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The influence of the stratosphere on the tropical troposphere

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Abstract

Observational and model studies suggest that the stratosphere exerts a significant influence on the tropical troposphere. The corresponding influence, through dynamical coupling, of the stratosphere on the extratropical troposphere has over the last 15-20 years been intensively investigated, with consequent improvement in scientific understanding which is already being exploited by weather forecasting and climate prediction centres. The coupling requires both communication of dynamical effects from stratosphere to troposphere and feedbacks within the troposphere which enhance the tropospheric response. Scientific understanding of the influence of the stratosphere on the tropical troposphere is far less developed. This review summarises the current observational and modelling evidence for that influence, on timescales ranging from diurnal to centennial. The current understanding of potentially relevant mechanisms for communication and for feedbacks within the tropical troposphere and the possible implications of the coupling for weather and climate prediction are discussed. These include opportunities for model validation and for improved subseasonal and seasonal forecasting and the effects, for example, of changes in stratospheric ozone and of potential geoengineering approaches. Outstanding scientific questions are identified and future needs for observational and modelling work to resolve these questions are suggested.

Keywords stratosphere, troposphere, convection, quasi-biennial oscillation, intraseasonal oscillation

1. Introduction

Chemical, radiative or dynamical coupling between troposphere and stratosphere is an important aspect of the climate system. For example: ozone produced in the stratosphere can, when transported into the troposphere, have an important effect on tropospheric chemistry and air quality (e.g. Monks et al. 2015); the stratospheric concentrations of radiatively active gases such as ozone and water vapour can play an important role in the thermal balance of the troposphere (e.g. Forster and Shine 2002, Forster et al. 2007); waves on scales of km to tens of thousands of km can communicate dynamical information between troposphere and stratosphere (e.g. Baldwin et al. 2018). Naive arguments suggest that since the mass of the troposphere is much more than the mass of the stratosphere, any important dynamical coupling will be from the troposphere to the stratosphere. But such arguments, which might also be applied to chemical and radiative coupling, neglect the sensitivity of the system. Just as chemical and radiative sensitivity means that very small stratospheric concentrations of ozone and water vapour can have strong effects on the chemical and radiative balance of the troposphere, dynamical sensitivity means that there can be strong dynamical coupling from the mid-stratosphere (20-25 km) to the mid-troposphere (5-10 km), notwithstanding the factor of 10 difference in density between those levels.

During the last 15-20 years there has been a major research focus on the coupling from the stratosphere to the extratropical troposphere (e.g. Gerber et al. 2010, Kidston et al. 2015). Research has progressed from a handful of individual observational and modelling studies, through parallel lines of investigation addressing key theoretical issues, testing hypotheses using models across a range of complexity and demonstrating important effects in state-of-the-art numerical models used for weather, climate and chemistry-climate prediction. This progress has led to exploitation in operational seasonal weather prediction, e.g. as reported by Fereday et al. (2012) who argue that including a better representation of the stratosphere allows a more accurate representation of the effects of initial conditions in sea-surface temperatures and equatorial stratospheric winds. It has also led to appreciation of the importance of model representation of the stratosphere for climate prediction, with studies such as Scaife et al. (2012), Manzini et al. (2014) and Simpson et al. (2018) arguing that model-to-model variation in predicted stratospheric change has a strong effect on the predicted change in tropospheric circulation in the Northern Hemisphere (NH) with important implications for predictions of mid-latitude weather and hydroclimate. The benefit of better representation of the stratosphere has also been demonstrated for seasonal forecasting for Southern Hemisphere midlatitudes (e.g. Hendon et al. 2020)

Coupling from the stratosphere to the tropical troposphere has received much less attention but could also potentially be exploited in significant ways in weather and climate prediction. Early studies such as that of Gray (1984), who found a statistical connection between the quasi-biennial oscillation (QBO) in tropical stratospheric winds and the frequency of Atlantic hurricanes, understandably prompted widespread interest (Gray's paper has ~500 citations). Subsequent analysis as the data record has lengthened (Camargo and Sobel 2010) has shown that there is no such statistical connection for the period mid-1980s to late 2000s. Whilst the existence of a robust QBO-hurricane connection might now be more uncertain, several other potential effects of the stratosphere on the tropical troposphere have been identified or suggested, including, quite recently an effect of the QBO on the Madden-Julian Oscillation (MJO) (Yoo and Son 2016) which dominates intraseasonal variability in the tropical troposphere. Furthermore effects of the stratosphere on the tropical troposphere have also been argued to be potentially important in future tropical climate (e.g. Nowack et al. 2015) and in the climate response to geoengineering (e.g. Simpson et al. 2019).

