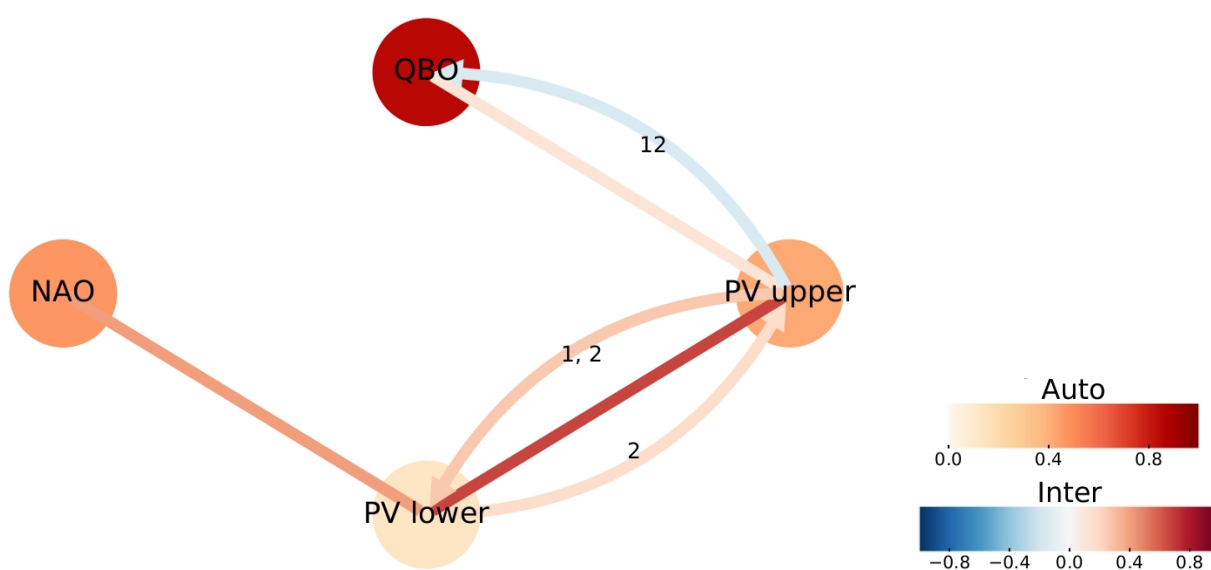


Figure VII.10.: Same as Figure VII.9, using the upper and lower stratospheric polar vortex in the network graph calculation.

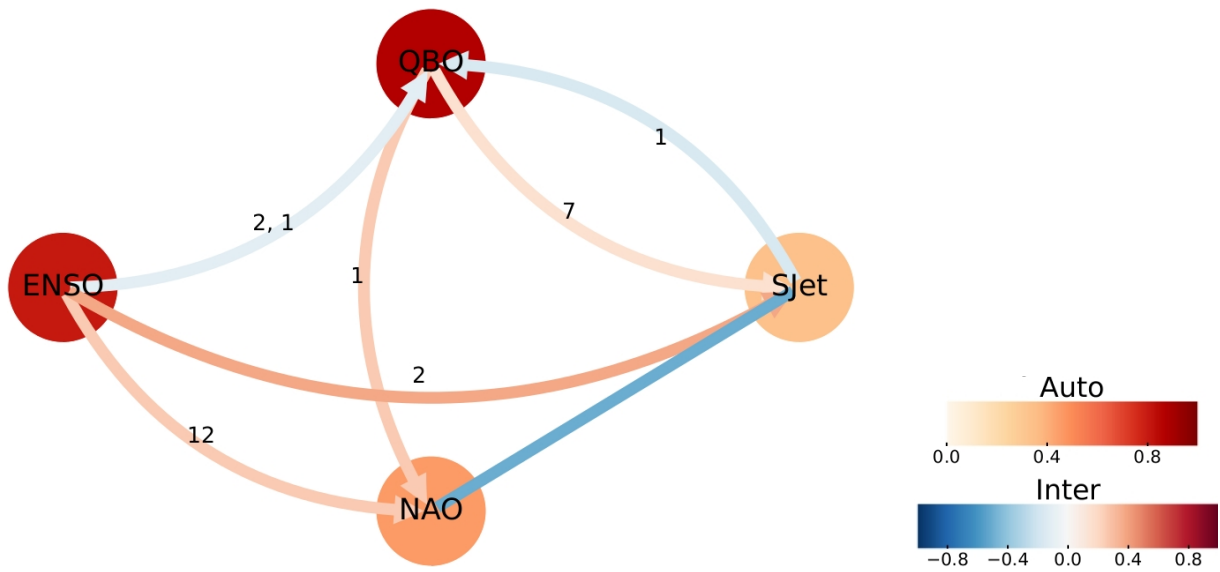


Figure VII.11.: Same as Figure VII.9, using the subtropical jet index and the ENSO index in the analysis.

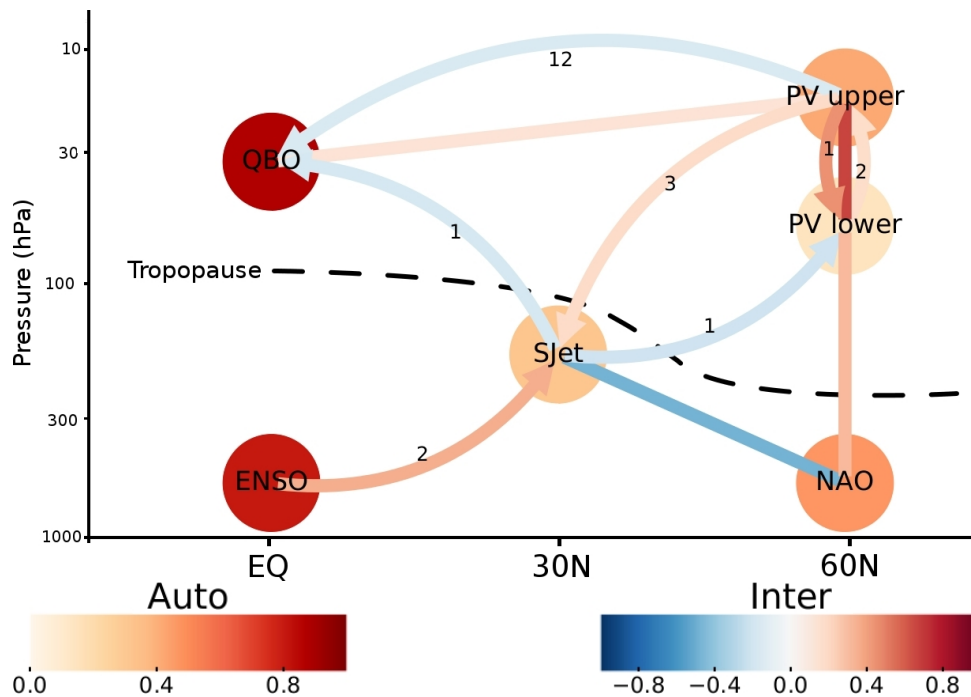


Figure VII.12.: Combining the networks from VII.10 and VII.11 into 1 network. Only the influence on NAO in early winter (DJ) is taken into account. The indications of latitude and pressure are for orientation and not an accurate representation of the atmosphere; approximate tropopause height is drawn dashed.

VII.4. The tropical network

Apart from the teleconnection from the tropical stratosphere (QBO) to the mid-/high-latitude troposphere (NAO), there is another possible network of processes the QBO is very likely to be involved in, located in the tropics only. The QBO, ENSO and the MJO are all, despite being named “oscillation”, not classical oscillations in the sense of fulfilling the harmonic oscillator equations. Nonetheless, all these processes involve modulating and being modulated by atmospheric waves to some extent.

This connection has been brought into focus rather recently (e.g. Moss *et al.* [2016]; Yoo & Son [2016]), partly due to the possibility of observing tropical waves. One of the aims from understanding the interdependency of the tropical processes is, like in the Northern Hemisphere winter teleconnection, the use of the QBO to improve longer range forecasts, more specifically of the Indian Monsoon (e.g. [Claud & Terray, 2006]).

In the tropics, the QBO and ENSO are probably the most dominant climate patterns. Their interaction has been studied extensively, with mixed results; Schirber [2015] mentions four studies that do not find an ENSO signal in the QBO, in contrast to one observational and one model study that does. The idealized simulations in Schirber [2015] show modulation of the QBO evolution depending on the state of ENSO. On the other hand, the QBO might modulate the onset of ENSO, as described by Gray *et al.* [1992].

Figure VII.13 gives an account of the various links between the different states of both QBO and ENSO. The wind at 50hPa is chosen as QBO index as it is likely that any influence from ENSO on the QBO will affect the lower layers of the stratosphere most. Schirber [2015] describes a weakening of lower stratosphere Easterlies and strengthening of Westerlies during El Niño conditions (i.e. a more westerly state of the lower stratosphere) due to increased wave activity. Both positive and negative states of ENSO and easterly/westerly states of the QBO are considered in both the driving variable X and the driven variable Y of the causal network analysis.

Figure VII.13 shows these eight different experiments as well as the network obtained from using both timeseries unaltered (panel (i)). A weak negative link QBO→ENSO at 4 months lag is found in the unmasked analysis, however, when taking the forcing from only one QBO phase into account (panels (a) and (b)), this link is not significant at the 95% level. Instead, a positive link QBO→ENSO at a lag of 11 months is found for QBO East, and none for QBO

West. When looking at the influence on the different phases of the QBO (panels (c) and (d)), an ENSO→QBO link is detected for the West phase, but not for the East phase. ENSO seems more influential on the QBO in its negative state (panel (e)), with a negative link at 4 months lag to the QBO, but no link in its positive state (panel (f)). Finally, the QBO is found to have more influence on the negative state of ENSO with a weak positive link at 10-12 month lag. In all the networks, the strength of the links, if any are found, are weak compared to the autocorrelation of the processes.

From this analysis, it is not possible to give a conclusive statement about the QBO-ENSO interaction. The relationship is probably a more complicated one than just a translation of sea surface temperature to zonal wind, which are the timeseries used by the algorithm. Inclusion of indices for wave fluxes or deep convection in the network could possibly shed more light on the QBO↔ENSO relation.

Still, numerous studies have used the QBO wind at one height and the SST timeseries in correlation and composite analyses to study QBO-ENSO interactions and implications for other tropical (and extratropical) processes. The tropics will be explored a bit further in the following.

Figure VII.14 shows the network graphs using the QBO index for $\varphi = -40^\circ$, the Niño 3.4 ENSO index and the OLR MJO index in the four seasons and all year round. All shown links are significant at the 95% level. The networks differ quite a bit during different seasons: The QBO→MJO link only appears in DJF, having a lag of 1 month, but not in any other season. Thus, in the all month analysis, this link appears weaker than in the seasonal analysis. While ENSO has an influence on the QBO all year round, an influence on the MJO is only detected in JJA at a lag of 7 months. This could be interpreted as ENSO having an influence on the MJO during its peak in boreal winter.

The links detected in this analysis, especially between the QBO and the MJO support the results of Yoo & Son [2016], who claim a connection between the boreal winter MJO amplitude and the QBO. However, they apply composite analysis, use a QBO index of wind at 50hPa and smooth all timeseries by a 3 month running mean. To test their results within the CEN framework, the equatorial $\bar{u}$ timeseries at 50hPa is used in a repeat of the above analysis. This index is highly correlated with the QBO $u^*(\varphi = -40^\circ)$ timeseries with a correlation coefficient of 0.91 (compare Table IV.5). Therefore, the results should be similar to the above findings presented in Figure VII.14 (a). Figure VII.15 shows the network for DJF using the QBO at 50hPa, ENSO and MJO index. The weak ENSO→QBO link disappears in this setup. However,

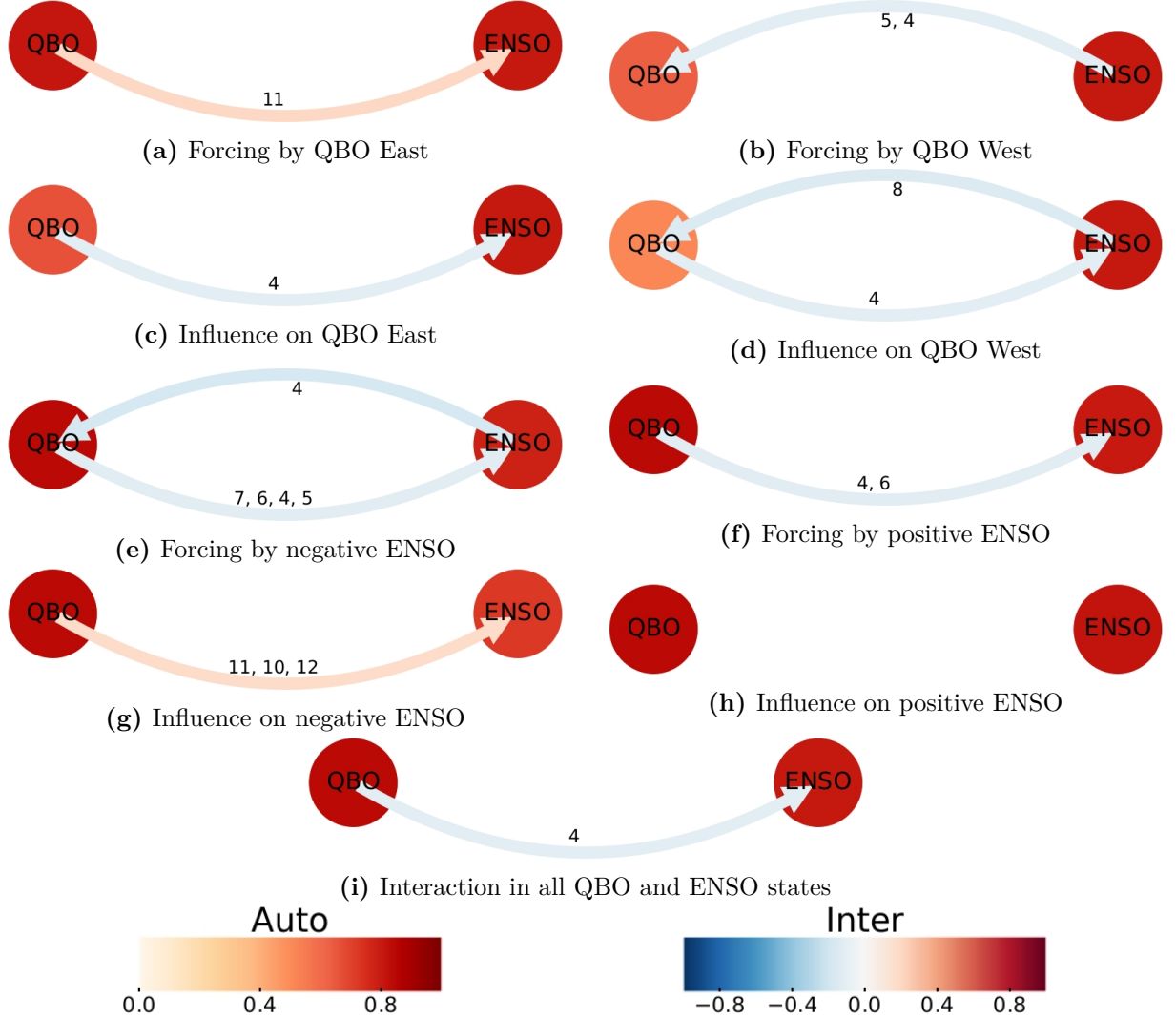


Figure VII.13.: Bivariate networks (95% significance) of QBO and ENSO monthly timeseries for different maskings: “Forcing by” means that in the respective timeseries all but the relevant state is masked in the driving variable X , “Influence on” describes masking of all but the relevant state in the driven variable Y . (i) depicts the QBO-ENSO network without any masking.

the QBO→MJO influence remains significant, with the lag changing from 1 to 2 months. Additionally, the MJO→QBO link is significant at the 95% level at a lag of 1 month. The networks in the other seasons look similar to the ones in Figure VII.14 (Figure B.6).

One question is whether the QBO-MJO connection also accounts for possible modulation of the Indian Monsoon by the QBO. Preliminary regression analysis analogous to the one shown in Figure VII.1 shows significant QBO regression coefficients in JJA for $\varphi \sim -60^\circ$ (negative coefficient) and in SON for $\varphi \sim 60^\circ$ (positive coefficient). However, the only QBO-Monsoon links detected with the CEN framework that are significant at the 95% level appear in DJF (Figure VII.16). Given the low autocorrelation of the Monsoon, it is unlikely that the boreal winter link influences the Monsoon in its active season. Nonetheless, the QBO-Monsoon network is expanded by the ENSO and the MJO timeseries to assess the possible tropical connections. The QBO index is set to $\varphi = -60^\circ$.

Figure VII.17 shows the networks equivalent to Figure VII.14, including the additional Central Indian rainfall Monsoon index. The slightly different links to the QBO are due to the different phase shifts. What is important, however, is that there are hardly any significant links to the Monsoon. In DJF, there is an indirect QBO→Monsoon influence via the MJO and in SON, ENSO acts as a common driver of both QBO and Monsoon rainfall, at similar lag times of about half a year. Sanity checks using $u^*(\varphi = 60^\circ)$ give similar results to the ones shown in Figure VII.17, especially the lack of links to the Monsoon. Looking at the full Monsoon season June-September leads to similar networks, as does taking only the late season (July-September or August-September) into account.

However, a single month analysis of the summer and autumn Monsoon (July-October) does show some links, as presented in Figure VII.18. While there is no connection of any of the processes to the Monsoon in June, the QBO has a strong influence on the Monsoon in July and August, both with a lag of 2 months, but of opposite signs. In September, ENSO seems to be a stronger factor and in October, the MJO. However, both of these links are at high lags (12 months/8 months respectively). In November, ENSO influences the Central Indian rainfall, the link being significant at a lag of 2 months.

These rather mixed findings make it questionable how reliable the QBO as a factor in Monsoon forecasting really is. The correlations found in various studies [Bhalme *et al.*, 1987; Claud & Terray, 2006] might be due to another factor influencing both QBO and Monsoon. Read & Castrejón-Pita [2012] suggest that both the QBO and the topospheric biennial oscillation (which

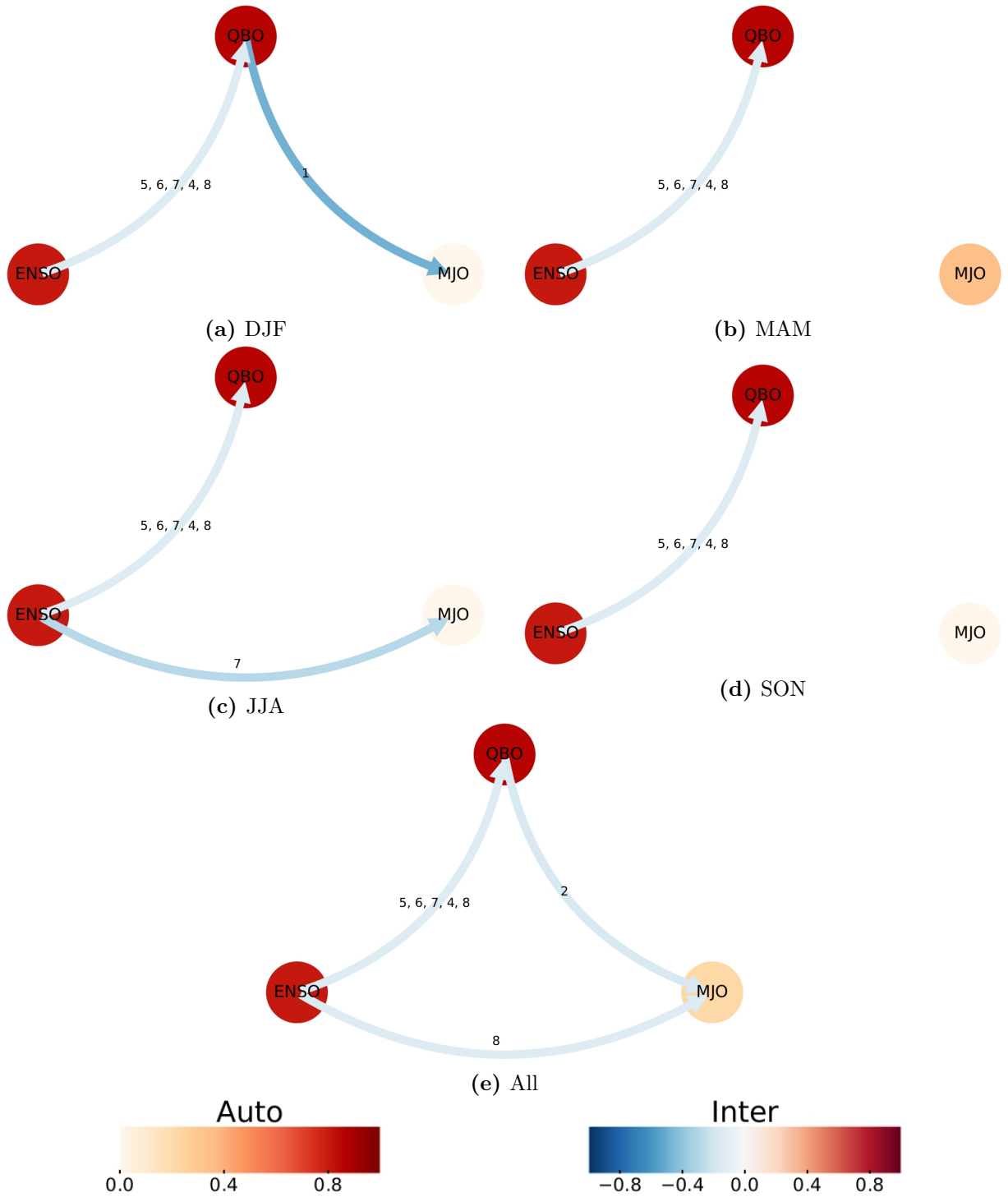


Figure VII.14.: Network graphs (95% significance) for the tropical processes QBO, ENSO, MJO, using either seasonal data for the latter (a-d) or an all year timeseries (e).

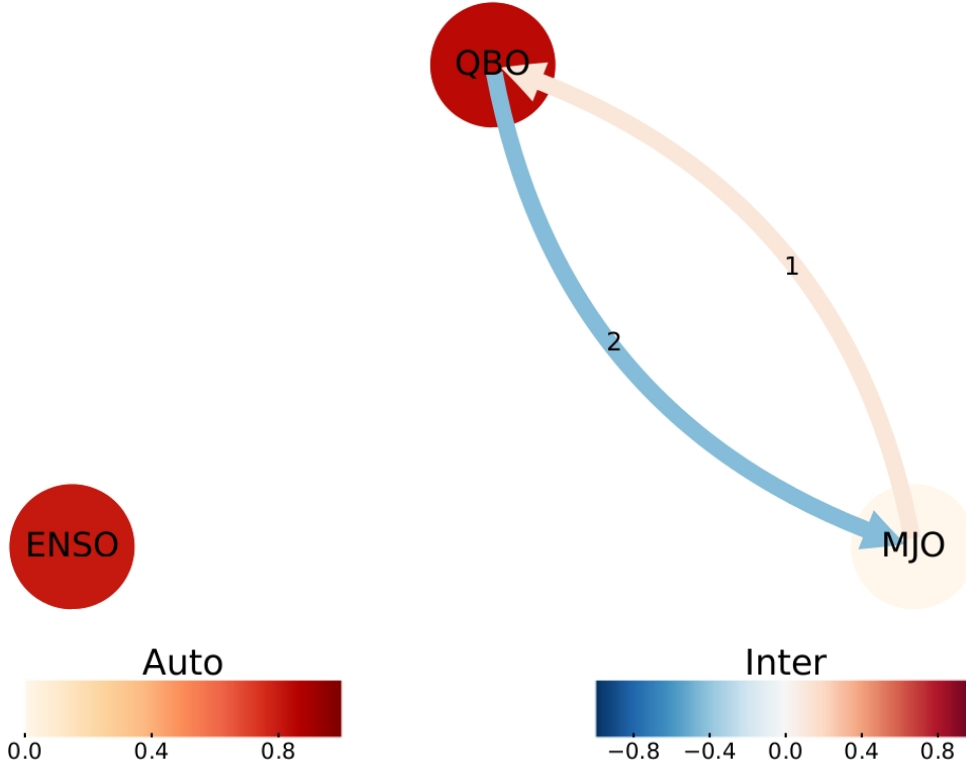


Figure VII.15.: As VII.14 a), using the equatorial zonal mean zonal wind at 50hPa as QBO index.