Many of the details of stratosphere-troposphere coupling in the tropics are expected to be very different to those in the extratropics. One aspect is that the potential dynamical mechanisms for communication between stratosphere and troposphere are different. The small values of the Coriolis parameter in the tropics mean that in balanced dynamics the natural aspect ratio of vertical to horizontal length scales, determined by the form of the potential vorticity (PV) inversion operator, is small, so dynamical structures are naturally shallow. Alongside this there is a larger role for unbalanced dynamics, in convection or in wave propagation, in communication of information in the vertical. The second distinct aspect is that the potential dynamical feedbacks within the troposphere, which may enhance the tropospheric response, are different because of the very different nature of the dynamics and thermodynamics of the tropical troposphere compared to that of the extratropical troposphere. The latter is dominated by interaction between baroclinic eddies and the larger scale environment of jets and planetary-scale Rossby waves. This interaction is now recognised as fundamental for coupling from the stratosphere to the extratropical troposphere and indeed more generally for determining future changes in the circulation of the extratropical troposphere. The tropical analogue is self-organisation and corresponding internal variability on scales of 100s to 10000s of km of strongly convective regions and their non-convective environment, interacting through dynamical and cloud-radiative processes and moisture transport. It is these interactions that are likely to play a major role in any coupling from the stratosphere to the tropical troposphere.

This review will summarise the current observational and modelling evidence for an influence of the stratosphere on the tropical troposphere and the possible implications of this for prediction. Outstanding scientific questions will be identified and future needs for observational and modelling work to resolve these questions will be discussed. As with many topics in climate science, ideas on stratosphere-troposphere coupling have developed over decades through interplay between the three different strands of observational studies, modelling studies and the development and application of theory for the relevant dynamical and physical processes and the interactions between them. Dividing between these three strands is difficult and to some extent arbitrary, but facilitates presentation. The choice made here is as follows. Section 2 will give a brief overview of possible pathways and mechanisms for coupling, based on theoretical ideas for large-scale tropospheric and stratospheric dynamics. Section 3 will then set out the observational evidence for coupling and Section 4 will give a more detailed account of model investigations relevant to identifying and assessing specific mechanisms, including many of the important aspects of the dynamics and physics of the tropical troposphere. These investigations cover phenomena on a wide range of timescales for diurnal to centennial, but they are presented together in this Section in order to emphasise that certain mechanisms are relevant across this range. Section 5 will discuss some of the practical implications of coupling for weather and climate prediction. Section 6 will summarise, identify outstanding scientific questions and suggest ways in which those questions might be addressed. Some of the topics included in Sections 2 to 4 have been discussed by Gray et al. (2018) who focus on the effect of the QBO on both the extratropical and the tropical troposphere and by Hitchman et al. (2021) in a review of historical development of evidence for links from the QBO to the tropical troposphere. Some of the prospects mentioned in Section 6 for exploiting stratosphere-troposphere coupling in the subtropics and tropics to improve subseasonal to seasonal forecasting have recently been reviewed independently by Butler et al. (2019), see also Alexander and Holt (2019). The intention of this review is to provide a more detailed discussion of observations, models and mechanisms relevant to stratosphere-troposphere coupling in the tropics, extending beyond QBO effects to cover as wide a range of timescales as possible.

2. Pathways and tropospheric feedbacks

A major stimulus to research on stratosphere-troposphere coupling in the extratropics has been the suggestion that there are tropospheric signals of the QBO and of the state of the stratospheric wintertime extratropical circulation, in particular the occurrence of stratospheric sudden warmings (SSWs), which are major disruptions to the extratropical wintertime stratospheric circulation.

In the large body of previous research on mechanisms for extratropical stratosphere-troposphere coupling, various pathways have been suggested for these signals, one apparently originating in the tropical stratosphere and the other in the extratropical stratosphere, to be communicated to the extratropical troposphere. It is useful to summarise these alongside the pathways that may be relevant for communicating stratospheric signals to the tropical troposphere. Note that in this previous research it has been important to consider not only (A) pathways for communication from stratosphere to troposphere but also (B) the feedbacks within the troposphere that determine the magnitude of the resulting response. This section will consider (A) first and then (B).

Figure 1 shows a schematic diagram of the different principal pathways that may be relevant for communication from the stratosphere to the troposphere, both the tropical troposphere and the extratropical troposphere, of (a) the QBO signal (or any other effect originating in the low-latitude stratosphere) and (b) the SSW signal (or any other signal originating in the mid/high-latitude stratosphere). Gray et al. (2018) showed a similar schematic diagram focusing on pathways relevant to the QBO signal. Note that what are shown in Figure 1 are pathways for communication of dynamical signals, not pathways for transport of chemical species. Figure 1 should be clearly distinguished from schematic diagrams of transport pathways for stratosphere-troposphere exchange, as shown in e.g. Holton et al (1995), Stohl et al (2003).