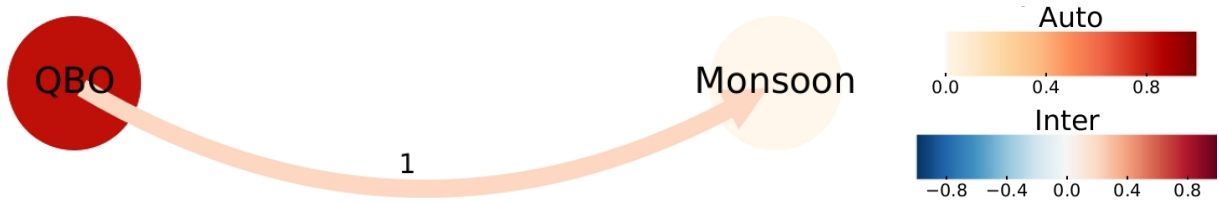


Figure VII.16.: Bivariate network graph (95% significance) using QBO index $u^*(\varphi = -60^\circ)$ and the central Indian rainfall in DJF as Monsoon index.

is linked to Monsoon wind/rainfall) are synchronised separately with the semi-annual oscillation and possibly an annual cycle as well. One difference between the studies finding QBO-Monsoon correlations and the one presented here is the fundamentally different approach from a simple linear one (correlation analysis) to a nonlinear one that also takes into account other processes. Some discrepancy of the results would therefore be expected. Regarding the influence of other processes (ENSO/MJO), the lack of connections to the Monsoon is rather surprising. However, they might happen on shorter timescales and therefore could not pass the significance test in a monthly mean analysis.

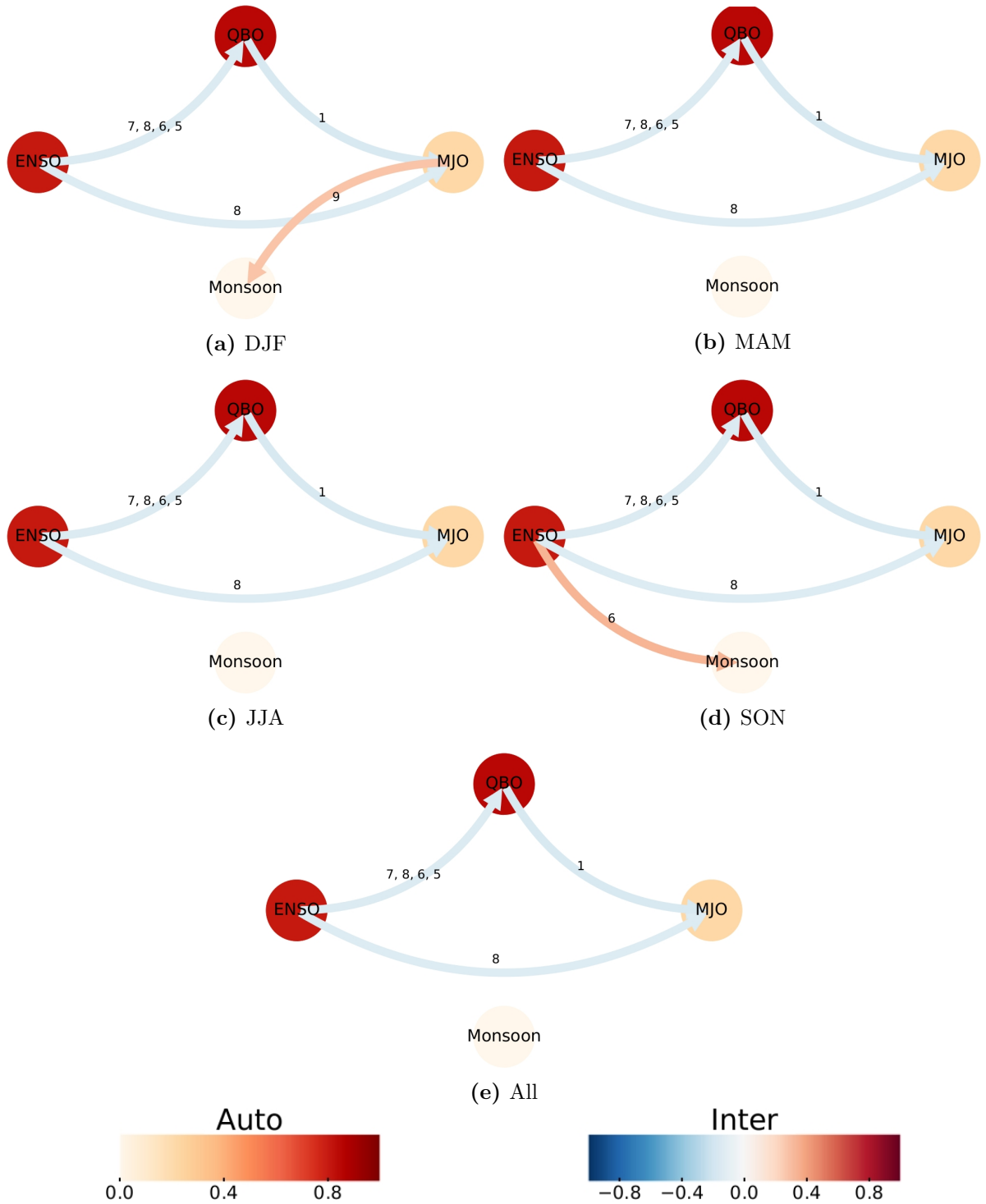


Figure VII.17.: As Figure VII.14, but including the Central Indian rainfall as Monsoon index and using the QBO index $u^*(\varphi = -60^\circ)$.

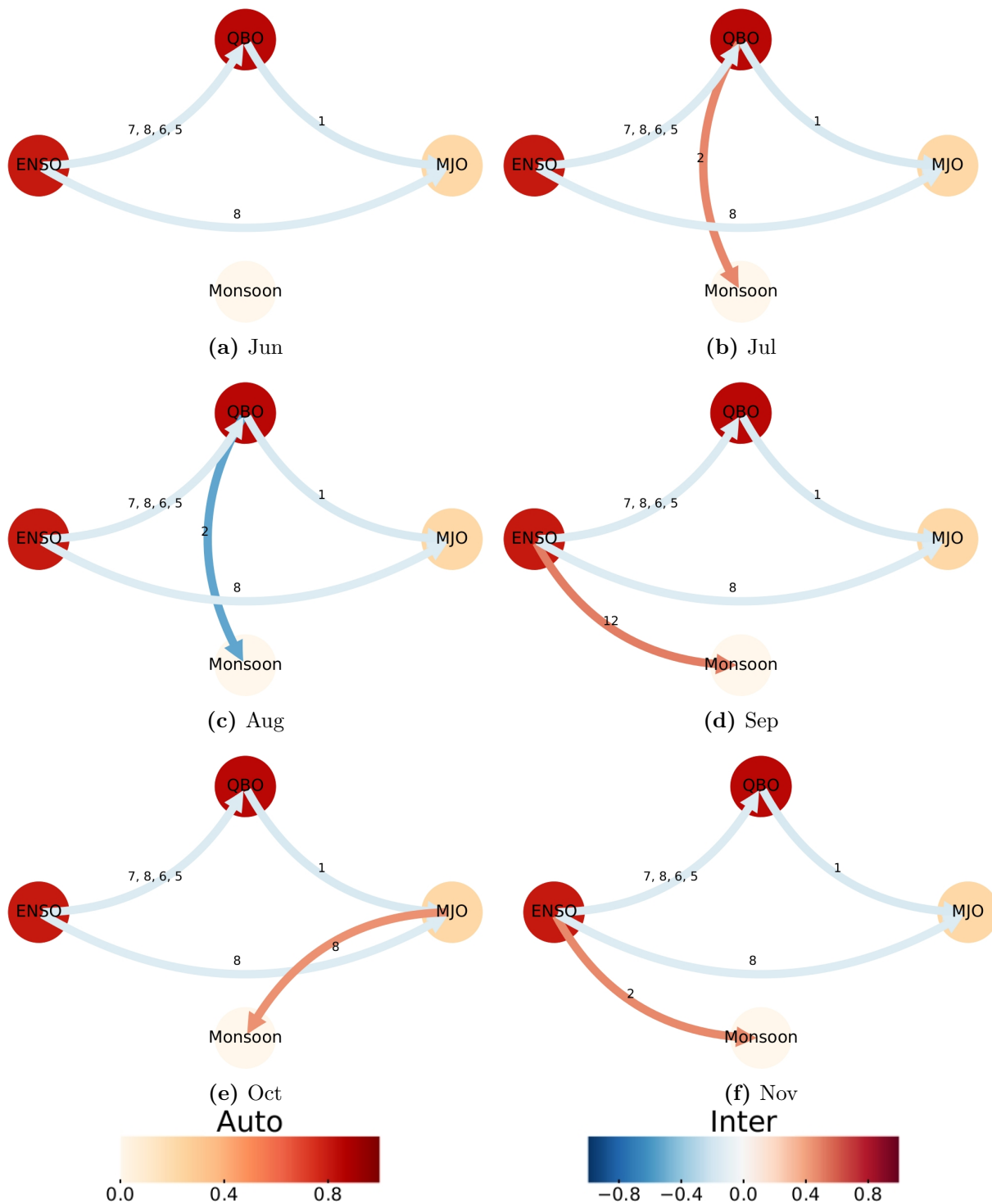


Figure VII.18.: As Figure VII.17, but looking at influences on the Central Indian Monsoon in single months during boreal summer and autumn.

VII.5. Higher time resolution

Given the low lag connection between the QBO and NAO/PV on a monthly timescale, the extratropical influence of the QBO is attempted to be resolved in higher time resolution data, which is also a test for the robustness of the above findings. Fortnightly and weekly setups are tested, averaging daily data 1^{st} - 15^{th} and 16^{th} - 30^{th} (1^{st} - 14^{th} / 15^{th} - 28^{th} for February) and 1^{st} - 7^{th} / 9^{th} - 15^{th} / 16^{th} - 22^{nd} / 24^{th} - 30^{th} (February 1^{st} - 7^{th} / 8^{th} - 14^{th} / 15^{th} - 21^{nd} / 22^{th} - 28^{th}), respectively. The maximum lag is set to 24 fortnights/48 weeks equalling the 12 months of the monthly analysis. For comparability with the monthly analysis, the significance level α is adjusted to account for the difference in maximum lag τ_{max} , following Kretschmer *et al.* [2016]: $\alpha^* = \frac{N_r}{N} \cdot \alpha$ where $N = n^2(\tau_{max} + 1) - n$, with n the number of included variables, denotes the number of tested links. N_r is the reference number from the monthly bivariate analysis for which $\tau_{max} = 12$, $n = 2$.

Figures VII.19 and VII.20 show the QBO-NAO networks equivalent to the monthly mean analysis at higher time resolution (95% significance), testing the links found for robustness, trying to resolve contemporaneous links and checking for additional interactions that happen on sub-monthly timescales. The QBO-NAO interaction is examined using timeseries representing the polar route (analogous to Figure VII.10) and the subtropical route (Figure VII.11). While on fortnightly resolution, a contemporaneous weak link between the QBO and the NAO appears, the path QBO→PV found in the monthly analysis disappears, which is quite surprising. Instead, the inverse influence of the upper stratospheric polar vortex becomes significant at lags of 2 and 9 fortnights. The lower stratospheric vortex is again directly linked to the NAO on fortnightly resolution, but not on the weekly timescale. At the 90% significance level (not shown), however, there is a negative link from PV lower→NAO at lags of 6 and 1 weeks. The backward influence from the upper polar vortex to the QBO appears at weekly resolution as well, with lags of 17 and 3 weeks being significant (which is in accordance with the links found at fortnightly resolution with lags of 2 and 9 fortnights). The only robust link in this network is, unsurprisingly, the connection between the upper and lower part of the stratospheric polar vortex, which is a downward influence at very low lag times.

In the network including ENSO and the subtropical jet (like Figure VII.11), the (negative) influence of the Jet on the NAO is significant at 95% at fortnightly and weekly time resolution, as is the ENSO→Subtropical Jet link. The QBO is again not playing a major role in the network,

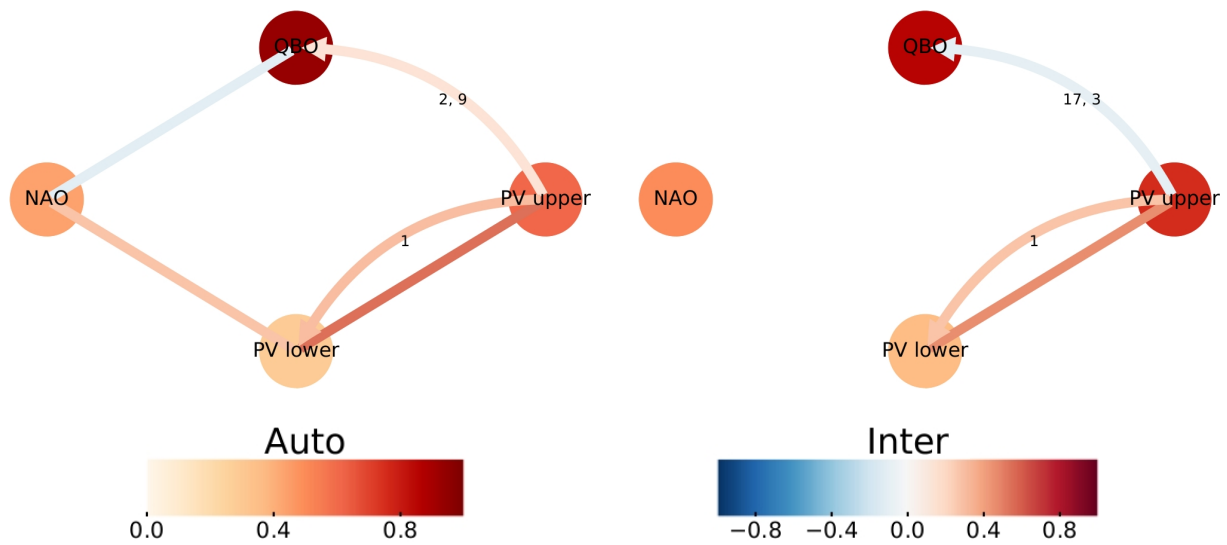


Figure VII.19.: Same as Figure VII.10, but at fortnightly (left) and weekly sampling (right). Note that the lags displayed refer to the respective time resolution.

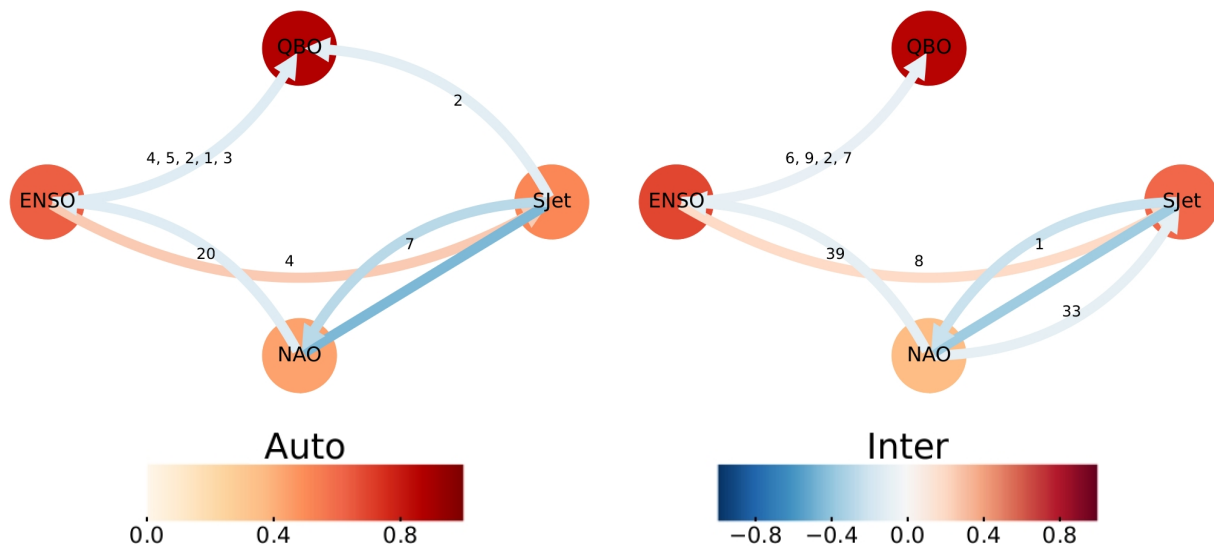


Figure VII.20.: Same as Figure VII.11, but at fortnightly (left) and weekly sampling (right). Lags refer to the respective time resolution.

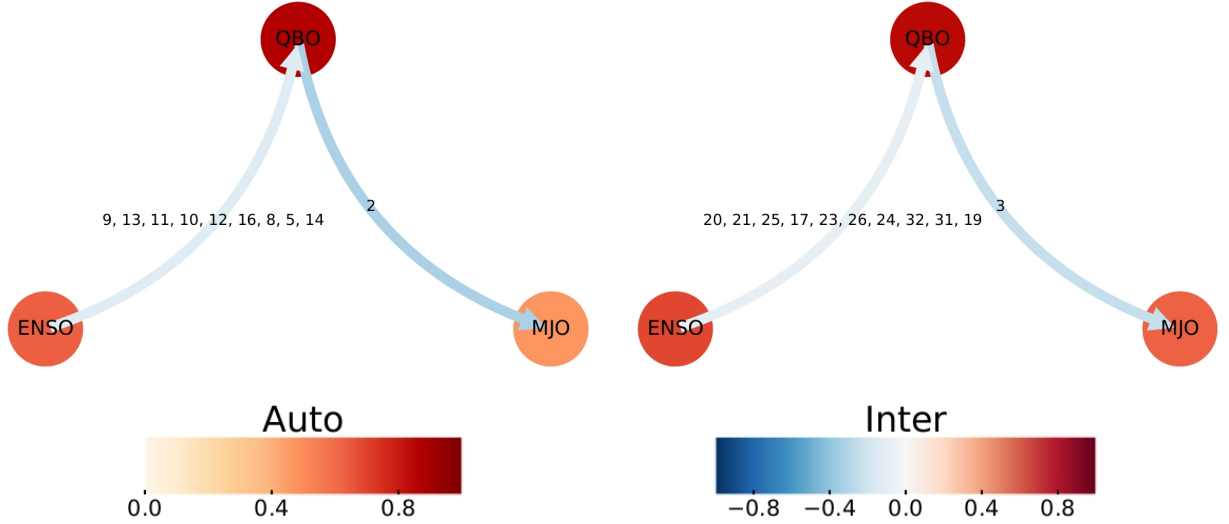
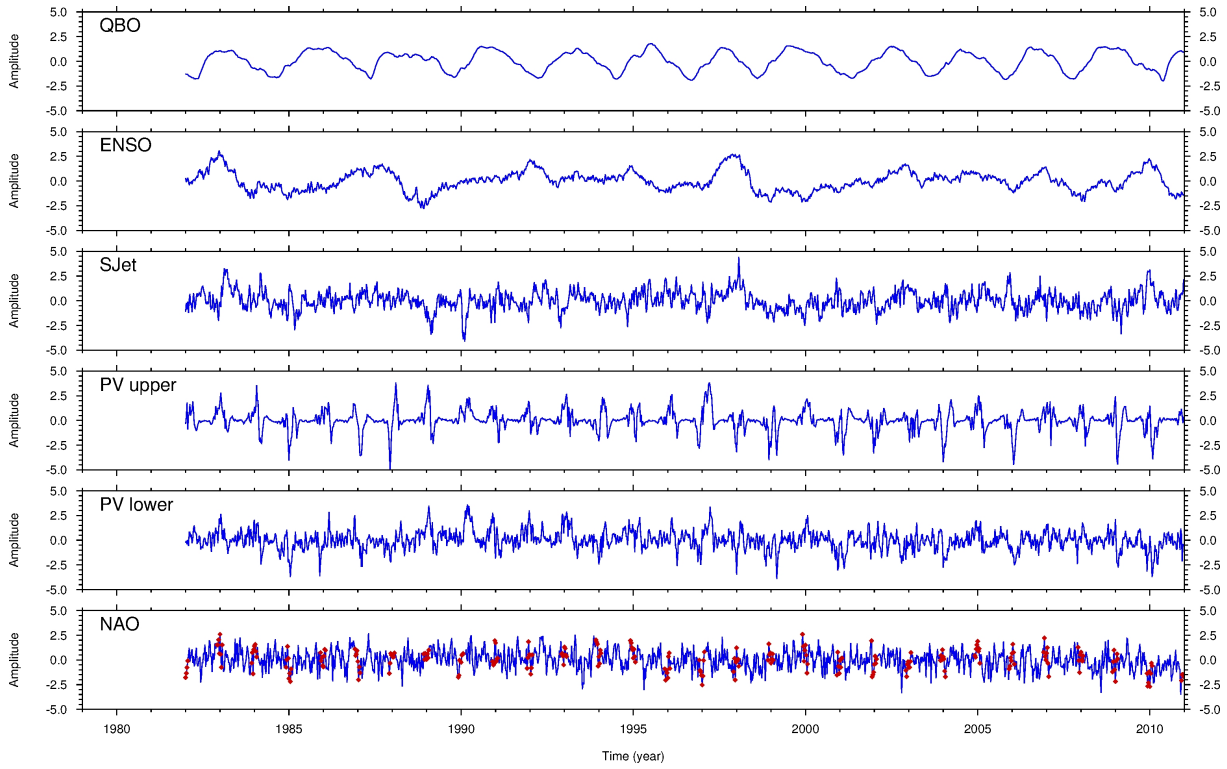


Figure VII.21.: Same as Figure VII.14 a), but at fortnightly (left) and weekly sampling (right). Lags refer to the respective time resolution.

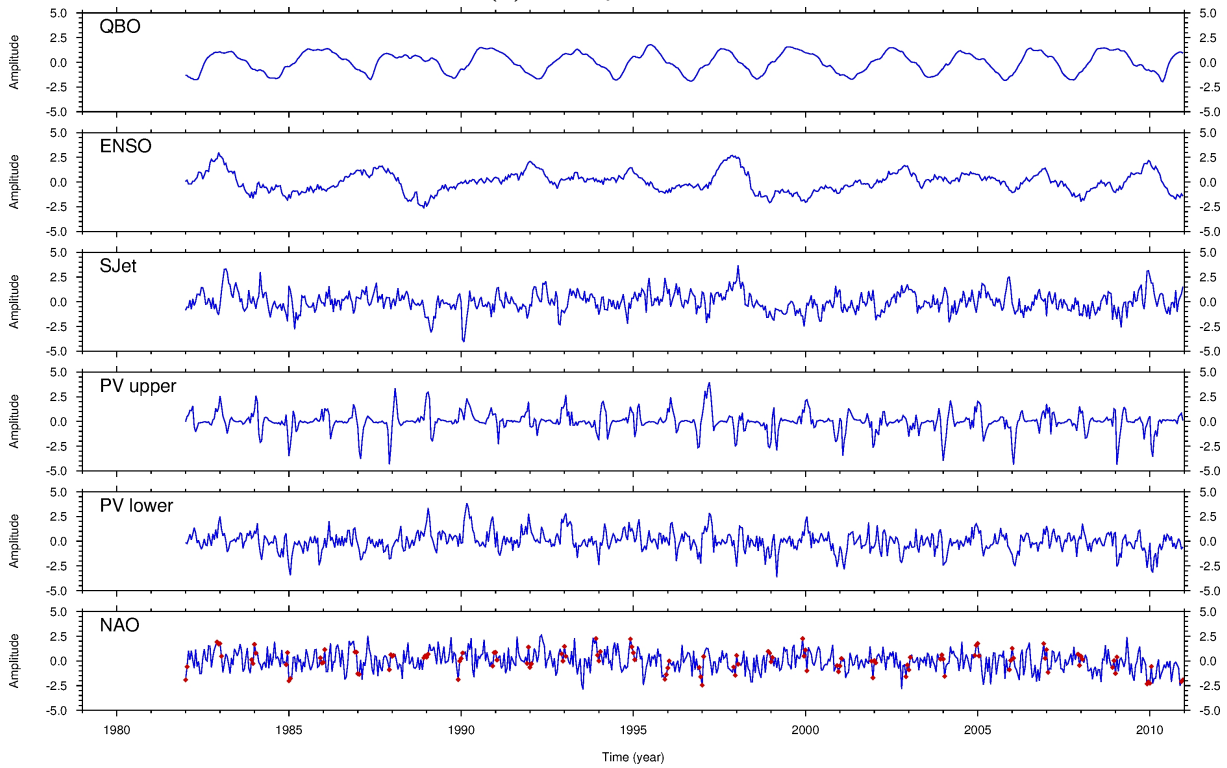
with no links from the QBO being significant at 95%. The weak ENSO→QBO link, however, is robust at similar lag times as in the monthly analysis.

The lack of QBO influence on the NAO on the higher time resolution might be due to the very long intrinsic QBO timescale and high autocorrelation. The jets and the NAO (and to some extent, ENSO), in contrast, vary on much shorter timescales (see the monthly mean, fortnightly and weekly timeseries in Figure VII.21). Thus, on shorter timescales other processes might be more influential on the NAO, like snow cover or sea ice, as explored by Kretschmer *et al.* [2016]. For QBO-NAO influences, monthly time resolution seems much more useful to assess connections. The differences in the networks when looking at different timescales could hint at possible interaction mechanisms: longer timescales as in the QBO-PV-NAO network (Figures VII.10 and VII.19) might be due to planetary waves travelling from the equator to the polar region, whereas the tropical connection, especially the Subtropical Jet→NAO link, could involve faster synoptic scale waves. The PV→QBO influence at a lag of 12 months in the monthly analysis could be due to a modulation of planetary wave propagation by the altered polar vortex state. This changes the strength of the Brewer-Dobson circulation, which itself influences upwelling at the equator. The latter is linked to the descent of the QBO; it takes about 1 year for each shear zone to descent through the stratosphere. However, this mechanism is speculative and requires further investigation.

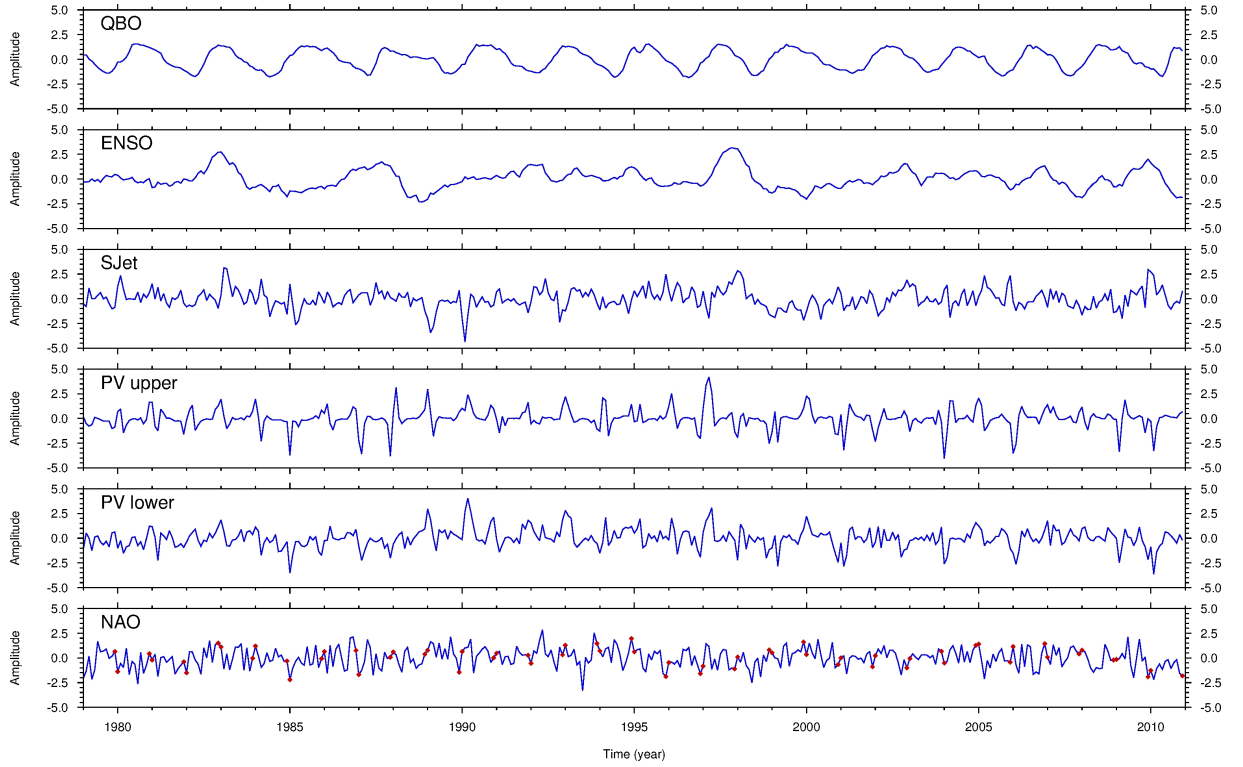
Additionally, higher levels of noise in fortnightly or weekly sampling might lead to reduced significance of the links found in the monthly analysis.



(a) Weekly resolution



(b) Fortnightly resolution



(c) Monthly resolution

Figure VII.21.: Timeseries (normalised) of the processes involved in the QBO-NAO interaction (Figures VII.19 and VII.20) at weekly, fortnightly and monthly sampling. Only December and January data (red dots) for the NAO is used in the analysis. The higher resolution timeseries start in 1982 due to availability of daily data from which these are calculated.

The tropical network involving QBO, ENSO and MJO in DJF (Figure VII.14 a)), is assessed as well on fortnightly and weekly timescales. Corresponding networks are shown in Figure VII.21. The QBO→MJO link is robust, staying significant at the 95% level at lags of 2 fortnights/3 weeks. There is an additional indirect influence from ENSO modulating the QBO, the link being significant at various lag times around 4-6 months as in the monthly analysis. The negative QBO→MJO link found is in agreement with the findings of Yoo & Son [2016], who explain it by the QBO modulating deep convection in the tropics. Including a timeseries of convection in the network could shed more light on this hypothesis.

Including the Monsoon in this network and repeating single month analysis as shown in Figure VII.18 at fortnightly and weekly resolution gives the networks in Figures B.7 and B.8. In contrast to the monthly mean analysis, ENSO now appears to be the driving factor of the other processes, directly influencing the QBO, which itself modulates the MJO, though with higher lag times than in the analysis without the Monsoon timeseries. In the autumn (Sep-Nov), ENSO also has a strong influence on the Monsoon rainfall, similar to the monthly findings.

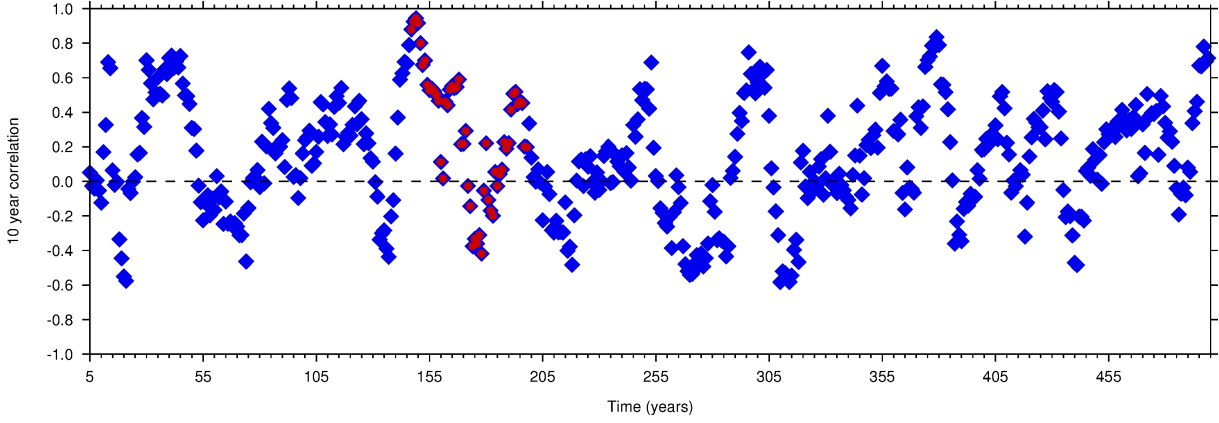


Figure VII.22.: 10 year running correlation between the winter (DJF) QBO and the polar vortex zonal winds for 505 years of the HadGEM2-CC control run (blue). The red dots show the 52 subsequent years with the highest correlation ($R = 0.40$) of any 52 year long subset, the closest equivalent to the observed value ($R = 0.49$).

The ENSO→QBO→MJO connection stay the same when increasing the resolution to weekly. Additionally to confirming the ENSO→Monsoon links in autumn, a direct link between ENSO and the Central Indian rainfall in these months appears. As ENSO is found in this analysis to modulate QBO and Monsoon alike, it would explain why some other methods find a connection between the QBO and the Monsoon. Similarly to the QBO-NAO connection, the modulation of the Monsoon by the QBO might only be relevant for monthly means, whereas on shorter timescales quicker varying processes like ENSO dominate the influence.

VII.6. Northern Hemisphere teleconnections in the Hadley Centre Model Control Run

Experiments equivalent to the ones described for the observational/reanalysis dataset above are carried out for the 505 years of the HadGEM2-CC control run. Its QBO representation is realistic (Table V.2) and the structure of the first two EOFs of zonal mean zonal wind in the equatorial stratosphere is comparable to the observations (Figure V.12). The Holton-Tan relationship is assessed in the same way as described by Lu *et al.* [2008]. The 10-year correlation between equatorial $\bar{u}$ and the polar vortex wind (60°N, 10hPa) is shown in Figure VII.22.

As seen for the observations (Figure VII.4), the QBO-PV correlation is not constant over time. In the control run, this leads to a correlation coefficient of $R = 0.13$ for the whole 505 year timeseries. The record is scanned for 52 years with the highest correlation, to find an equivalent subset of years to the observations ($R = 0.49$). These years are shown in Figure VII.22 as red

dots; the correlation coefficient is slightly less than in the observations: $R = 0.40$.

Timeseries from both the whole control run (“all”) and the 52 high correlation years (“HC”), are used for the network analysis to assess the Northern Hemisphere winter teleconnections in the control run. The definition of the timeseries (all monthly means) is as follows; they are shown in Figure VII.23:

- QBO: Reconstructed timeseries u^* from the first two EOFs at $\varphi = 20^\circ$
- ENSO: Average of sea surface temperature anomalies over the Niño 3.4 region
- NAO: First PC timeseries of the Northern Hemisphere (North of 20°) mean sea level pressure¹
- PDO: First PC timeseries of the North Pacific ($20^\circ\text{N} - 60^\circ\text{N}$, $120^\circ\text{E} - 150^\circ\text{W}$) sea surface temperature anomalies
- Polar Vortex and Subtropical Jet: Average over $\bar{u}$ as defined for the observations (Figure III.2).

Figure VII.24 shows the influences on the winter (DJ) NAO via the polar route (95% significance). As in the observations (Figure VII.10), the connection between the upper and the lower polar vortex is strong, as is the connection between the lower stratospheric polar vortex and the NAO. All other connections, however, are very weak. This applies in particular to the QBO influence on the upper stratospheric polar vortex. Using the HC years only slightly strengthens the QBO→PV connection to about the same level as in the observations. The downward influence from the upper to the lower PV is marginally weaker, while the contemporaneous link PV lower→NAO is strengthened again. Additional backward connections from the NAO and the lower polar vortex to the upper polar vortex, that are not present in the observations, appear in the HC years or are strengthened.

Including the PDO index in this network leads to the networks shown in Figure VII.25. In contrast to the observations, the PDO does not modulate the QBO→PV influence a lot. There are additional links in the network, showing weak connections of the PDO to all four other processes, the strongest ones are the PDO-lower PV and the PDO-NAO connection. Similarly, in the years of high QBO-PV correlation, the PDO does not influence the QBO interaction with the upper stratospheric polar vortex, but rather has an independent influence on the NAO and

¹This actually is an Arctic Oscillation index; however the NAO and the AO are strongly correlated.

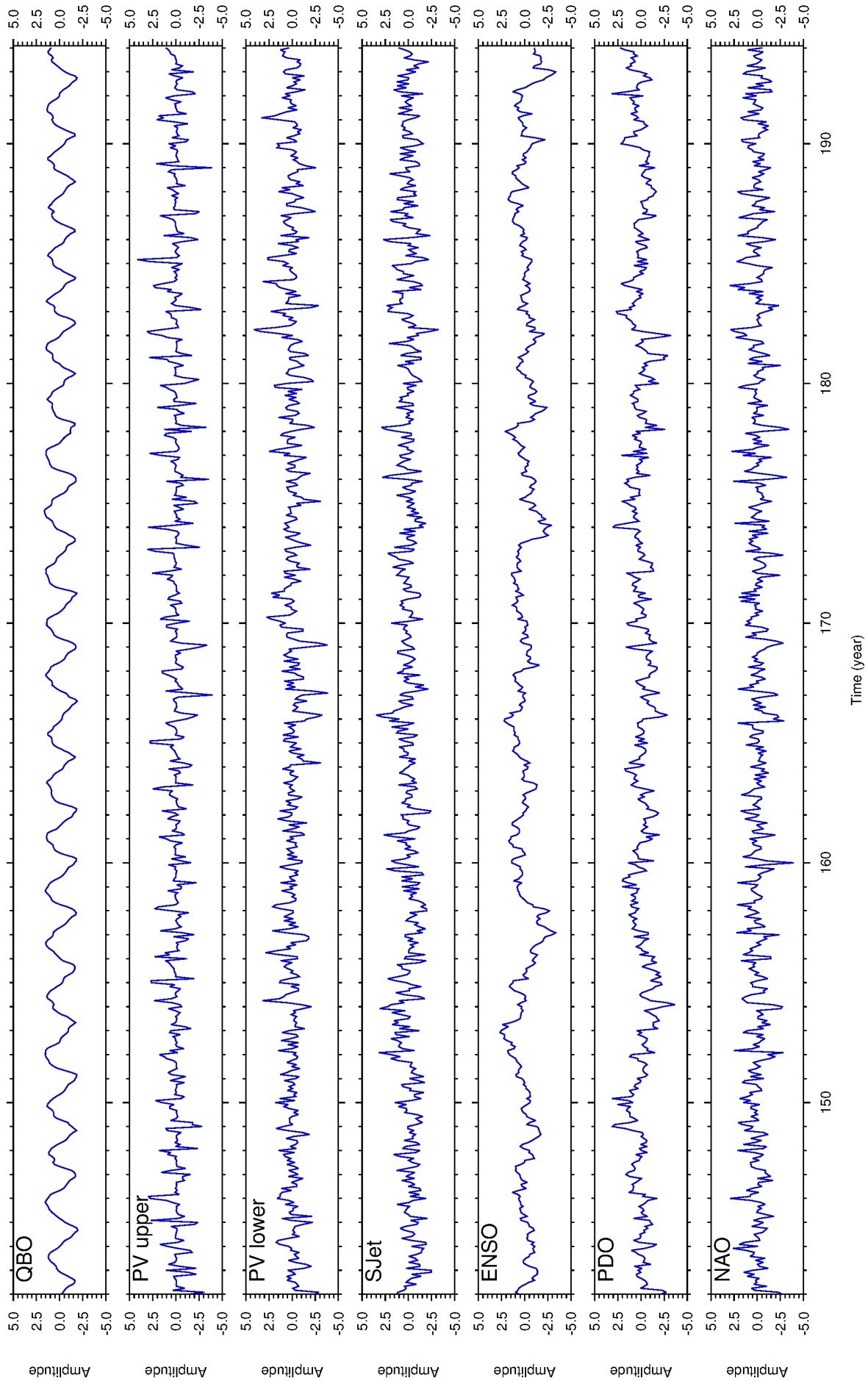


Figure VII.23.: Timeseries of the indices in the 52 HC years from the control run.

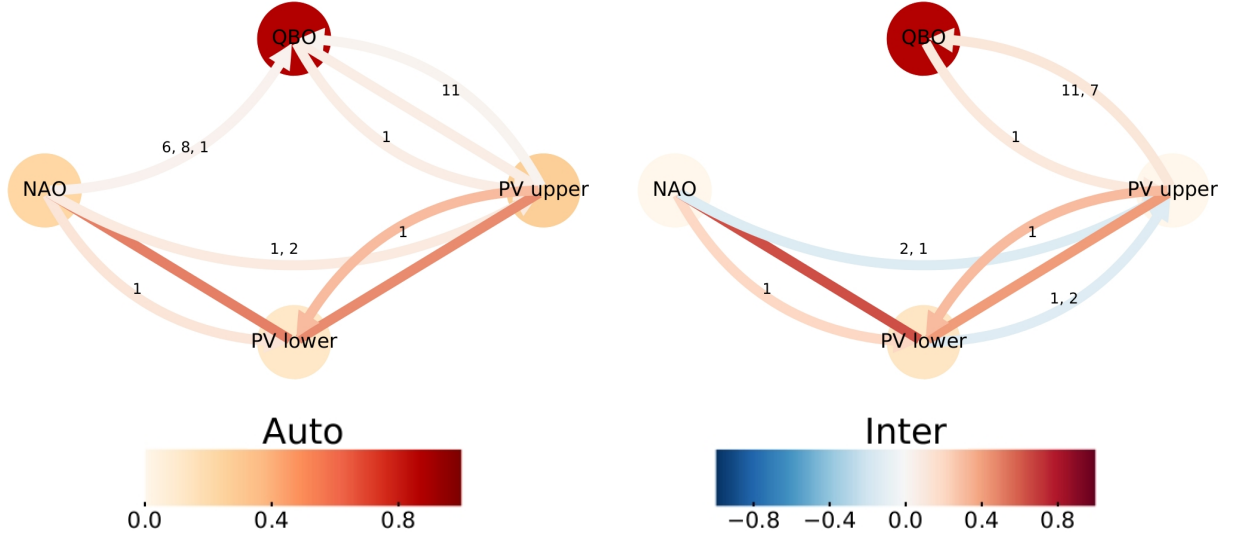


Figure VII.24.: Same as Figure VII.10, using HadGEM2-CC control run data. Left: full record, right: 52 high correlation years.

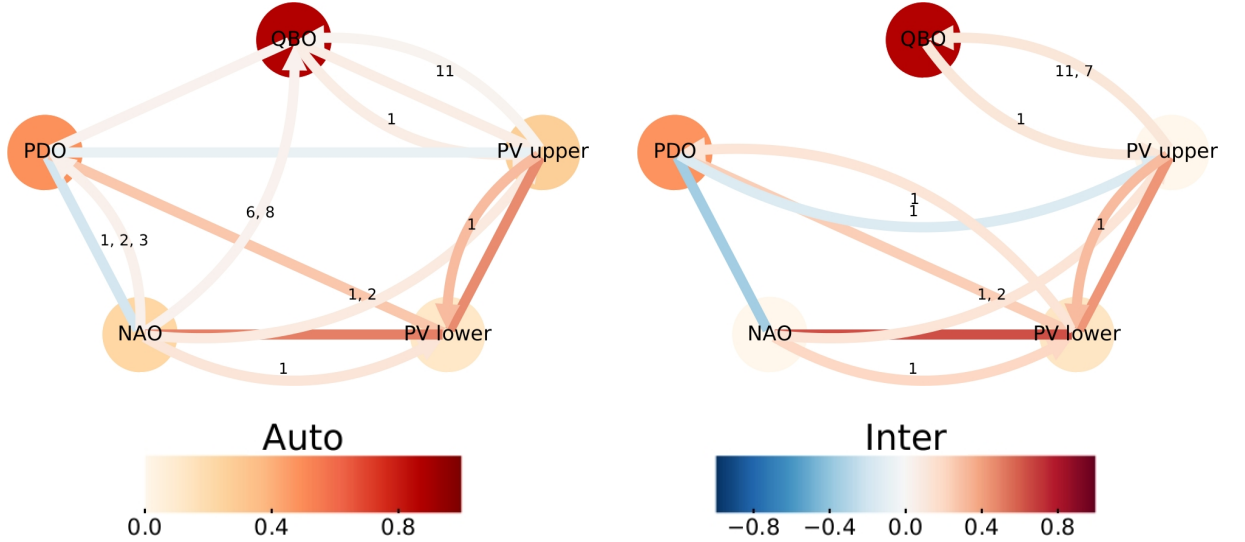


Figure VII.25.: Same as Figure VII.24, using an additional PDO index.

the lower stratospheric polar vortex. The QBO-NAO influence via the stratosphere remains unchanged.

For comparison, the QBO-NAO interaction via the subtropical route is analysed as well. Figure VII.26 shows the network graphs at 95% significance using the QBO, ENSO, NAO and the subtropical jet index. Only influences on the winter NAO (DJ) are taken into account. The network using the full 505 year record shows weak links from ENSO, NAO and SJet on the QBO; the QBO only influences the subtropical jet, but the connection is very weak. Stronger connections exist between ENSO and the subtropical jet, as well as the subtropical jet and ENSO and the NAO. The latter is in agreement with the observations. Using the HC years

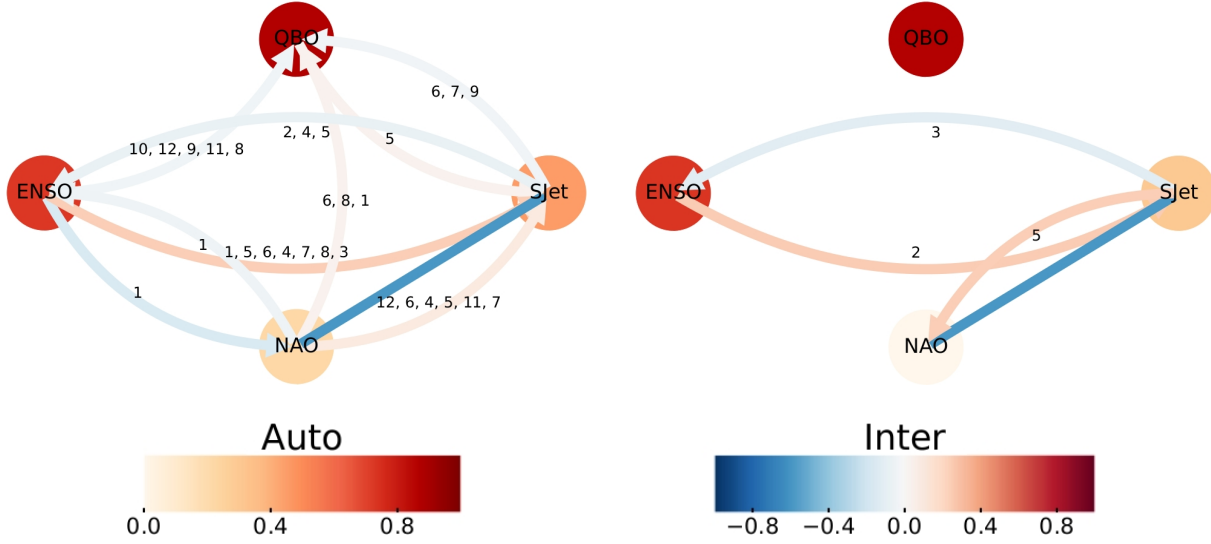


Figure VII.26.: Same as Figure VII.11, using HadGEM2-CC control run data. Left: full record, right: 52 high correlation years.

only, the QBO is disconnected from the other processes in the network; the interaction between ENSO and the subtropical jet at lag times of 2-3 months remains, as does the direct connection between the subtropical jet and the NAO.

Combining the networks from the two possible interaction routes in HC years gives the one shown in Figure VII.27, which is the control run equivalent to VII.12. As in the observations, the QBO interacts with the NAO via a direct connection to the upper stratospheric polar vortex and the lower stratospheric polar vortex. The connection strength between the lower polar vortex and the NAO is stronger than in observations. There is an additional direct link of negative sign between the subtropical Jet and the NAO, consistent with observations, as is the ENSO→SJet modulation with lag of two months. Furthermore, weak links are detected between the subtropical jet and the lower stratospheric Vortex as well as NAO→ENSO, which do not appear in the observations. Using all 505 years from the Control run weakens the links shown in Figure VII.27. The strongest links that remain are the PV upper-PV lower-NAO, SJet-NAO and ENSO→SJet ones. There are loads of weak links appearing between all processes in the network (Figure B.9). This is due to higher statistical detection power for large samples [Runge *et al.*, 2014]. Given the weakness of the links, especially compared to the ones found in the shorter record, it is questionable whether these interactions actually play a significant role in the atmosphere.

Overall, the findings from the Control run regarding the QBO-NAO interaction are consistent with the ones based on the observations (compare Figures VII.27 and VII.12). The analysis sup-

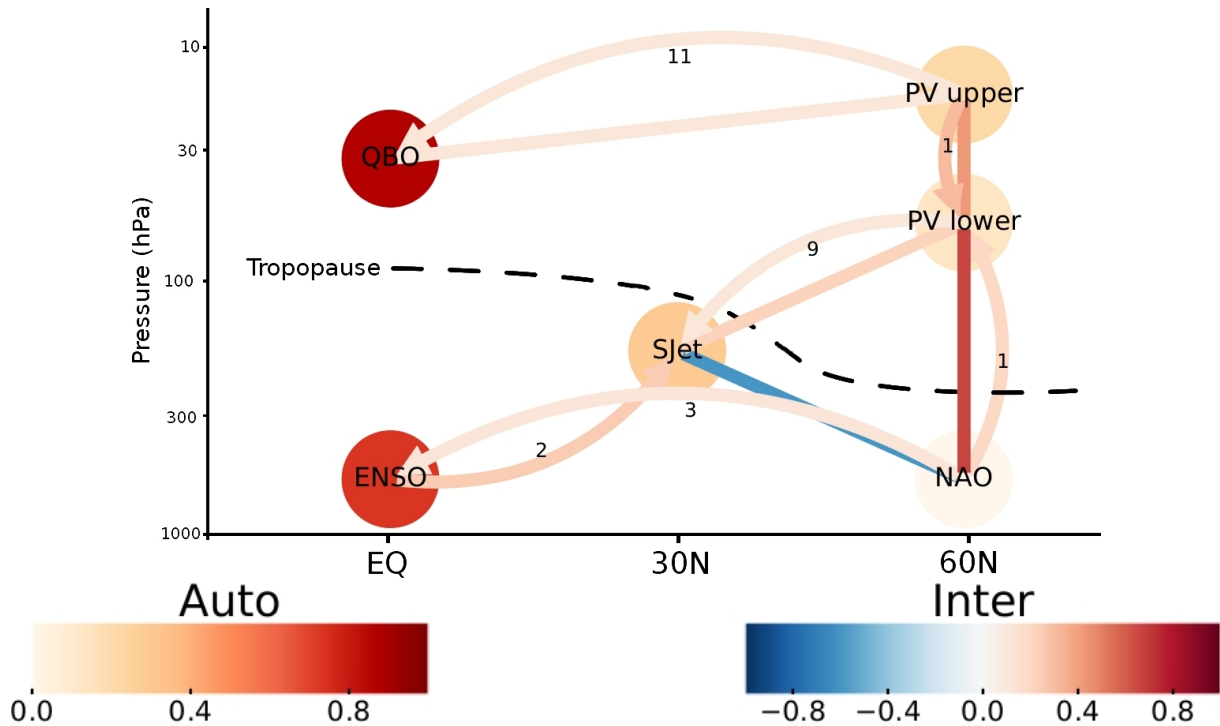


Figure VII.27.: Combining the networks from VII.24 and VII.26, using 52 HC years from the control run.

ports the hypothesis of the QBO modulating the NAO via the “stratospheric” route, influencing the upper stratospheric polar vortex; this modulation is then propagated down to the surface. The subtropical jet is closely connected to the NAO, which is expected, but appears not to be influenced by the QBO. ENSO indirectly influences the NAO via the subtropical jet.

VII.7. Chapter summary

Possible links between the QBO and other climate indices, namely the NH Polar Vortex, the NAO, the MJO and the Indian Monsoon, were tested using a recently developed framework called causal effect networks [Runge *et al.*, 2012a,b, 2014, 2015]. The network of the QBO and the NH winter was also assessed in the 505 year HadGEM2-CC control run.

In the observations, the link between the QBO and the Polar Vortex and surface NAO, first described by Holton & Tan [1980], is found in the analysis of monthly data (Figure VII.12). In the network, the QBO, upper and lower stratospheric Polar vortex, and the NAO are linked by contemporaneous links, meaning that it happens on monthly timescales or faster. Additionally to the QBO influence, the subtropical jet is closely linked to the NAO and is found to modulate the QBO and the lower stratospheric polar vortex, an interaction that cannot be distinguished by methods such as composite and regression analysis. ENSO is found to directly influence the

subtropical jet, but no other variables in the network.

The interactions found in the observations were also established in the control run using a subset of 52 years in which there is a high correlation between the QBO and the Polar Vortex (i.e. a Holton-Tan effect; Figure VII.27). In the full 505-year dataset, the QBO-Polar Vortex link is much weaker, whereas the Subtropical Jet-NAO link is robust, as well as the connection between the upper and lower stratospheric polar vortex and the NAO. This hints towards the model getting the basic interaction mechanism right, but means that the observed Holton-Tan effect of a high correlation between the equatorial stratospheric winds and the NH Polar Vortex (Figure VII.4) is not a constant feature in the atmosphere (Figure VII.22).

Attempts to resolve the QBO-NAO interaction at higher timescales (fortnightly and weekly) gave inconsistent results (Figure VII.10), even breaking the QBO-NAO connection. The reasons for the differences between monthly means and higher time resolution are still to be investigated. The subtropical pathway of ENSO influencing the subtropical jet and subsequently the NAO, is confirmed at fortnightly and weekly timescales (Figure VII.20).

In the tropical network, the link between the QBO and the MJO is robust in boreal winter, consistent with the findings of Yoo & Son [2016]. ENSO, however, is only indirectly influencing the MJO via the QBO. QBO-ENSO interactions assessed in a bivariate analysis using only these two indices, are inconclusive (Figure VII.13). One reason could be the relatively short dataset, given that ENSO varies on comparably long timescales (Figure VII.21).

The interaction between the QBO and the Indian summer Monsoon is questionable as well (Figure VII.18). A possible link only is found in July and August and is of opposite sign. This puts limits on the usefulness of the QBO in Monsoon forecasting, despite its long-standing incorporation in these [Bhalme *et al.*, 1987; Claud & Terray, 2006]. However, Read & Castrejón-Pita [2012] suggest a separate coupling of the QBO and the Monsoon to the semi-annual oscillation which subsequently appears as a correlation between the two. This theory could be further explored with the CEN method by incorporation of an SAO term in the analysis and not deseasonalizing the data.

VIII. Quest from bizarre Observations

VIII.1. Unprecedented behaviour of the QBO

Since systematic observations of the equatorial zonal wind began in 1953, the QBO has exhibited a fairly regular cycle (Figure II.2). However, the winter 2015/16 saw very unusual behaviour as shown in Figure VIII.1: Instead of westerlies descending, as observed during the winter 2013/14, the shear zone showed an upward motion, giving rise to a layer of easterlies near 40hPa. Over the course of 2016, this easterly strengthened, halting the descent of the overlying westerly. With the westerly winds in the lower stratosphere (70-50hPa) weakening in mid 2016, the high-stratosphere westerly resumed its descent, due to the fast waves carrying westerly momentum not getting absorbed anymore in the lower stratosphere. From boreal summer 2016 onwards, the structure of the stratospheric winds looks like a regular cycle, with just two shear zones between 10 and 70hPa.

Newman *et al.* [2016] and Osprey *et al.* [2016] were the first studies to describe this unprecedented behaviour of the QBO. While Newman *et al.* [2016] point out that the easterly phase at 10hPa and the westerly phase at 40hPa are the shortest in the 63 year radiosonde record, Osprey *et al.* [2016] classify the irregularity of the January equatorial zonal winds by looking at the zonal wind profiles: To each profile, a closest match is found by minimizing the sum of the squares of the differences in $\bar{u}$ at each height, $S_k = \sum_{l=0}^n (u_t(l) - u_k(l))^2$. l stands for the single height levels (between 10 and 70hPa), u_t for the wind at time t and u_k for the wind at time k with $|t - k| \geq 6$ months. For each point in time t , S_k is calculated and the profile with the smallest S_k value is classified as best match.

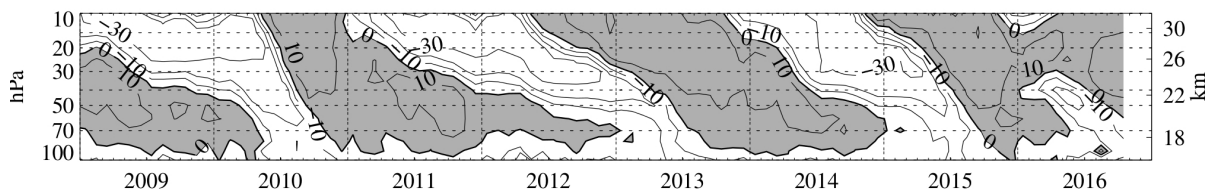


Figure VIII.1.: Same as Figure II.2, lowest row, extended to October 2016.

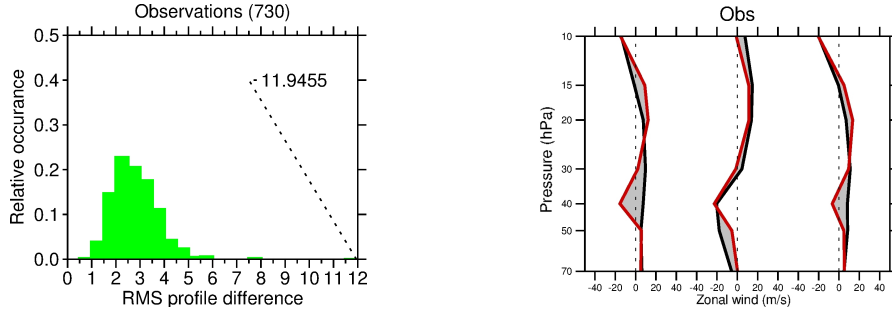


Figure VIII.2.: Left: Distribution of $\sqrt{S_{min}}$ in the FU Berlin observational record, January 1956 - October 2016 (730 months). The highest value of $\sqrt{S_{min}}$ is indicated. Right: 3 profiles corresponding to the highest $\sqrt{S_{min}}$ values (red) and their closest matches (black).

The distribution of the square root of the corresponding sum, $\sqrt{S_{min}}$, is shown in Figure VIII.2. The vast majority (98.1%) of values lies between 1 and 5. However, there are 3 clear outliers, 2 between 7.5 and 8 and one even at 11.9. All three correspond to the unusual QBO cycle at the beginning of 2016. The easterlies developing at 40hPa between 2 layers of westerlies lead to 3 zero crossings between 70 and 10 hPa. This is the only occasion that this happens in the FU Berlin record. The profiles correspond to February $\sqrt{S_{min}} = 11.9$, March $\sqrt{S_{min}} = 8.0$ and January $\sqrt{S_{min}} = 7.9$ of 2016. Interestingly, the last two months of 2015 were not that unusual, with $\sqrt{S_{min}}$ values of 3.8 (November) and 4.8 (December). The vertical structure of the equatorial stratosphere in November/December of 2013 also was very similar to that of 2015 (Figure VIII.1), however the 2015/16 cycle developed rather differently.

Using the MERRA-2 data, Newman *et al.* [2016] find that not only the zonal mean zonal wind, but also the temperature structure of the atmosphere in early 2016 was unusual compared to the 1980-2015 mean. Figure VIII.3 shows the equatorial (5°S-5°N) zonal mean wind $\bar{u}$ and temperature anomalies $\Delta\bar{T}$ in the ERA-Interim reanalysis. The temperature anomalies are calculated with respect to the monthly means 1979-2009. The easterly anomaly in the westerly jet develops slightly later than in the observations and further upwards, though the latter is probably due to the 40hPa level not being included in this dataset. As Newman *et al.* [2016] point out for the MERRA-2 data, the troposphere in ERA-Interim shows a large positive temperature anomaly compared to the mean state 1979-2009, which is most likely due to the strong ENSO event at the end of 2015 (which was the strongest one on record with weekly mean SST anomalies exceeding 3° C in the Niño 3.4 region¹).

Taking a latitudinal cross section of the zonal mean wind field at 30hPa and 50hPa (Figure

¹see for example the NOAA Climate Prediction Centre El Niño/Southern Oscillation (ENSO) Diagnostic Discussion (http://www.cpc.ncep.noaa.gov/products/analysis_monitoring/enso_disc_jan2016/ensodisc.pdf)

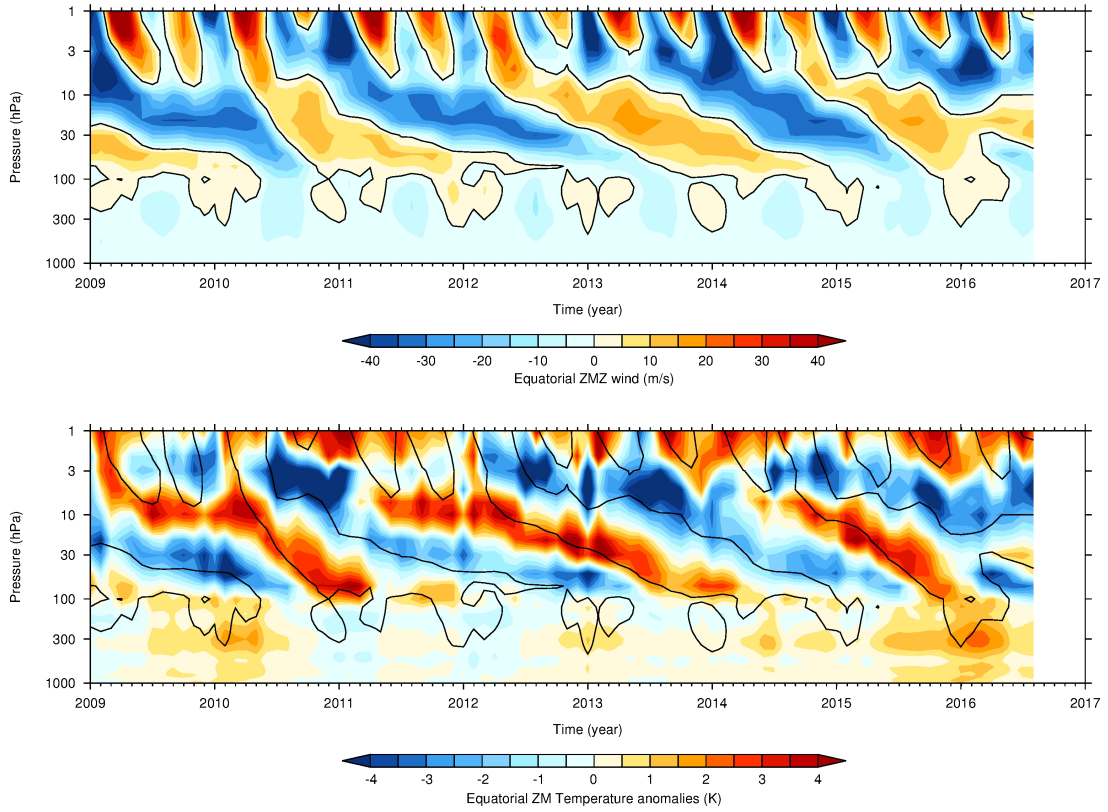


Figure VIII.3.: Equatorial zonal mean zonal wind (upper) and temperature anomalies (lower) from ERA-Interim. The T anomalies are calculated with respect to the monthly means 1979-2009. The zero wind line is shown on both panels.

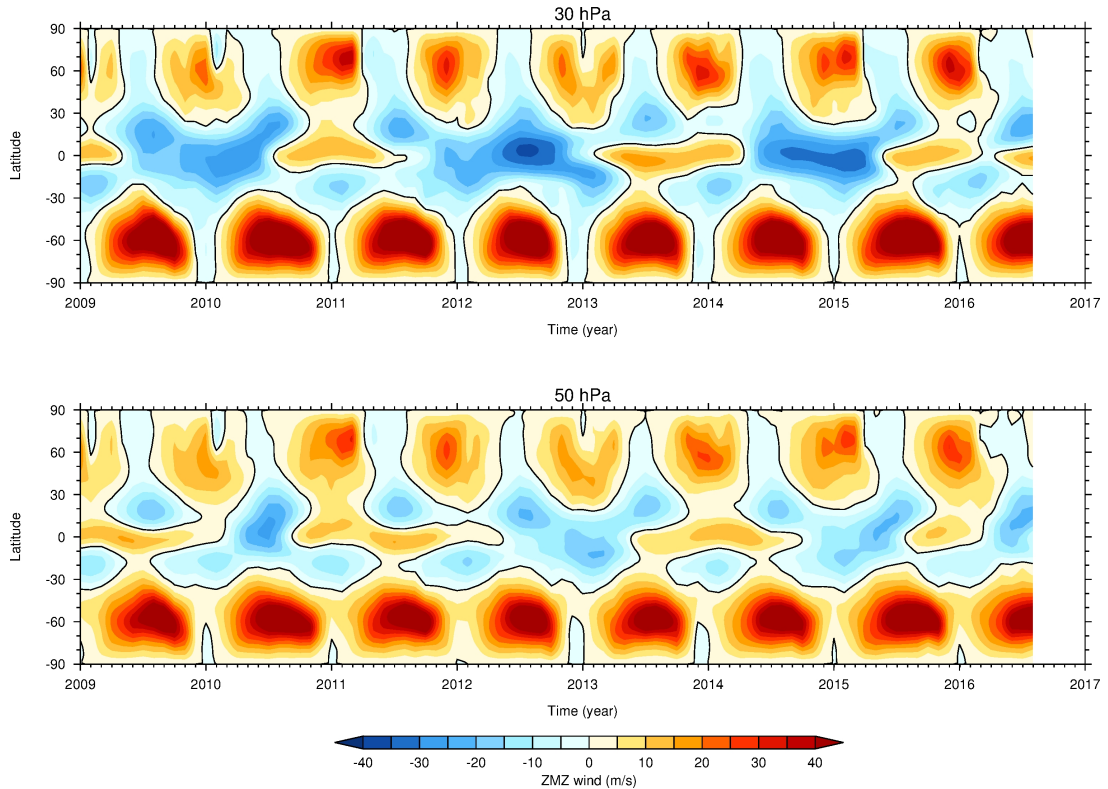


Figure VIII.4.: Zonal mean wind latitude-timeseries at 30hPa (upper) and 50hPa (lower) from ERA-Interim.

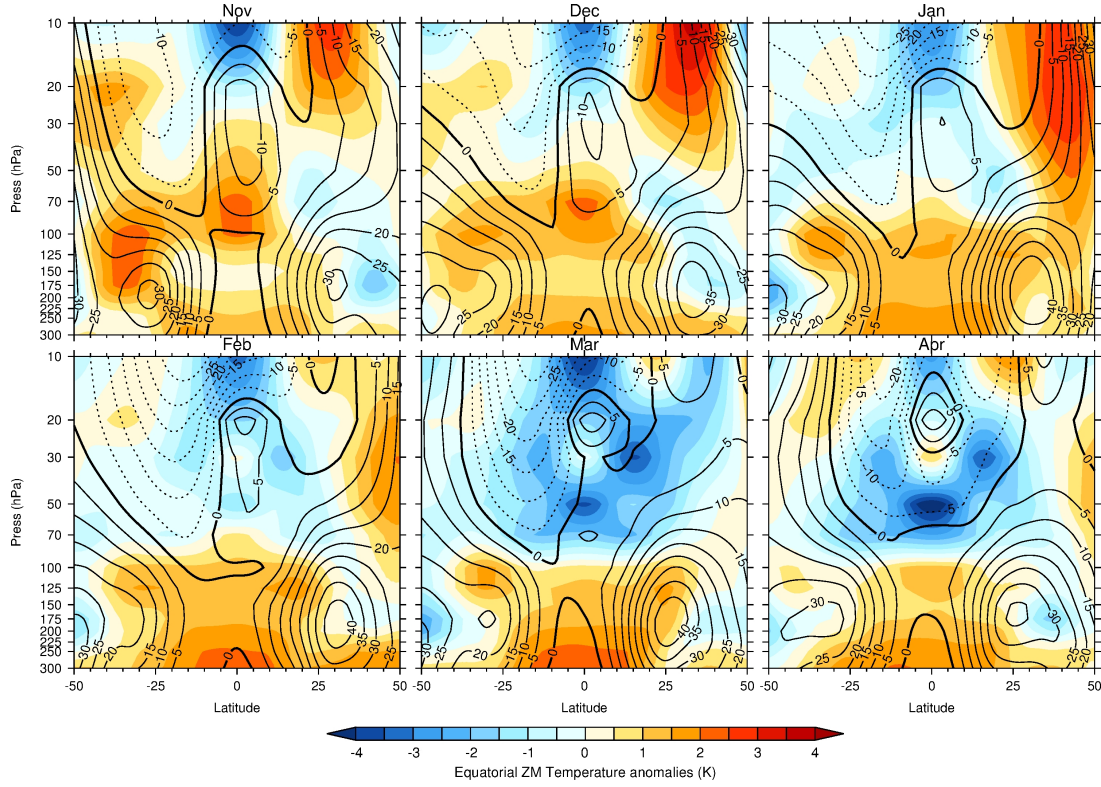


Figure VIII.5.: Latitude-Height cross section of ERA Interim November 2015 - April 2016 $\bar{u}$ (lines spaced 5 m/s) and temperature anomaly ΔT (colours) data.

VIII.4), the 2016 anomaly is clearly visible, especially the short westerly phase at 50hPa compared to previous years, as well as the short intermittent easterly phase at 30hPa, before the upwardly displaced westerly jet comes down again (compare Figure VIII.3). Newman *et al.* [2016] explain that the extension of the subtropical NH westerlies ($\sim 30^\circ$) at 40hPa to the equator starting mid 2015 could be contributing to a steady and anomalous QBO forcing. However, this connection between the subtropical and equatorial winds at 30hPa and 50hPa is not that uncommon, as can be seen in Figure VIII.4. The development at 50hPa is comparable to the year 2013/14, which developed as a normal QBO cycle. Still, the absence of a zero wind line between the NH subtropics and the equatorial region could facilitate equatorward Rossby wave propagation and momentum transport.

Figure VIII.5 shows the development of the zonal wind and the temperature anomalies from November 2015 to April 2016. While the equatorial stratospheric temperature anomaly is consistent with the zonal wind structure of the descending westerly according to the thermal wind relation, a positive anomaly at about 30hPa develops in January. As the easterly shear zone forms in the lower stratosphere, this temperature anomaly gets stronger; additionally, a negative anomaly forms at 50-70hPa due to the positive wind shear at the easterly-westerly zone bound-

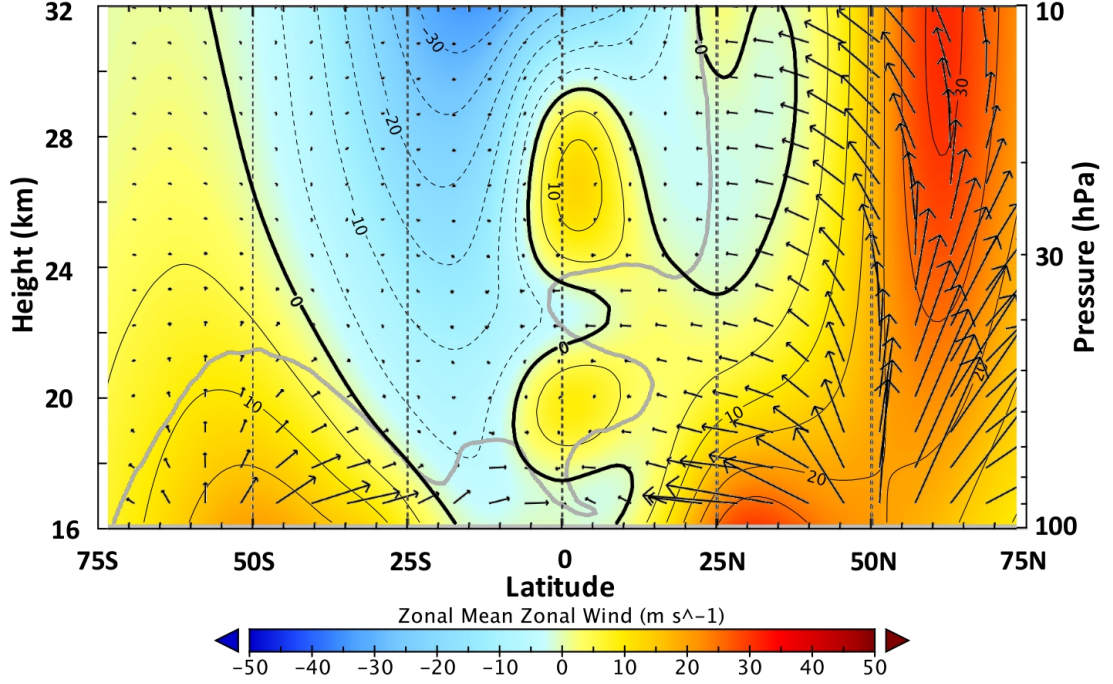


Figure VIII.6.: Latitude-Height cross section showing zonal mean zonal wind (colours and black contours, negative values dashed), Eliassen-Palm fluxes (arrows) and their extent (grey contour). Plot from Osprey *et al.* [2016] (their figure 2A).

ary. The strong positive temperature anomaly in the troposphere persists throughout these six months. The absence of a zero wind line between subtropics and the equatorial region from December to March is quite evident as well. These findings are in accordance with the analysis of the Merra-2 dataset by Newman *et al.* [2016]. Osprey *et al.* [2016] look more closely at the forcing of the 2015/16 anomaly. Figure VIII.6 shows zonal mean zonal wind and Eliassen-Palm (EP) fluxes in February 2016 from the ERA-Interim operational reanalysis. The EP flux is indicative of momentum of waves; in this case, the origin of the wind forcing comes from the Northern Hemisphere subtropics. Osprey *et al.* [2016] further break down the zonal acceleration $\frac{\partial \bar{u}}{\partial t}$ into wave flux and advection terms, which clearly shows the contribution of the waves to the easterly forcing at 40hPa.

VIII.2. A model analogue

Since the 2015/16 QBO disruption was not predicted, it is particularly important to understand the mechanism behind the event and its implications for forecasting of the QBO, and its impacts. If the anomalously warm tropical troposphere contributed to the event, then such irregular QBO cycles might become more common under climate change. Studying models and analysing

observations can provide an opportunity to learn about the forcing and development of possible future events.

The closest-profile analysis used for the observational record (Figure VIII.2) is applied to the climate models that exhibit a QBO in the tropical stratosphere. Figure VIII.7 shows the $\sqrt{S_{min}}$ statistics for the models. Similar to the observations, most values are low, with 89.8% lying between 0.5 and 2. No obvious model outliers can be identified, with the maximum $\sqrt{S_{min}}$ value being 5.6 (HadGEM2-CC) and thus much less than in the observations. However, some of the model runs are much longer than the ~ 60 year radiosonde observations, which means that there will be more profiles in the models to compare to and thus the “extreme” ones may not stand out as much. One example for this is CMCC-CMS, which shows a rather bizarre QBO (Figure IV.1), but does not have a maximum $\sqrt{S_{min}}$ value comparable to the observations.

The wind profiles corresponding to the three highest $\sqrt{S_{min}}$ values for each model are shown in Figure VIII.8. The only two models whose “extreme” profiles exhibit 3 phase changes between 10 and 70hPa are MPI-ESM-MR and CMCC-CMS. For its obvious shortcomings in QBO modelling, CMCC-CMS is disregarded in this study, though an analysis of its forcing might still be useful as a reference. MPI-ESM-MR, however, shows only one event in its 156 years of data, which makes it a valuable model to compare the 2015/16 QBO disruption to.

Comparing the $\bar{u}$ profile outliers of MPI-ESM-MR and the Observations, the unique west-east-west vertical structure can be identified in both. However, the shear in the MPI model is less compared to the observations and it lacks two pressure levels (15hPa and 40hPa) compared to the observations, which might contribute to this bias. Interpolation onto the observation’s pressure levels gives a shear of $-10.6 \frac{m/s}{km}$ between 40 and 30hPa for the model, compared to $-21.2 \frac{m/s}{km}$ between 50hPa and 40hPa in the observations. Furthermore, the easterly sandwiched between the two westerly jets occurs higher in the atmosphere (20-30hPa), consistent with the whole QBO structure in the model being shifted upwards (Table IV.1).

Figure VIII.9 shows the unusual QBO cycle in MPI-ESM-MR, which appears during the winter 1964/65 in the historical **r1i1p** run (the other historical runs do not show this behaviour). One apparent difference between the model and observations (and the ERA-Interim data) is that the westerly phase at 70hPa from the previous QBO cycle is not terminated by the descending easterlies. This will lead to some filtering of upward propagating gravity waves. Still, the westerly shear zone in the upper stratosphere exhibits a normal amplitude of about 25-30 m/s throughout 1964. As already noted in Chapter IV, the westerly shear zone descends slowly

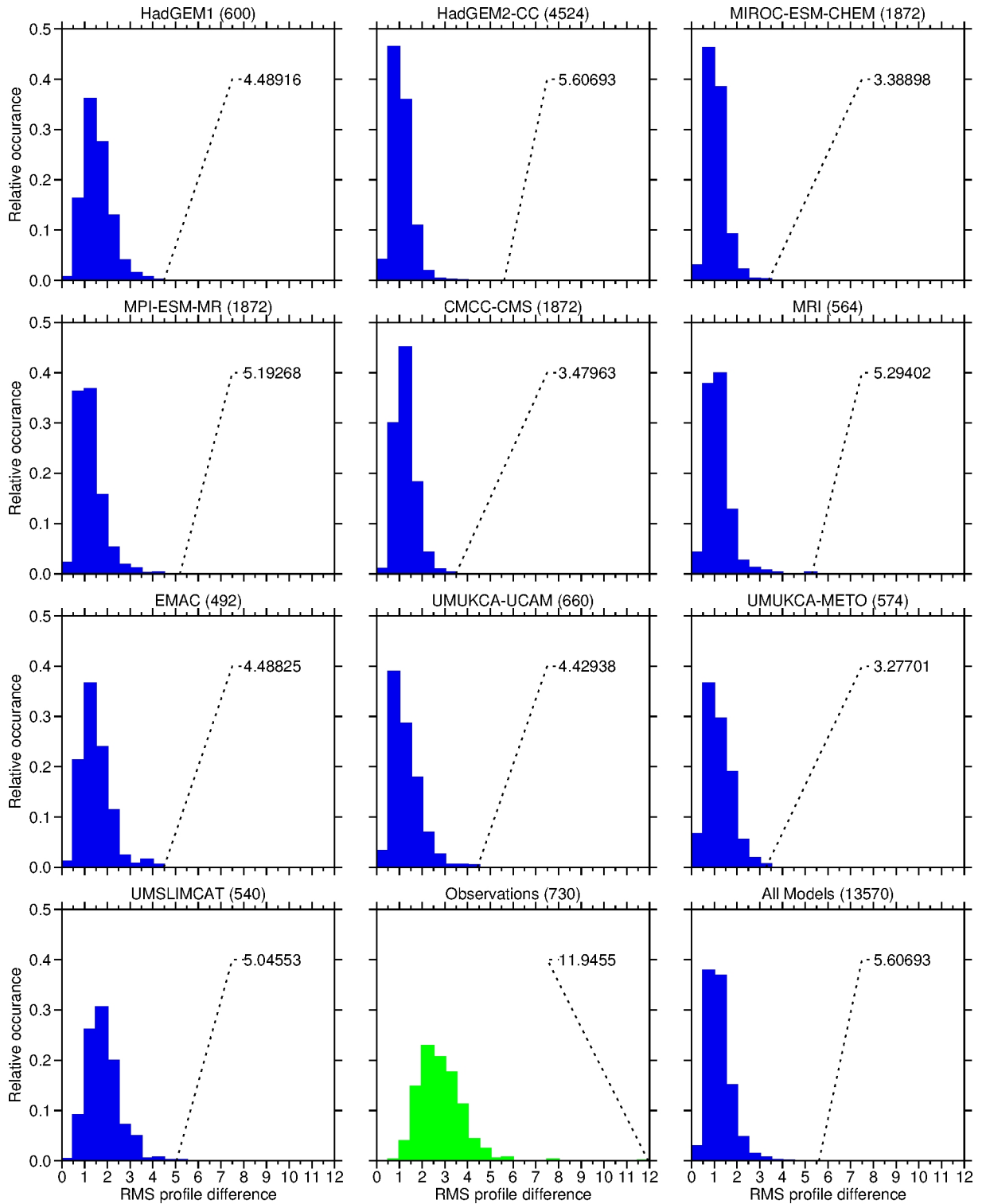


Figure VIII.7.: Same as Figure VIII.2, for all models. The numbers in brackets give the number of months used in the calculation. The last panel shows the statistics for all models. The statistics from the observations (Figure VIII.2) is shown again as for comparison.

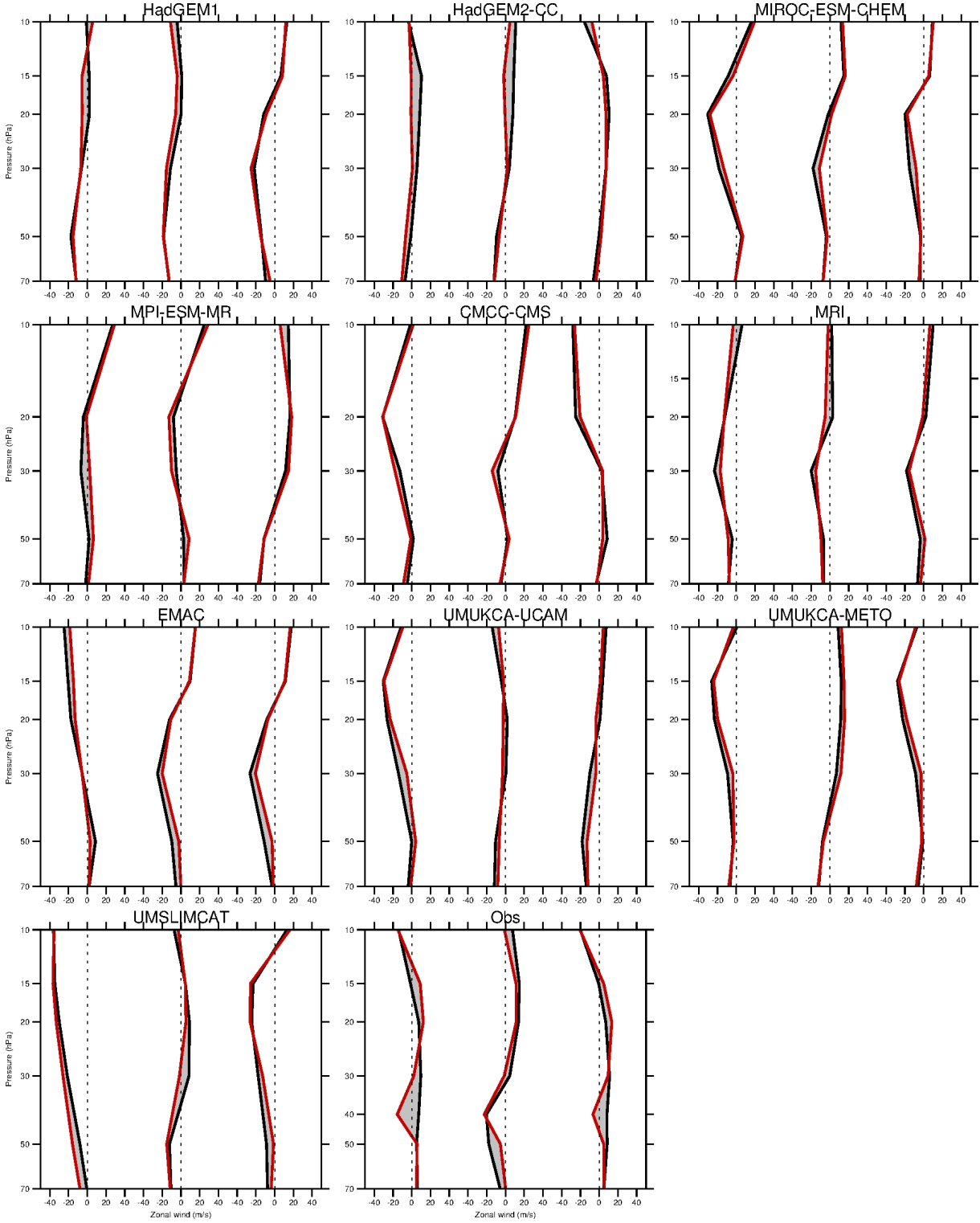


Figure VIII.8.: Three worst matched zonal wind profiles (highest $\sqrt{S_{min}}$ values) for each model and the FU Berlin observations (as Figure VIII.2). The red profile shows the outlier, the black its best match and the grey shading is an indicator for $\sqrt{S_{min}}$. The shown pressure levels are the levels used for the calculation.

compared to observations. The upward propagation of the westerly shear zone that occurs simultaneously with the development of the 40hPa easterlies in the observations is not as evident in the MPI-ESM-MR model.

As in the ERA-Interim data (Figure VIII.3), there is a large positive temperature anomaly in the upper troposphere before the disruption of the westerly occurs. In contrast to the observations, the model's anomaly is located mainly above the tropopause, whereas the troposphere exhibits a weak negative anomaly. Checking the sea surface temperatures in the model (Figure VIII.10) shows that there the ENSO index was positive in 1964, though its magnitude was much less than the observed 2015 El Niño event. This would explain the discrepancy in the tropospheric anomalies preceding the QBO disruption.

Figure VIII.11 shows two latitude time series at 20hPa and 30hPa, similar to VIII.4. These heights are chosen as the model anomaly develops higher than in the observations. Compared to ERA-Interim, the polar jets do not extent quite as far into the subtropics. However, during the disruption in the 1964/65 winter, there is a westerly bridge from the equator to the Northern Hemisphere; the peak amplitude in this bridge region is slightly less than in ERA-Interim.

The evolution of the anomalous event in a latitude-height cross section is shown in Figure VIII.12, from November 1964 until April 1965. The chosen sector of the atmosphere is the same as in Figure VIII.6. Similar to the reanalysis, there is no subtropical zero wind line in the Northern hemisphere, so that Rossby waves can travel to the equator. The Eliassen-Palm (EP) flux (F_ϕ, F_z) is calculated following Andrews *et al.* [1987b] from daily means of zonal wind u , meridional wind v , vertical wind w and potential temperature $\theta = T \left(\frac{p}{p_0} \right)^\kappa$ (p indicating pressure, $p_0 = 1000\text{hPa}$ and κ the isentropic factor for air) as

$$F_\phi = \rho_0 a \cos \phi \left(\bar{u}_z \frac{\overline{v'\theta'}}{\theta_z} - \overline{w'v'} \right)$$

$$F_z = \rho_0 a \cos \phi \left(\left(\beta - \frac{(\bar{u} \cos \phi)_\phi}{a \cos \phi} \right) \frac{\overline{v'\theta'}}{\theta_z} - \overline{w'u'} \right)^2$$

Primed values x' indicate zonal anomalies of the field x , $x' = x - \bar{x}$, $\bar{x}$ being the zonal mean of x . Subscripts indicate a partial derivative by the respective variable. a is the Earth's radius and $\beta = 2\Omega \sin \phi$ the Coriolis parameter. ϕ stands for the latitude, ρ_0 for uniform density via the ideal gas law $\rho_0 = \frac{p}{R_{gas} T_0}$, with gas constant R_{gas} and $T_0 = 288K$.

Especially in the early winter months November-December the equatorward wave fluxes in the middle stratosphere (20-30hPa) are evident. The momentum from these waves is very likely the

²This is identical to the calculations of EP Flux in the ERA-Interim reanalysis dataset (http://meteo.mcgill.ca/srip/S-RIP_ZMD_description.pdf); however, the flux data is available for ERA-Interim, but had to be calculated for MPI-ESM-MR.

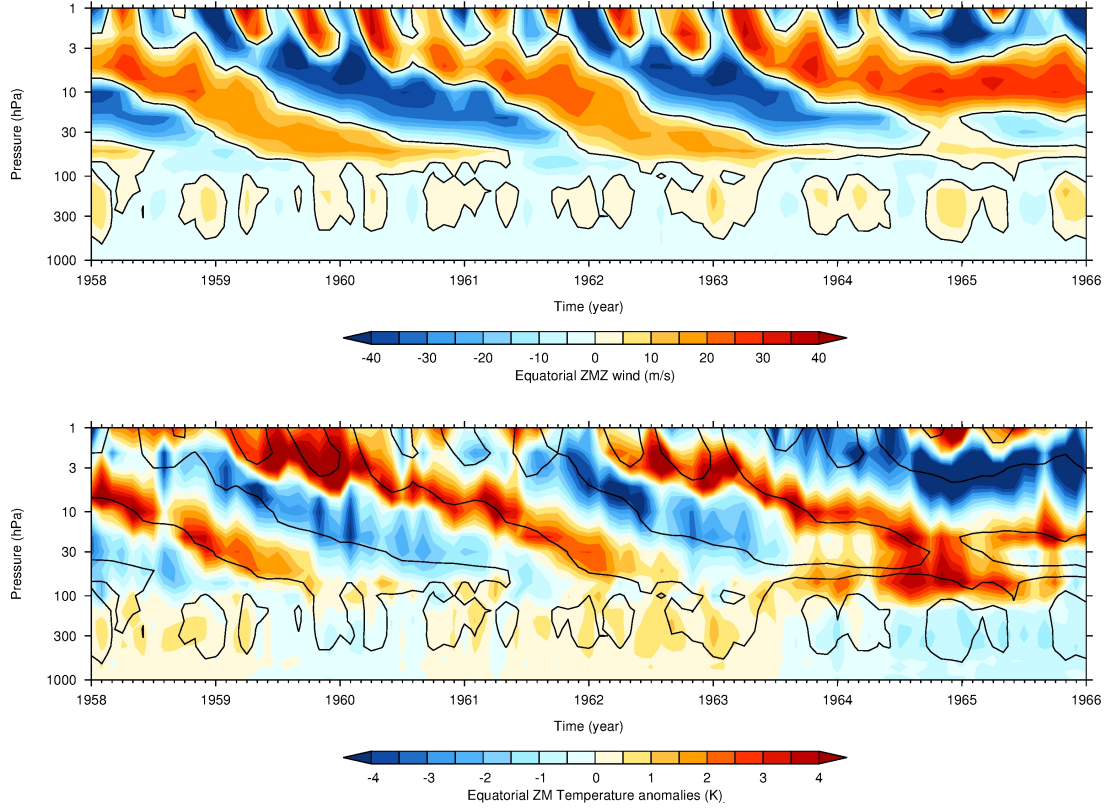


Figure VIII.9.: Equatorial $\bar{u}$ (upper) and $\Delta\bar{T}$ (lower) from the MPI-ESM-MR historical run. The T anomalies are calculated with respect to the monthly means of the entire 156 year run, excluding the years 1964 and 1965. The zero wind line is shown on both panels.

factor contributing to the formation of the easterly jet within the westerly shear zone. Thus, the mechanism behind the forcing of the disruption in the MPI-ESM-MR model shows similarities to the ones found by Osprey *et al.* [2016] for the observations.

The question remains whether the observed 2015/16 QBO event was singular or whether one might see more disruptions to the QBO which has been fairly regular for the past 60 years. With rising global temperatures, positive tropospheric temperature anomalies, which have preceded the 2015/16 event, will be more common. While Latif & Keenlyside [2009] do not find a clear trend in future ENSO behaviour in CMIP5 models, Cai *et al.* [2014], who evaluate a subset of CMIP3 and CMIP5 models, claim that extremes such as the 1997/98 and 1982/83 events will appear more frequently. The state of ENSO is important to predict the QBO, as the strong 2015 El Niño phase is very likely to have played a role in its disruption [Dunkerton, 2016]. Further implications of climate change involve shifts of the Hadley Circulation and the subtropical jets' strength and position, which then alters the critical wind line for wave propagation. As the 2015/16 QBO disruption was forced by waves emerging outside the tropics it is important to understand how atmospheric wave creation and propagation will change in the future to be able

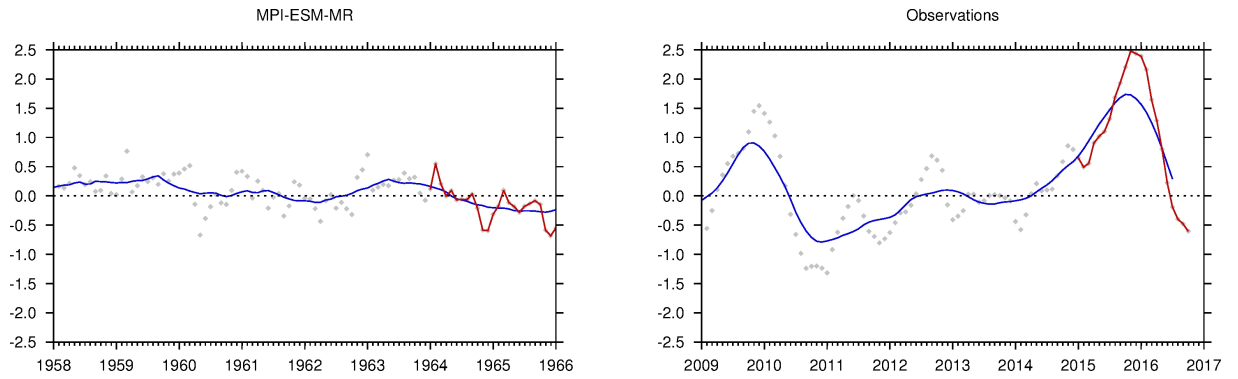


Figure VIII.10.: Niño 3.4 index calculated from the MPI-ESM-MR sea surface temperature (left) and the NOAA observational index (right). Grey dots indicate monthly means, the blue line the 3 month running mean and the red line the monthly anomalies in the year preceeding the QBO disruption and the following 12 months (model) or 10 months (Observations).

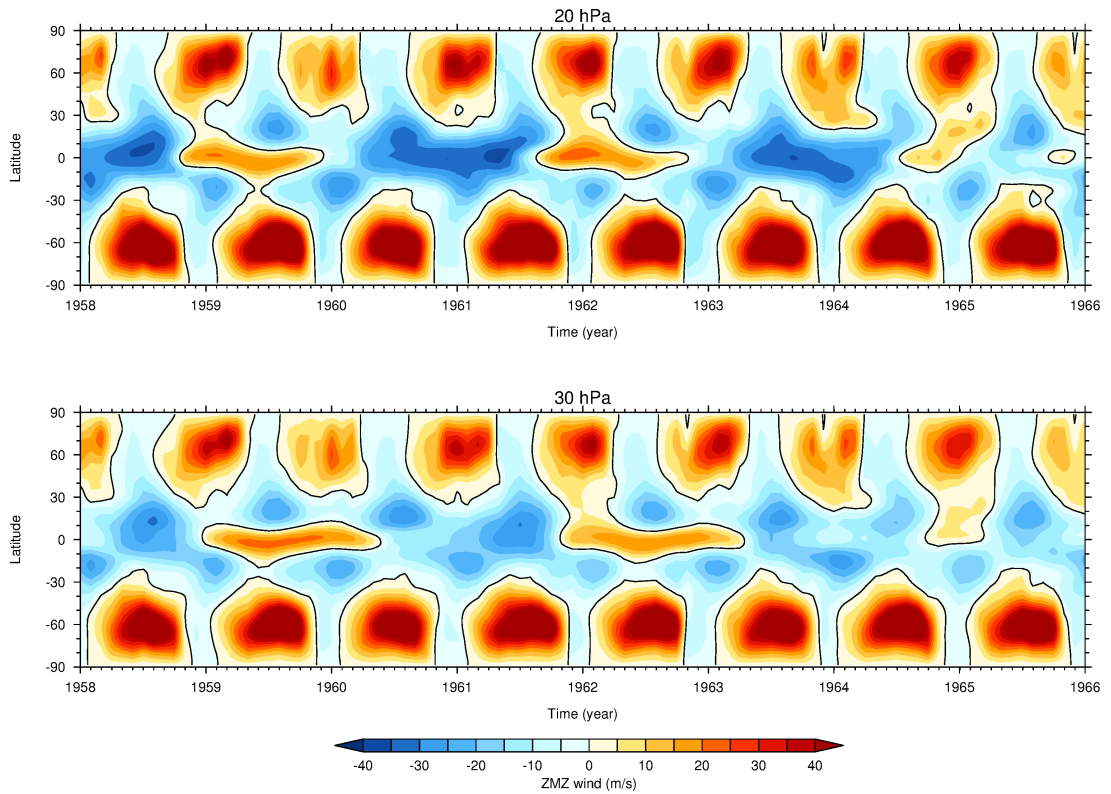


Figure VIII.11.: Zonal mean wind latitude-timeseries at 30hPa (upper) and 50hPa (lower) in MPI-ESM-MR.

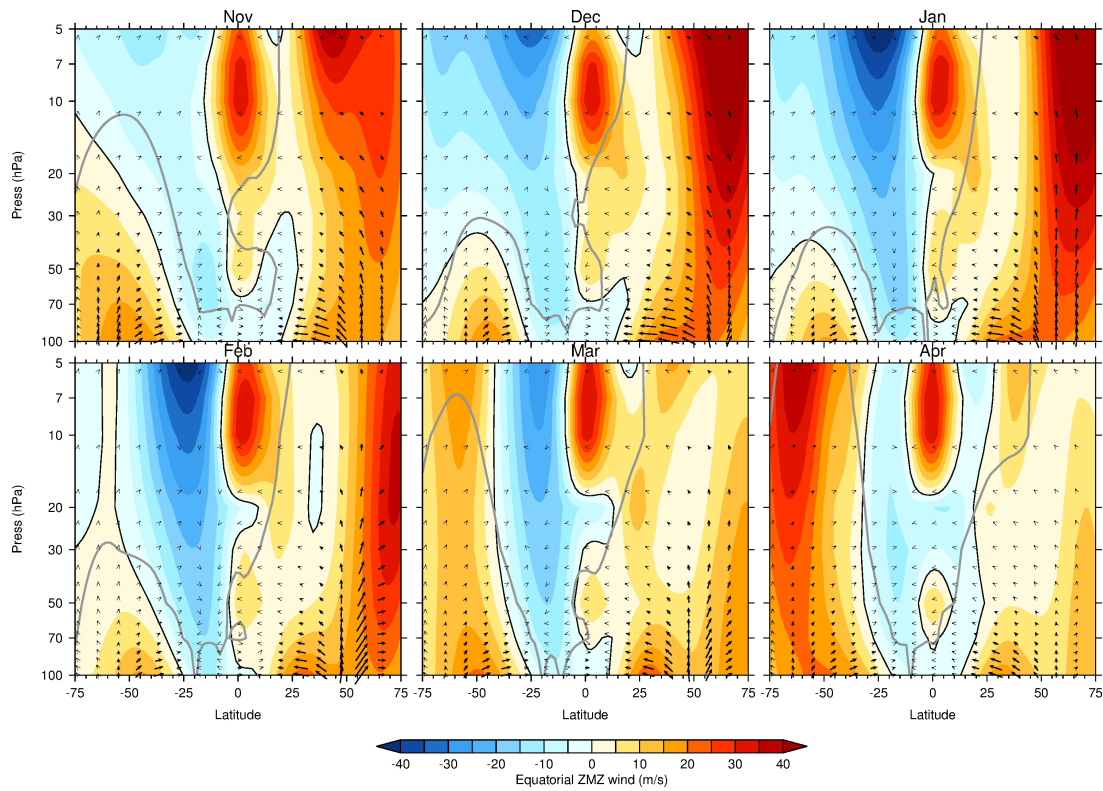


Figure VIII.12.: Latitude-Height cross section of MPI-ESM-MR $\bar{u}$ (colours) and the Eliassen-Palm flux (vectors) November 1964 - April 1965. The black contour shows the zero wind line, the grey line the extent of the EP flux. EP flux vectors are scaled in the horizontal by $\frac{1}{111.1}$ to account for the aspect ratio of the panel.

to forecast potential QBO irregularities.

Despite some shortcomings of the model compared to the observed event in 2015/16, such as the observed 1964/65 QBO cycle actually being normal, the ability of the MPI-ESM-MR model to produce an event that is similar to this unprecedented QBO disruption make it a useful reference tool for future QBO projections. Figure VIII.13 shows examples of 100 years of future runs of the MPI-ESM-MR model with different emission scenarios (Representative Concentration Pathways (RCPs) [Taylor *et al.*, 2012]). The frequency of unusual events where there is a prolonged westerly phase in the upper stratosphere over an easterly phase evolving in the mid-stratosphere seems to be increasing, from 1 in 458 years (3 historical runs) to 2 in 100 (RCP 2.6), 3 in 100 (RCP 4.5) and 4 in 100 (RCP 8.5). The other two CMIP5 models that can simulate a QBO oscillation (HadGEM2-CC and MIROC-ESM-CHEM) do not show any similar behaviour in their future climate projections (Figure B.10).

VIII.3. Chapter summary

The winter 2015/16 has seen an unprecedented disruption of the QBO pattern, which was not predicted by any forecast. Analysis of the MERRA-2 reanalysis [Newman *et al.*, 2016] showed an unusually warm upper troposphere due to the strong ENSO event in 2015, as well as a westerly bridge from the equator to the Northern Hemisphere. The ERA-Interim reanalysis shows similar anomalies in the wind and temperature structure. Closer analysis of wave fluxes (Figure VIII.6) presented by Osprey *et al.* [2016] shows that the disruption of the descending westerly shear zone by easterlies at about 40hPa was due to extratropical waves emerging at about 30°N travelling towards the equator.

An analogous event was identified in the MPI-ESM-MR historical run. The development of the easterly anomaly within the westerly QBO phase at 30hPa is similar in the model, making it a useful tool to study the exact dynamics behind the unusual behaviour. Future projections of the MPI-ESM-MR model predict an increase in the number of QBO disruptions (Figure VIII.13). Understanding the mechanics behind the 2015/16 event and in the model can improve the accuracy of predicting the future QBO development.

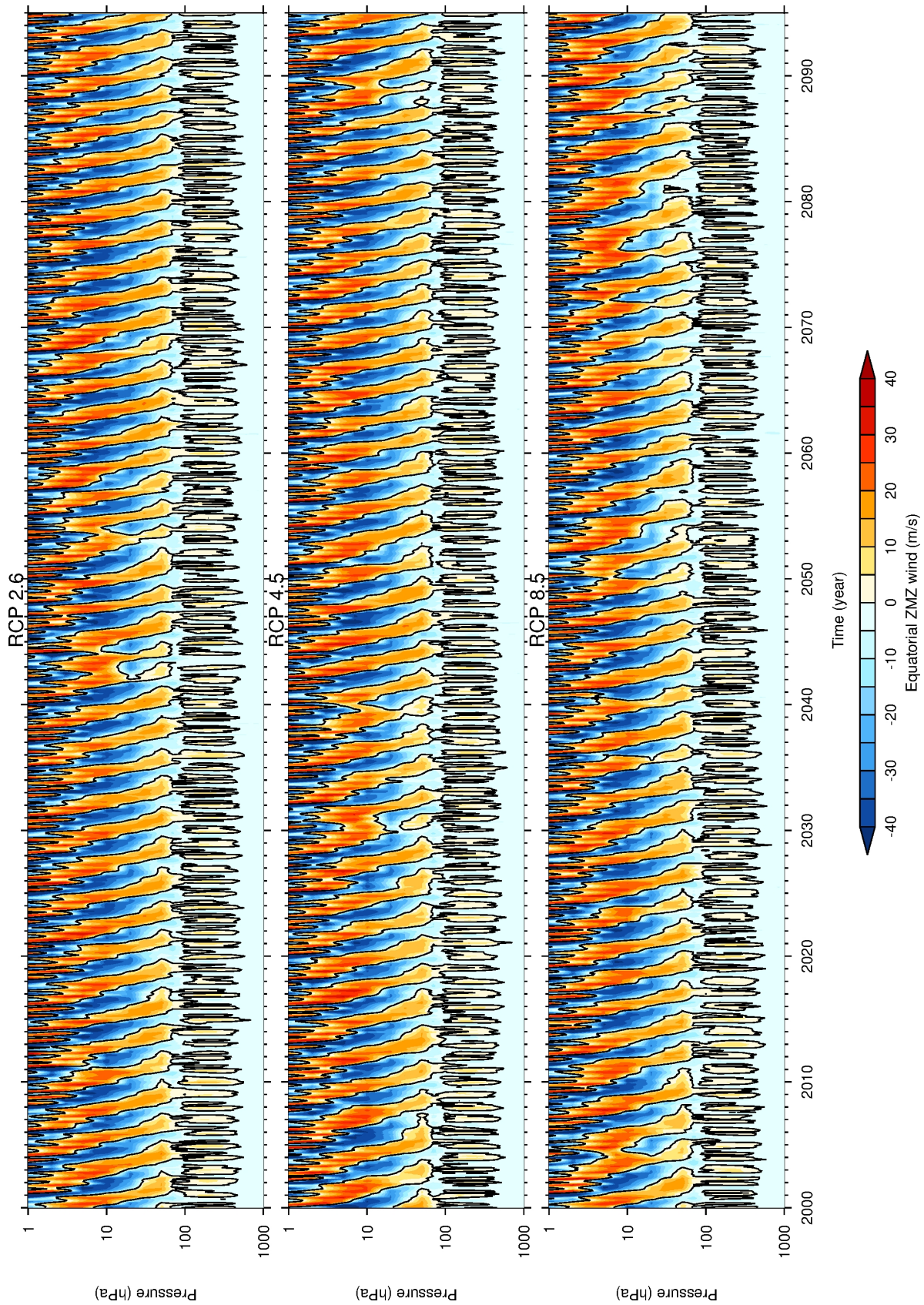


Figure VIII.13.: 100 years of equatorial zonal mean zonal wind in three different scenarios from the MPI-ESM-MR model. The zero wind line is shown in black.

IX. Summary and outlook

IX.1. Results of this study

The Quasi-Biennial Oscillation is the dominant mode of variability in the equatorial stratosphere. Therefore, and due to possible teleconnections to other parts of the climate system, it is very important that global climate models can adequately represent it.

Out of the CMIP5 and CCMVal-2 model intercomparison projects, only 10 models exhibit a QBO-like structure in equatorial zonal mean zonal wind and temperature. A set of characteristic metrics, described in section IV.1, is developed to assess the quality of these simulations. For reference, the FU Berlin radiosonde dataset and three reanalyses are used. In the multi-model mean, the period of the oscillation is well represented, as is the difference in maximum wind amplitude for the two shear zones. The definition of the metrics as well as the model assessment have been written up as Schenzinger *et al.* [2016].

Common biases of models, described in section IV.3, are the upward shift in the QBO structure, which implies that the peak level of the QBO in models is too high and that the lowermost affected level lies well above the tropopause. Furthermore, the latitudinal extent in models is slightly too narrow, a discrepancy that is stronger at lower stratospheric levels. This structural difference between the model and reanalysis QBO can be seen in the wind and temperature field (Figure IV.8) and is, in the case of the temperature bias, likely to influence the QBO-induced meridional circulation (Figure II.7). Another discrepancy between models and observations is the representation of the shear zone descent, especially the slow descent of the westerly shear zone (Figure IV.9).

A further development of the metrics uses empirical orthogonal function analysis of equatorial $\bar{u}$ between 10 and 70hPa to quantify QBO behaviour (section IV.5). The first two EOFs account for 71% of the variance in the FU Berlin radiosonde dataset. They can be used for an alternative definition of the QBO period to overcome the limitation of having to choose one particular pressure level for the calculation. Furthermore, the polar angle ψ between the first two EOFs

can be used to classify the state of the QBO in the stratosphere (Figure IV.13, Tables IV.4 and V.1) and its time derivative $\dot{\psi}$ as an alternative measure of the QBO descent rate (Figure IV.11).

While the wind in the equatorial stratosphere represents a challenge to model developers, the influences of the QBO on other variables, like ozone, the extratropics and even the surface, are important and have been investigated since the discovery of the QBO (e.g. Landsberg [1962]; Angell & Korshover [1964]). One of the most well-known teleconnections is between the QBO and the strength of the wintertime NH stratospheric polar vortex, as discussed by Holton & Tan [1980]. As the polar vortex influences the state of the North Atlantic Oscillation [Baldwin & Dunkerton, 2001], it is an important factor in determining the weather in the Northern Hemisphere.

The strength of the QBO teleconnection to the surface in the early winter months (December-January) is assessed using the observational mean sea level pressure dataset HadSLP. Classic composite analysis using the state of the zonal mean zonal wind at 50hPa shows an NAO pattern in January, while in December and February the Siberian and Aleutian low are most prominent (Figure V.3). As there are a lot of different profiles of equatorial $\bar{u}$ in these states, a further distinction into 4 QBO states, defined by their EOF polar angle ψ , has been made. The January signal intensifies when the equatorial stratosphere between 10 and 70hPa is dominated by wind of the same sign (Figure V.6).

To eliminate influences of other factors such as the 11-year solar cycle or ENSO in the composite analysis, multilinear regression analysis of mean sea level pressure is carried out. Differences in the response to QBO forcing dependent on the choice of the pressure level as an index of the regression (Figure V.8) suggest, that rather than wind at one pressure level, the state of the whole equatorial stratosphere might be relevant for the surface impacts. Employing an index reconstructed from the first two EOFs of equatorial zonal mean zonal wind in the analysis ($u^* = R \sin(\psi + \varphi)$), following Baldwin & Dunkerton [1998], shows a clearer NAO response, which maximises for $10^\circ \leq \varphi \leq 30^\circ$. This corresponds to a QBO state with one dominant shear zone between 10 and 70 hPa (Figure IV.18), which is consistent with the composite analysis results.

An equivalent analysis for the QBO-surface connection is carried out for the 505 year HadGEM2-CC control run, which only simulates internal climate system variability. The patterns found in the composite and regression analysis are similar to the ones from observations, though much weaker in amplitude. One of the reasons for the reduced amplitude in the control run might be

the different climatological state of the mean sea level pressure (compare Figures V.4 and V.13), another one the model biases regarding the QBO representation (Table V.2).

Having established a connection between the equatorial stratosphere and the Northern Hemisphere surface, the question of the interaction route is addressed with regression analysis of the zonal mean circulation and wave fluxes. The regression analysis suggests that the QBO mainly modulates the residual circulation and wave fluxes in the stratosphere (Figure VI.3); influences reaching down to the surface in the zonal mean analysis can be seen in geopotential height (Figure VI.4). Compared to other factors, like the 11-year solar cycle and ENSO, the QBO has the strongest influence on the polar stratosphere. This extratropical response to the equatorial zonal wind can be seen in the Northern Hemisphere circulation (Figure VI.5), where it shows in the stratospheric polar vortex, as well as in the tropospheric cyclonic pattern over the North Atlantic, consistent with the QBO-NAO connection found in the surface regression analysis.

The regression analysis is applied to the MJO region as well, as recent publications suggest a tropical stratosphere-troposphere interaction [Liu *et al.*, 2014; Yoo & Son, 2016; Moss *et al.*, 2016]. A circulation response to the QBO is found in the regression analysis, though it is located south of the MJO center.

One of the main problems with composite and regression analysis is that the direction of the influence cannot easily be determined. A novel method, developed by Runge *et al.* [2012a,b, 2014, 2015], is applied to address the various possible QBO teleconnections. In this causal effect network analysis, the “polar route”, where the QBO influences the upper stratospheric polar vortex, which subsequently modulates the lower stratospheric polar vortex, and then the surface NAO, has been found to be the main interaction route in early Northern Hemisphere winter (December-January). An additional influence on the surface NAO from the tropics via the subtropical jet occurs in the network analysis. However, it is attributed to ENSO modulating the subtropical jet, which subsequently modulates the QBO, instead of the other way round (Figure VII.12). The QBO-Polar Vortex interaction is found to be modulated by the Pacific Decadal Oscillation, a factor that is not taken into account in the composite or regression studies. Applying the CEN analysis method to timeseries of monthly means (as opposed to higher time resolution data) has been found most useful in the context of QBO teleconnections.

A QBO-MJO interaction in DJF could be established with the CEN analysis, which is robust when sampling monthly, fortnightly or weekly (Figures VII.14, VII.21). An influence of the QBO on the central Indian Monsoon is detected in the monthly analysis, though it varies in sign

during the main Monsoon season (July and August). At higher time resolution, this connection is ascribed to either ENSO or the MJO, with ENSO being the more important factor.

Despite being called an oscillation, which is commonly associated with harmonic motion, QBO cycles exhibit some variation (e.g. in amplitude and period). Nonetheless, the appearance of easterly and westerly shear zones in the upper stratosphere and their downward propagation has been fairly regular for the past 63 years of radiosonde observations (Figure II.2). The winter 2015/16, however, saw a very unusual QBO cycle that has not been forecasted. The development of the disrupted QBO cycle in the ERA-Interim reanalysis is presented in section VIII.1, Figures VIII.3-VIII.5.

In one of the CMIP5 models historical runs, a similarly disrupted QBO cycle is found (winter 1964/65 in the MPI-ESM-MR model, `r111p` historical run). Its development is described in section VIII.2 and similarities to the observed disruption are discussed. Both events exhibit a stratospheric temperature anomaly near the tropopause. Other common features include a stratospheric bridge to the Polar westerly jet and the absence of a zero wind line in the Northern Hemisphere that could facilitate wave propagation from the subtropics to the equator. Latitude-height cross sections of EP flux indicate wave fluxes into the middle stratosphere emerging from around 30°N (Figures VIII.6, VIII.12). Future projections of the MPI-ESM-MR model predict an increase of these disrupted QBO cycles.

IX.2. Discussion and future work

The assessment of CMIP5 and CCMVal-2 models had some promising results regarding model development, with certain QBO features being well represented. It also showed where models need to be improved, which is the height structure of the QBO and an adequate representation of the westerly shear zone development. The latter is problematic in reanalyses as well, as noted by Kawatani *et al.* [2016]. They explain the reanalysis bias as arising through inadequate representation of gravity waves in the forecast model. In the climate models analysed in this study, these waves are parametrised, and moreover, not tied to their sources. However, some newer climate models also incorporate spatial and temporal variance in their wave parametrisation (e.g. de la Cámara *et al.* [2016]). The assessment of the representation of the QBO in these more recent GCMs will be part of the continuing QBOi project. The set of QBO characteristics developed in this work will provide the framework for analysing QBO morphology

in the reanalysis intercomparison S-RIP and in model runs that are part of QBOi. New model developments with higher resolution, better representation of gravity waves and their sources, will be an important part in understanding what minimum requirements models need to fulfill to adequately simulate a QBO. The set of metrics could also be used in the WCRP-SPARC Chemistry-Climate Model Initiative (CCMI)¹ to address the potential impact of chemistry on stratospheric dynamics.

What makes the equatorial stratosphere so interesting is the potential to improve forecasts of other parts of the climate system, such as the Indian Monsoon, the MJO and Northern Hemisphere winter circulation (c.f. Claud & Terray [2006]; Inoue & Takahashi [2013]; Marshall *et al.* [2016]; Anstey & Shepherd [2014]; Scaife *et al.* [2014]). Given the long period of the QBO, knowing its impacts would be very useful for long range forecasting (e.g. Scaife *et al.* [2014]).

To assess the teleconnections between the QBO and the Northern Hemisphere in winter, it is found that single month analysis as presented here is important as signals might get smoothed out in the commonly used seasonal average (Figures V.1 and V.3). Composite analyses of mean sea level pressure by QBO state at the 50hPa level are consistent with previous studies [Anstey & Shepherd, 2014]. However, if the composite analysis is done based on certain wind profiles, a stronger surface response emerges in “pure” QBO states, where one shear zone dominates between 10 and 70hPa. The same four state composite analysis also shows a strong polar vortex response in January for a westerly dominating the equatorial stratosphere. The latter confirms the study of Anstey & Shepherd [2008], despite their composites being based on the QBO phase change a couple of months earlier. Therefore, it is important to take the vertical profile of the QBO into account when studying teleconnections, rather than only one level.

Following the idea of the composite analysis that the state of $\bar{u}$ at more than one equatorial level matters, a QBO index u^* is constructed from the first two EOFs of the equatorial zonal mean zonal wind field for regression analysis. Compared to regression analysis that uses wind at one level, the surface NAO signal is larger and more significant when using the EOF index u^* , confirming the notion of the wind profile instead of one level being relevant for the surface response.

The analysis is applied to the HadGEM2-CC control run data as well, which spans 505 years in contrast to the ~ 60 years of the observations. The principal results are confirmed, however the surface response is weaker and less significant. This could have to do with the decadal

¹<http://www.igacproject.org/CCMI>

variation of the Holton-Tan effect noted by Lu *et al.* [2008], or biases in the representation of the QBO and the interaction mechanisms in the model. In the observations, the QBO is found to modulate circulation and wave fluxes, mainly in the stratosphere; an insufficient resolution or inadequate representation of gravity waves in the model thus could interfere with the QBO-NAO teleconnection.

Despite various studies claiming a modulation of the MJO by the QBO that stress its usefulness in forecasting [Liu *et al.*, 2014; Yoo & Son, 2016; Marshall *et al.*, 2016], the standard regression analysis of the tropospheric circulation could not confirm these results. Differences between this study and Yoo & Son [2016] are the analysis of single winter months (compared to DJF average), using the reconstructed index u^* (compared to a 3 month smoothed timeseries of $\bar{u}$ at 50hPa) and applying a multivariate regression analysis (compared to a correlation and composite analysis); Marshall *et al.* [2016] analyse model hindcasts regarding QBO predictability and acknowledge an easterly bias in the simulated QBO. Given these differences in the basic analysis methods, no conclusive remarks about the QBO-MJO interactions can be made on this basis.

To shed more light on possible QBO teleconnections and further distinguish them from other influence factors, the novel approach of causal effect networks is applied. This method has been shown to be useful in the context of climate interactions [Runge *et al.*, 2014; Kretschmer *et al.*, 2016]. For the Northern Hemisphere winter teleconnection, the network analysis suggests that the QBO-NAO interaction takes place via a modulation of the upper stratospheric polar vortex, which subsequently travels down to the surface. The downward influence of the polar stratosphere on the surface is consistent with Baldwin & Dunkerton [2001] and the QBO-Vortex interaction confirms the presence of the Holton-Tan effect at least in the past 60 years. The possible route via modulation of the subtropical jet could not be confirmed within the network; rather, a “backward” influence of the subtropical jet on the QBO is detected. The exact mechanism of this interaction, possibly happening via variation of the wave fluxes towards the equator, is subject to further investigation.

Analysing the QBO-MJO connection within the CEN framework did show a modulation of the winter (DJF) MJO by the QBO, which is independent of ENSO, as claimed by Yoo & Son [2016] and Marshall *et al.* [2016]. Further possible implications for the central Indian Monsoon, however, could not be established, despite the QBO having been used for years to improve Monsoon prediction [Bhalme *et al.*, 1987; Claud & Terray, 2006; Inoue & Takahashi, 2013]. Additional indices in the analysis might be needed to conclusively test for the QBO-

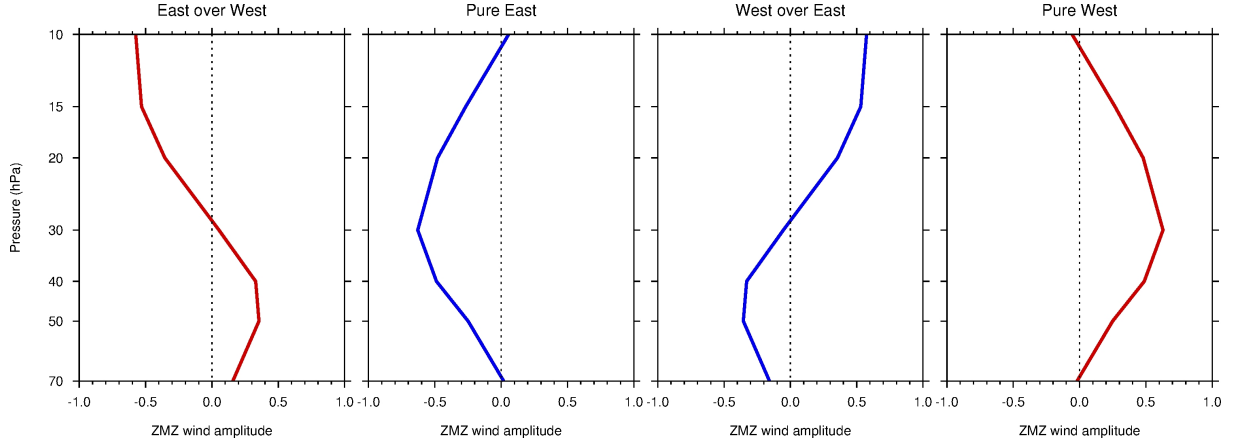


Figure IX.1.: Idealized equatorial $\bar{u}$ profiles for nudging studies. Profiles are constructed from the first two EOFs in the FU Berlin radiosonde dataset.

Monsoon link. As Read & Castrejón-Pita [2012] suggest a synchronisation of the QBO and the tropospheric biennial oscillation to the semiannual oscillation, including an index representing the SAO would be an important step to disentangle the tropical network.

As approaches like composite and regression analysis cannot distinguish between correlation and causality, this new method is a promising approach to establish causal networks in the atmosphere. Specifically for QBO influences, using monthly mean data is shown to be most useful compared to higher time resolution time series. This does not, however, imply that factors varying on shorter time scales, as investigated by Kretschmer *et al.* [2016], can be disregarded.

However, there are caveats to this method. It derives connections between single timeseries, which might be an oversimplification for the climate system. The choice of parameters has to be made very carefully. For example a change in stratospheric zonal mean zonal wind might not directly lead to a change in the amount of rainfall over central India, but rather be due to a modulation of several other parameters. Extension of the CEN approach in the Northern Hemisphere and the Tropics by timeseries representing heat/momentum/wave fluxes or convection would be an obvious part of future analysis to better understand the QBO interaction mechanisms.

One question that could not be addressed yet is which characteristic of the QBO would be the most important to improve in climate models in light of the QBO's relevance in long term forecasting. Given the limited observational record of ~ 60 years, this is likely only to be solved by model studies. As one common model bias is the upward shift of the whole QBO structure, one model experiment would be the comparison of the surface response in a free run and a “nudged” run in which the zonal winds in the equatorial stratosphere from reanalyses are imposed, so that

the QBO extends down to a more realistic level in the atmosphere. This approach could give hints whether the depth of the QBO plays an important role in the teleconnections to the polar stratosphere and/or the surface.

The importance of the QBO structure in the whole stratosphere could be examined by prescribing idealized wind profiles as shown in Figure IX.1. Analysis of wave propagation in these runs could be used to establish the mechanism of the Holton-Tan relationship and the QBO-NAO connection. With these nudged runs it would also be possible to distinguish whether deep states with one QBO phase dominating the stratosphere are indeed the influencing factor for the surface response or whether they develop simultaneously due to the state some months earlier.

Further model studies could involve ensembles of “stalled” vs. “normal” QBO cycles to understand the development and impacts of a prolonged easterly phase in the mid-stratosphere and whether it is an important QBO characteristic.

Exclamations of “Who ordered that?” may have been uttered by stratospheric researchers at the beginning of 2016, when the QBO fooled all forecasts with behaviour unprecedented in the 63 year radiosonde record. Dunkerton [2016] addresses whether this event was singular or might be the precursor to a permanent alteration in the stratospheric circulation, or, as he puts it: was the QBO affected by a hiccup or rather being dragged into a death spiral? There are hints from observations and the MPI-ESM-MR model as to what has caused this unusual event. Newman *et al.* [2016] and Osprey *et al.* [2016] suggest anomalous wave forcing from the extratropics as a likely cause, though understanding the complete mechanism will require more investigations. For example, a GCM could be initialised with the state of the atmosphere/ocean of the winter 2015/16 to establish how likely it was for this disruption to occur. The initial conditions could be varied to check which conditions are necessary (e.g. strong ENSO or the westerly bridge to the North Pole) for disruptions to occur.

Given the future predictions of the MPI-ESM-MR model, the QBO is unlikely to die in the next 100 years, but the occasional hiccup might occur more often. The coming winter will tell whether one should be concerned about this unique disruption in the tropical stratosphere - the 2015/16 event has effectively shifted the QBO by half a cycle, leading to another dominant westerly during Northern Hemisphere winter. After 2013/14, is the UK now in for another stormy and wet winter season?

Appendices

Appendix A - Model overview

Institution	Model Name	Resolution (lat x lon)(°)	# of Levels	Spacing (km)	Top (hPa)	Top (km)	QBO
GFDL	AMTRAC3	2 x 2.5	48	1.59	0.017	76	✗
NCAR	CAM3.5	1.9 x 2.5	26	1.51	3.5	39	✗
NIES	CCSRNIES	2.8125 x 2.8125	34	2.32	0.012	79	✗
Univ. Toronto/CCCma	CMAM	3.8 x 3.8	71	1.37	$8.1 \cdot 10^{-4}$	97	✗
CNRM/GAME, CNRS	CNRM-CACM	2.8 x 2.8	60	1.11	0.07	66	✗
DLR	E39CA	3.75 x 3.75	39	0.82	10	32	✗
MPI-C	EMAC	2.8125 x 2.8125	90	0.89	0.01	80	✓
NASA/GSFC	GEOSCCM	2 x 2.5	72	1.07	0.015	77	✗
CNRS	LMDZrepro	2.5 x 3.75	50	1.33	0.07	66	✗
MRI	MRI	2.8125 x 2.8125	68	1.18	0.01	80	✓
NIWA	NiwaSOCOL	3.75 x 3.75	39	2.05	0.01	80	✗
PMOD/WRC, IAC	SOCOL	3.75 x 3.75	39	2.05	0.01	80	✗
Univ. L'Aquila	ULAQ	11.5 x 22.5	26	2.71	0.04	70	✗
NIWA	UMETRAC	2.5 x 3.75	64	1.25	0.01	80	✗
Univ. Leeds	UMSLIMCAT	2.5 x 3.75	64	1.25	0.01	80	✓
Met Office	UMUKCA-METO	2.5 x 3.75	60	1.40	0.01	84	✓
Univ. Cambridge	UMUKCA-UCAM	2.5 x 3.75	60	1.40	0.01	84	✓
NCAR	WACCM	1.9 x 2.5	66	1.99	$5.96 \cdot 10^{-6}$	132	✗

Table A.1.: Overview over submitted models to the CCMVal-2 intercomparison project. Resolution denotes the horizontal spacing on the latitude-longitude grid in °. The spacing in km is an approximate value for the vertical resolution, calculated as the mean distance between levels assuming an evenly spaced level distribution. Top values indicate the highest computational level. Note that the tropopause is at about 100hPa and the stratopause at about 1hPa. For further details of modelling groups and references see the CMIP5 and CCMVal-2 references (World Climate Research Programme [2010]; SPARC CCMVal [2010]).

Institution	Model Name	Resolution (lat x lon)(°)	# of Levels	Spacing (km)	Top (hPa)	Top (km)	QBO
CSIRO-BOM	ACCESS1.0	0.9 x 1.5	38	1.03	3.52	39	✗
	ACCESS1.3	0.9 x 1.5	38	1.03	3.52	39	✗
BCC	BCC-CSM1.1	2.8125 x 2.8125	26	1.56	2.92	41	✗
	BCC-CSM1.1(m)	1.125 x 1.125	26	1.56	2.92	41	✗
GCESS	BNU-ESM	2.8125 x 2.8125	26	1.64	2.19	43	✗
CCCMA	CanCM4	1.875 x 1.875	35	1.51	0.50	53	✗
	CanESM2	1.875 x 1.875	35	1.51	0.50	53	✗
NCAR	CCSM4	0.9 x 1.25	27	1.58	2.19	43	✗
NSF-DOE-NCAR	CESM1(BGC)	0.9 x 1.25	27	1.58	2.19	43	✗
	CESM1(CAM5)	0.9 x 1.25	27	1.58	2.19	43	✗
	CESM1(CAM5.1.FV2)	1.9 x 2	30	1.07	10	32	✗
	CESM1(WACCM)	1.9 x 2.5	66	2.01	$5 \cdot 10^{-6}$	133	✗
	CESM1(FASTCHEM)	0.9 x 1.25	27	1.58	2.19	43	✗
CMCC	CMCC-CESM	3.75 x 3.75	39	2.05	0.01	80	✗
	CMCC-CM	0.75 x 0.75	31	1.03	10	32	✗
	CMCC-CMS	1.875 x 1.875	95	0.84	0.01	80	✓
CNRM-CERFACS	CNRM-CM5	1.4 x 1.4	31	1.03	10	32	✗
CSIRO-QCCCE	CSIRO-Mk3.6.0	1.875 x 1.875	18	2.09	4.50	38	✗
EC-EARTH	EC-EARTH	1.1 x 1.1	62	0.77	1.00	48	✗
LASG-CESM	FGOALS-g2	2.8125 x 2.8125	26	1.64	2.19	43	✗
LASG-IAP	FGUALS-s2	2.81 x 1.66	26	1.64	2.19	43	✗
FIO	FIO-ESM v1.0	2.8125 x 2.8125	26	1.51	3.55	39	✗
NOAA GFDL	GFDL-CM2.1	2.5 x 2	24	1.63	3.65	39	✗
	GFDL-CM3	1.8 x 1.8	48	1.67	0.01	80	✗
	GFDL-ESM2G	2.5 x 2	24	1.63	3.65	39	✗
NASA GISS	GISS-E2-H	2.5 x 2	40	1.60	0.10	64	✗
	GISS-E2-H-CC	1 x 1	40	1.60	0.10	64	✗
	GISS-E2-R	2.5 x 2	40	1.60	0.10	64	✗
	GISS-E2-R-CC	1 x 1	40	1.60	0.10	64	✗
NIMR/KMA	HadGEM2-AO	1.875 x 1.25	60	1.40	0.01	84	✗
MOHC	HadCM3	3.75 x 2.5	19	4.46	0.01	85	✗
	HadGEM2-CC	1.875 x 1.25	60	1.40	0.01	84	✓
	HadGEM2-ES	1.875 x 1.25	38	1.03	3.52	39	✗
INM	INM-CM4	2 x 1.5	21	3.81	0.01	80	✗
IPSL	IPSL-CM5A-LR	1.9 x 3.75	39	1.80	0.04	70	✗
	IPSL-CM5A-MR	1.25 x 2.5	39	1.80	0.04	70	✗
	IPSL-CM5B-LR	1.9 x 3.75	39	1.80	0.04	70	✗
MIROC	MIROC4h	0.5625 x 0.5625	56	0.87	0.90	49	✗
	MIROC5	1.40625 x 1.40625	40	1.02	2.90	41	✗
	MIROC-ESM	2.8125 x 2.8125	80	1.10	$3 \cdot 10^{-3}$	88	✓
	MIROC-ESM-CHEM	2.8125 x 2.8125	80	1.10	$3 \cdot 10^{-3}$	88	✓
MPI-M	MPI-ESM-LR	1.8 x 1.8	47	1.70	0.01	80	✗
	MPI-ESM-MR	1.8 x 1.8	95	0.84	0.01	80	✓
	MPI-ESM-P	1.8 x 1.8	47	1.70	0.01	80	✗
MRI	MRI-EGCM3.2H	0.5625 x 0.5625	64	1.25	0.01	80	✗
	MRI-AGCM3.2S	0.1875 x 0.1875	64	1.25	0.01	80	✗
	MRI-CGCM3	1.125 x 0.7	48	1.67	0.01	80	✗
	MRI-ESM1	1.125 x 1.125	48	1.67	0.01	80	✗
NCEP	CFSv2-2011	0.9375 x 0.9375	64	1.13	0.03	72	✗
NCC	NorESM1-M	1.9 x 2.5	26	1.64	2.19	43	✗
	NorESM1-ME	1.9 x 2.5	26	1.64	2.19	43	✗

Table A.2.: Overview over submitted models to the CMIP5 intercomparison project. As A.1

Appendix B - Additional Figures

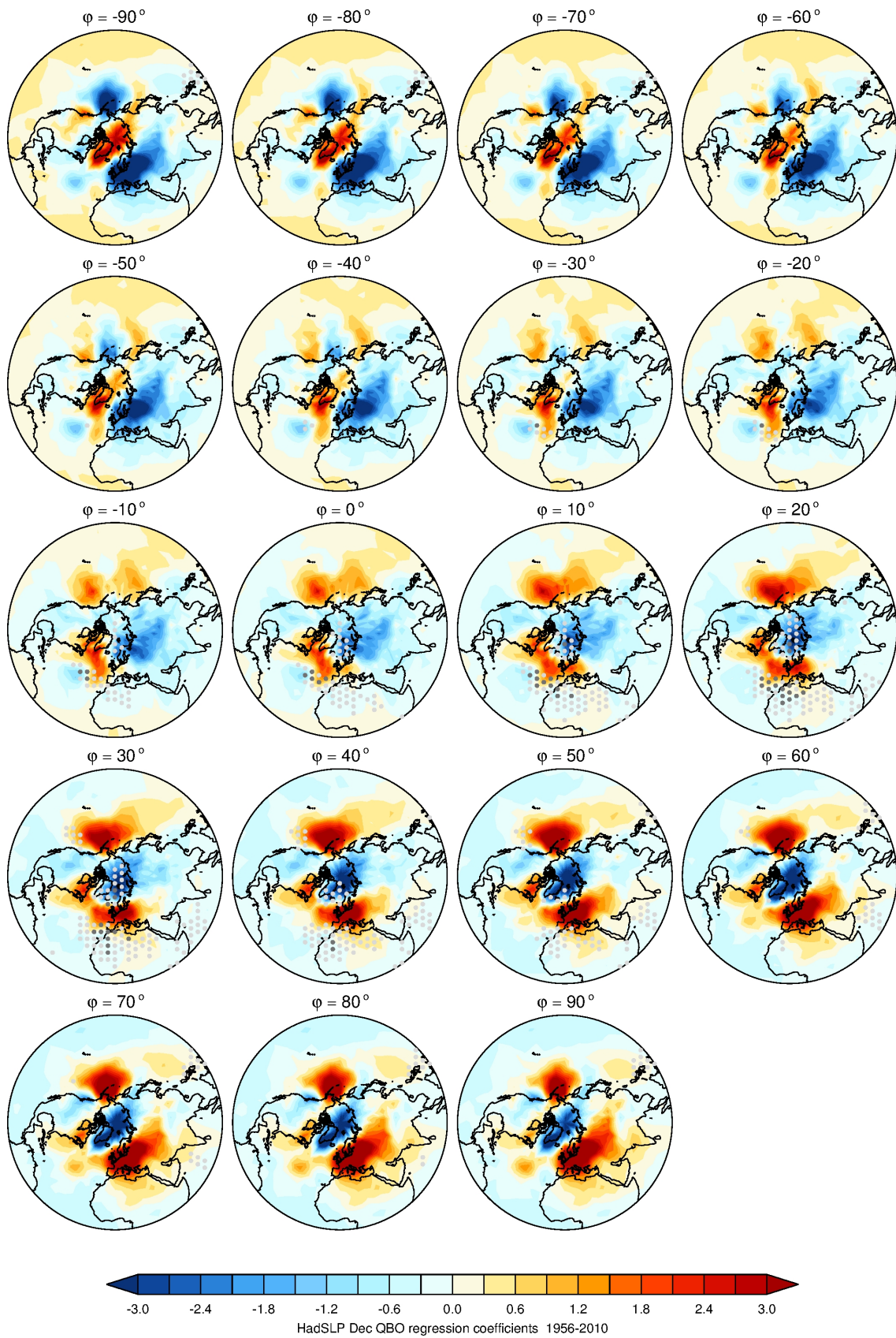


Figure B.2.: As Figure V.9 for December.

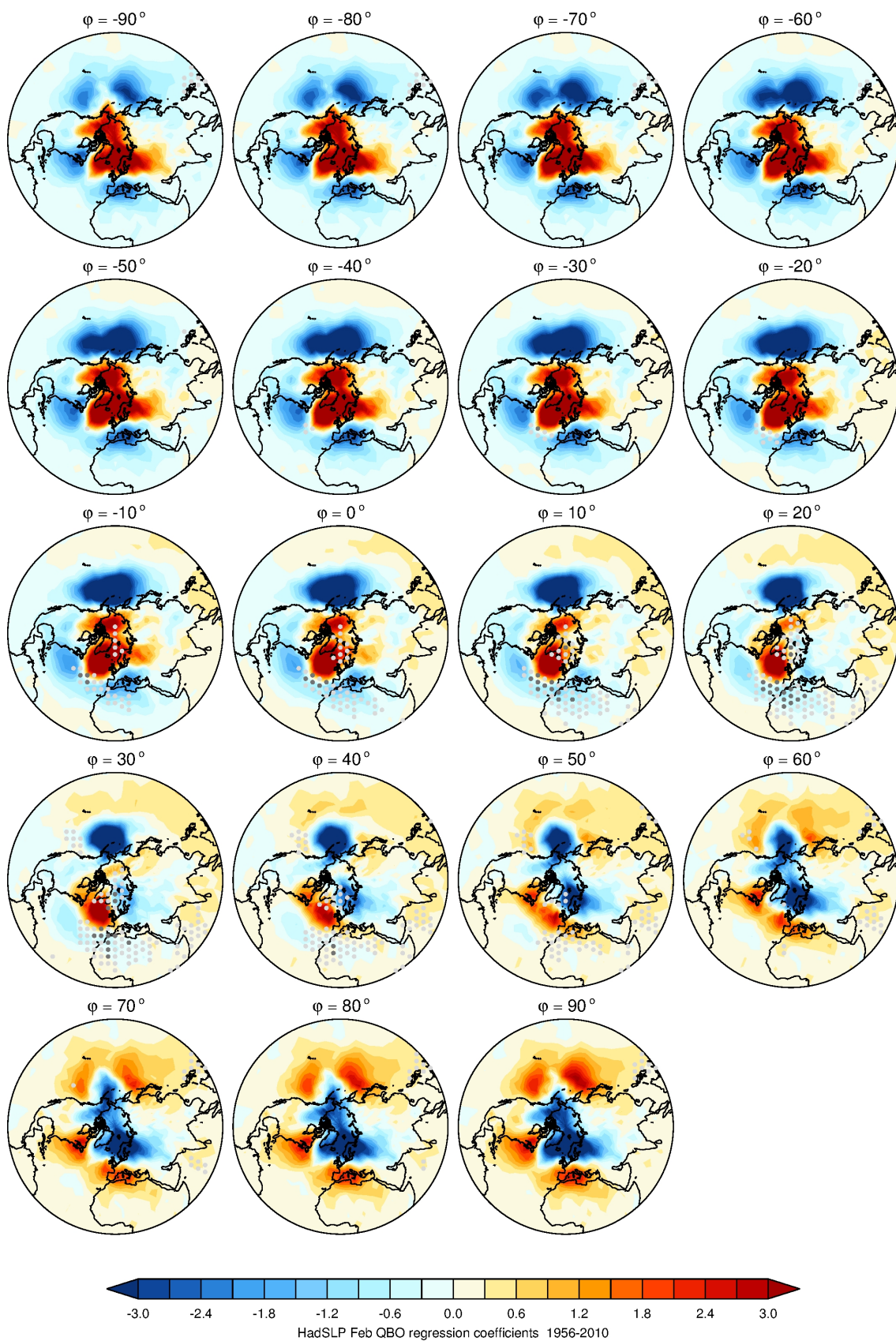


Figure B.3.: As Figure V.9 for February.

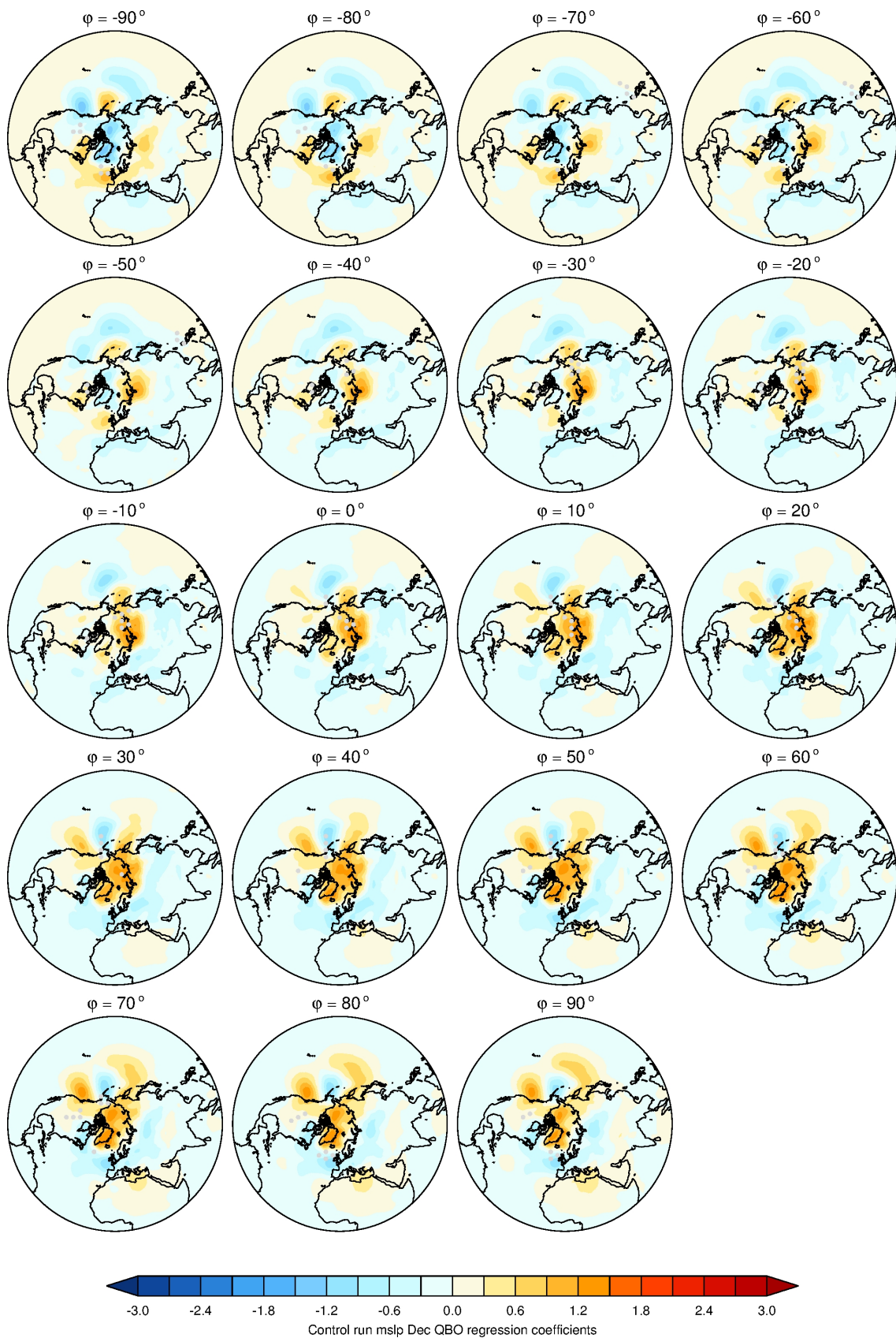


Figure B.4.: As Figure V.16 for December.

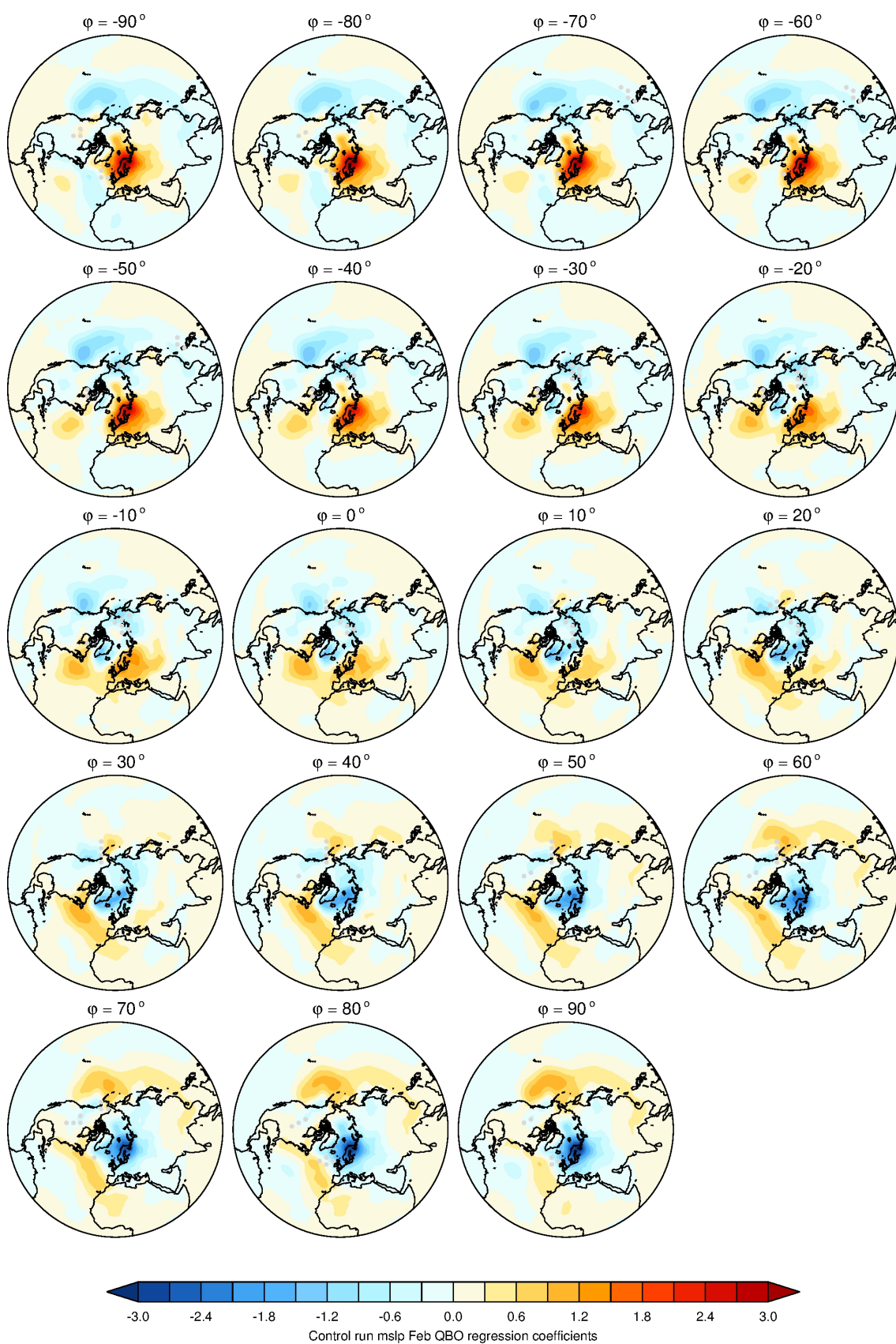


Figure B.5.: As Figure V.16 for February.

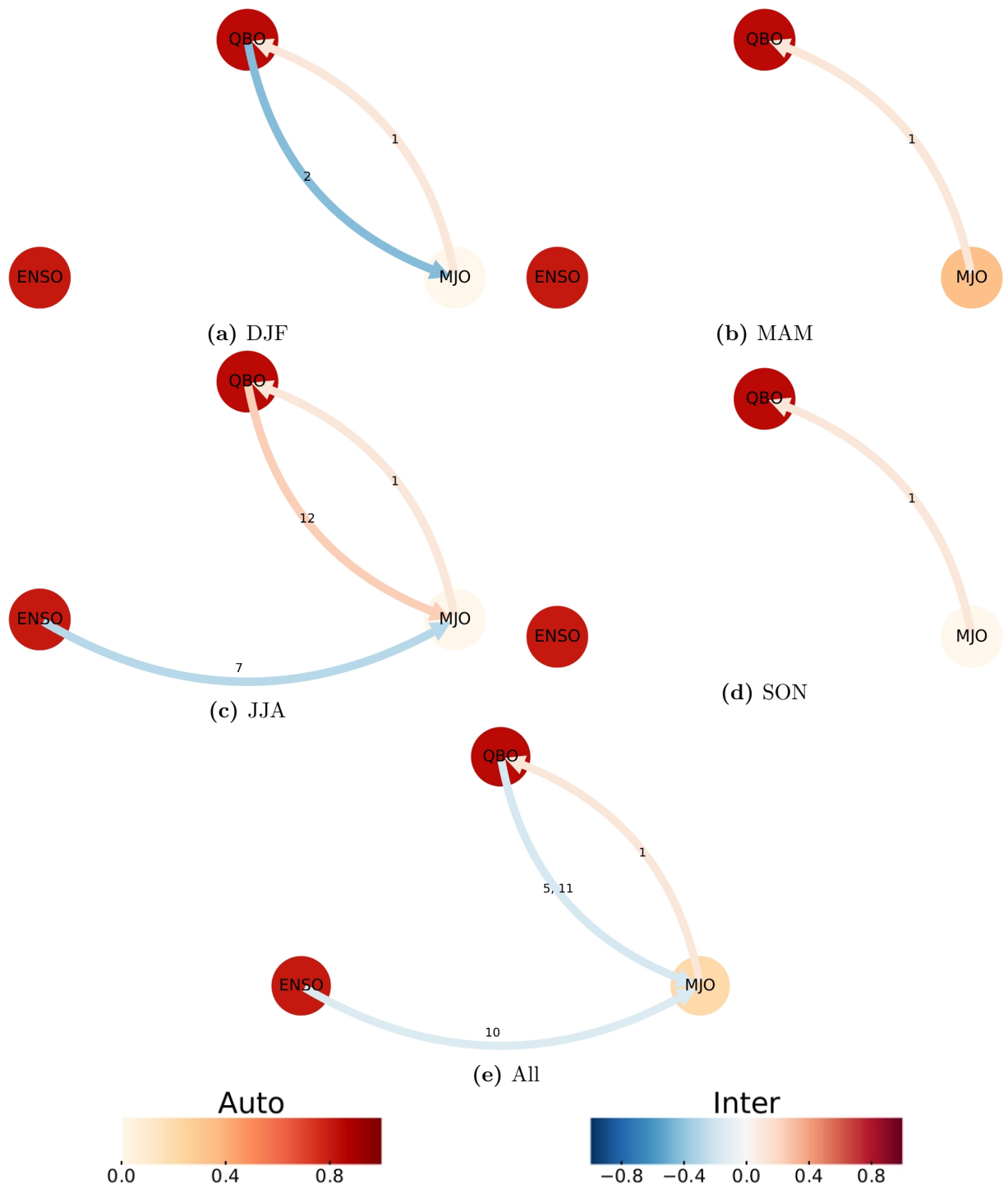


Figure B.6.: As Figure VII.14, using the equatorial zonal mean zonal wind at 50hPa as QBO index.

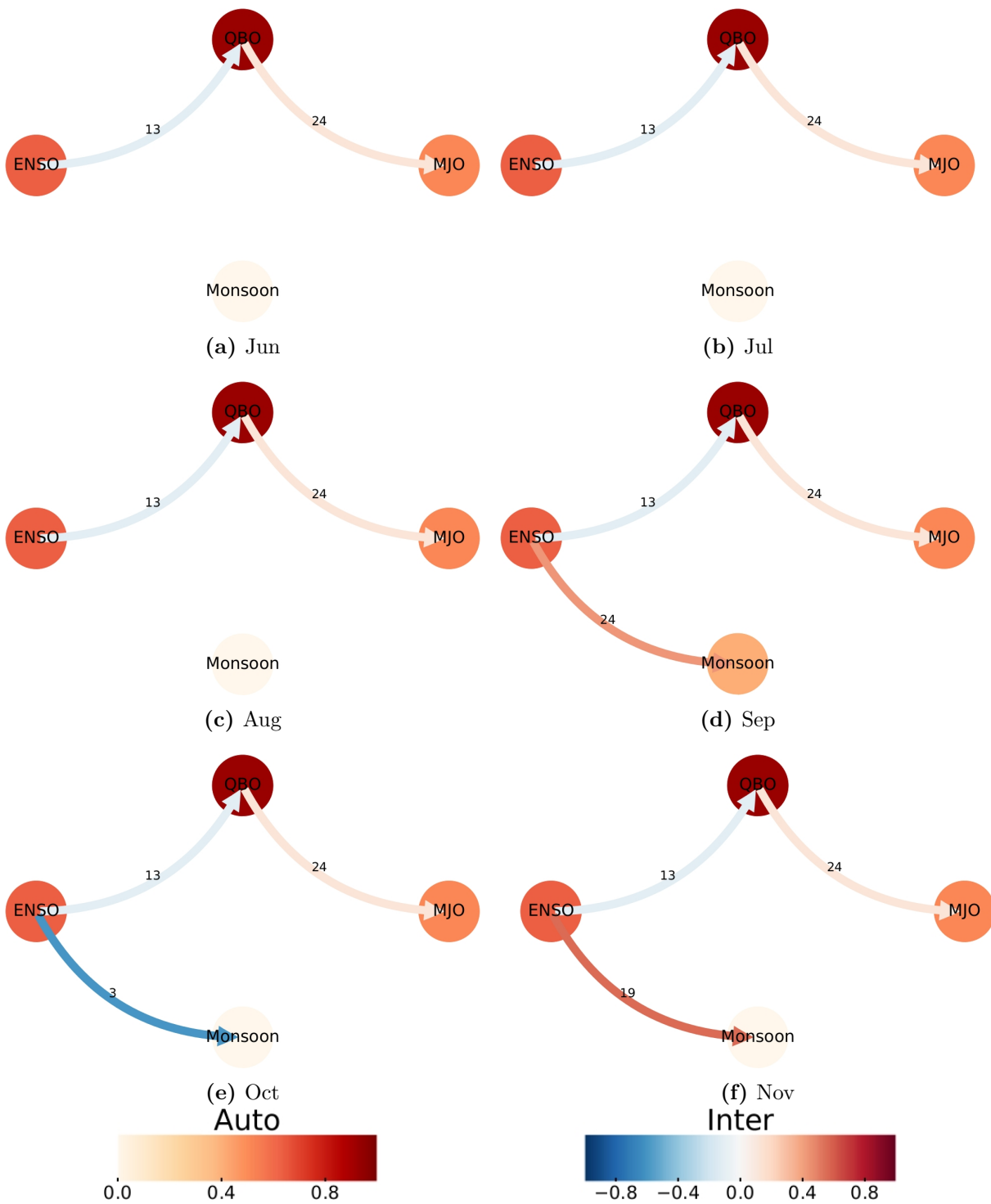


Figure B.7.: As Figure VII.18, but at fortnightly time resolution. Lags are given in fortnights.

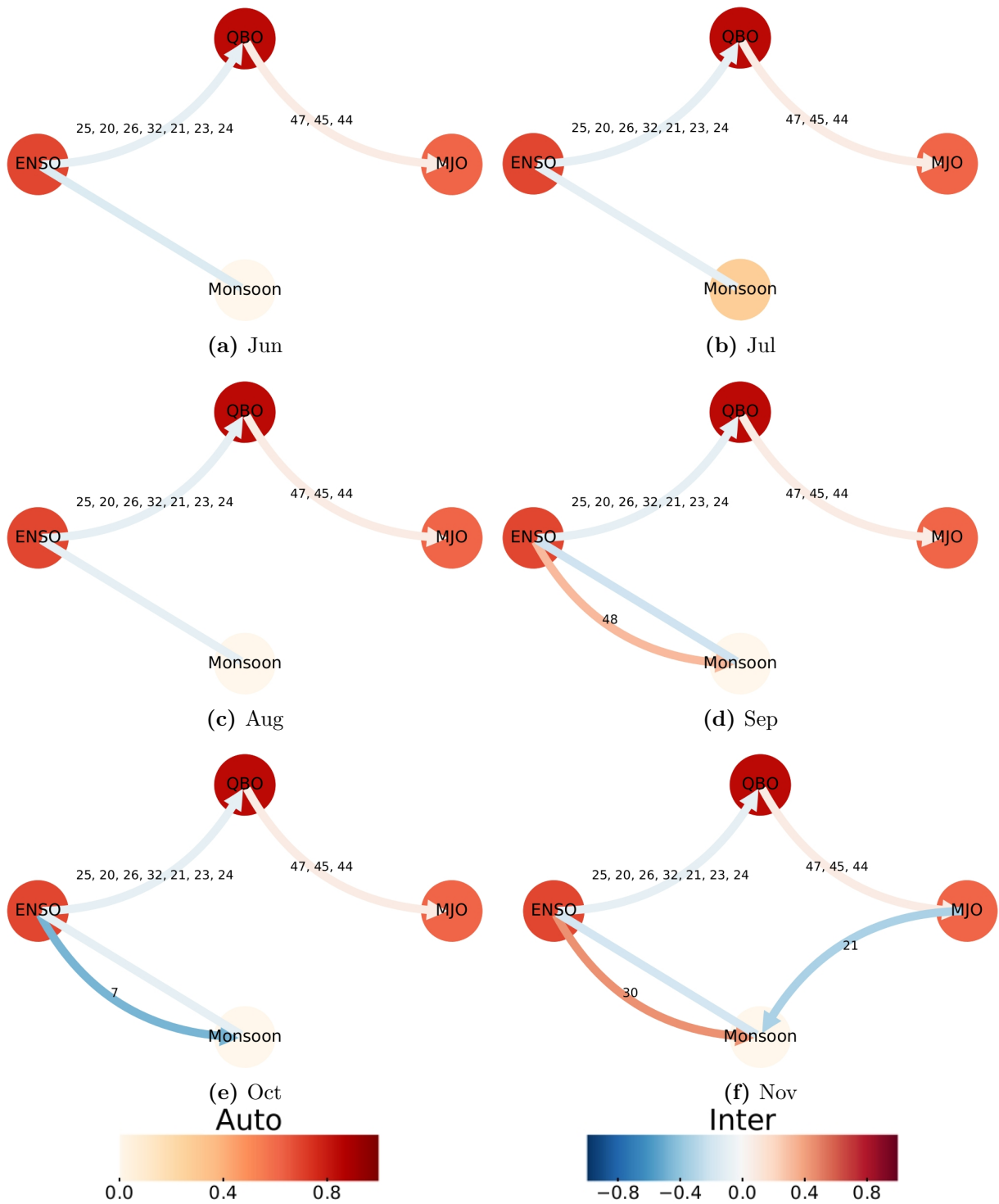


Figure B.8.: As Figure VII.18, but at weekly time resolution. Lags are given in weeks.

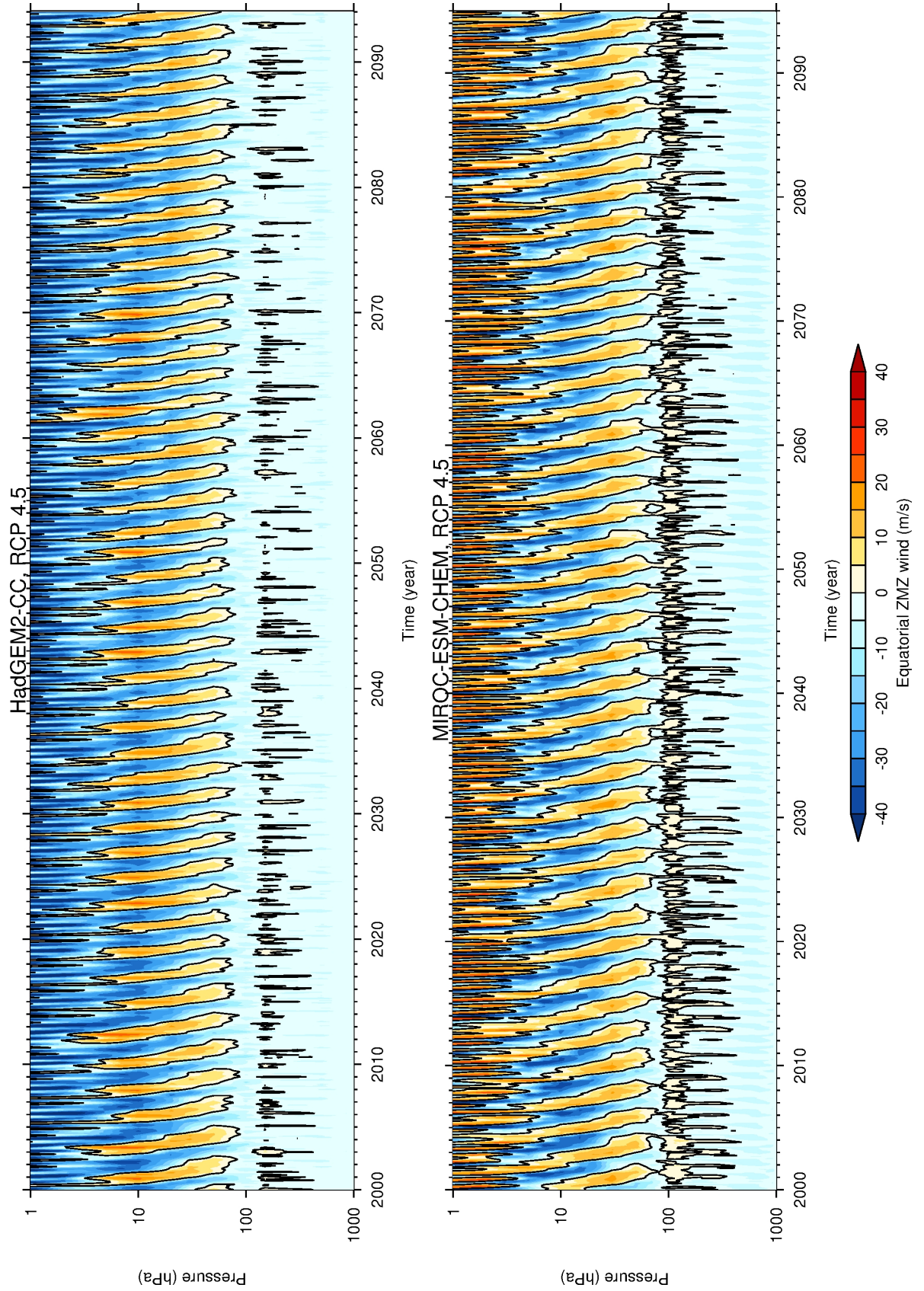


Figure B.10.: Equatorial zonal mean zonal wind in the RCP 4.5 emission scenario from HadGEM2-CC (upper) and MIROC-ESM-CHEM (lower). The zero wind line is shown in black.

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