2.1 The QBO as a source of variability in the low-latitude stratosphere

Given the prominence of the QBO as an example of potential stratospheric influence on the tropical troposphere, this sub-section gives a very brief review of its primary characteristics. The QBO is manifested by quasi-periodic variation, on a time scale of about 28 months, in winds and temperatures in the tropical stratosphere. The basic dynamics of the QBO is well understood and has been reviewed, for example, by Baldwin et al. (2001). Whilst the QBO is fundamentally a tropical phenomenon its effects extend to the extratropical stratosphere and hence to the extratropical troposphere. (The term 'tropical QBO' will sometimes be used to emphasise that what is meant is the phenomenon of oscillation in low-latitude winds and temperatures rather than a 'QBO signal' which extends away from the tropical stratosphere.) The key features of the tropical QBO that might affect the troposphere are the changes in the stratospheric winds and corresponding changes to stratospheric temperatures. The latter arise because the tropical QBO has a finite latitudinal width. Whilst the Coriolis force is zero at the equator itself, it is non-zero away from the equator. Therefore to meet the requirement of thermal wind balance, there must be latitudinal and vertical variation in temperature. The relation between winds and temperatures is captured in 2-D models such as that of Plumb and Bell (1982) and Figure 2 shows this relation schematically. The implication of Figure 2 for the lowest part of the stratosphere, which is likely to be the most important part for any stratosphere-troposphere coupling, is that temperatures will be relatively warm or relatively cold according to whether the QBO winds just above are westerly or easterly. Observations show that the dominant temperature structure is confined to [15S,15N]. Outside this range of latitudes there is a weaker temperature signal of the opposite sign. The magnitude of the QBO-related temperature signal averaged in longitude and across tropical latitudes is around 1 K peak-to-peak at the tropical tropopause (e.g. Huessman and Hitchman 2001, Zhou et al. 2001), though significantly larger in certain regions and seasons (Hitchman et al. 2021), increasing to more than 5 K peak-to-peak above 20 km (e.g. Randel and Wu 2015). Figure 3 shows some further details of interannual variation in temperatures and the relation to the QBO winds.

2.2 Pathways

The three pathways depicted in Figure 1 are as follows. The Extratropical Pathway (1), vertically from the extratropical stratosphere to the extratropical troposphere, is the generally accepted route for extratropical coupling (e.g. Kidston et al. 2015). The mechanisms that are likely to play a role in this pathway are (i) the instantaneous vertical non-locality of extratropical dynamics implied by PV inversion, as considered by Charlton et al. (2005), (ii) the modification of that by radiative transfer acting on temperatures, which acts to deepen dynamical structures (Haynes et al. 1991, Song and Robinson 2004) and, very importantly, (iii) downward propagation of information by large-scale waves[1], even if net large-scale wave propagation, e.g. as measured by wave fluxes, is upwards (Perlwitz and Harnik 2004, Song and Robinson 2004, Scott and Polvani 2004, Martineau and Son 2015, Hitchcock and Simpson 2016, Hitchcock and Haynes 2016). This pathway is clearly relevant for communication of SSWs (and of other dynamical events in the extratropical stratosphere). It is also relevant for the communication of the tropical QBO, if one accepts that the latter affects the circulation in the extratropical stratosphere. There is convincing evidence from modelling and observational studies that there is such an effect, though the mechanism is probably more complicated than that originally suggested by Holton and Tan (1980, 1982) in their papers which identified an extratropical QBO signal in observations (e.g. see Yamashita et al. 2011, Garfinkel et al. 2012, Anstey and Shepherd 2014).

The Extratropical Pathway is relevant for coupling from the stratosphere to the tropical troposphere if a change in the extratropical troposphere can be subsequently communicated, within the troposphere, to the tropics. For example, Kuroda (2008) has argued that such communication is relevant to correlations between SSWs, when the westerly polar vortex is

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[1] Note that ‘downward propagation of information’ implies that ‘propagation’ is being used here in the sense of ‘group propagation’. Where ‘phase propagation’ is meant that will be explicitly stated. See beginning of Section 4 for further comment on this point..

unusually weak, or ‘Vortex Intensification’ events, when it is unusually strong, and the tropical troposphere.

The Subtropical Pathway (2), from the tropical lower stratosphere to the troposphere via the subtropical jet is another possible route for stratosphere-troposphere coupling. This has been suggested by Garfinkel and Hartmann (2011) as a pathway for the QBO to affect the extratropical troposphere and was also discussed by Inoue et al. (2011) and Inoue and Takahashi (2013). Garfinkel and Hartmann (2011) described this as the effect of the ‘meridional circulation of the QBO’, though it is important to realise that this requires more than the zonally symmetric dynamics included in the Plumb and Bell (1982) description of this meridional circulation. The ability of such dynamics to extend a QBO signal into the subtropics is limited (e.g. Plumb 1982) and it is likely that the mechanism acting in the Garfinkel and Hartmann (2011) simulations is better described, following Inoue et al. (2011) and Inoue and Takahashi (2013), as a coupled response of the mean flow and synoptic-scale and planetary-scale eddies which originate in the extratropics and dissipate in the subtropics. Changes in the subtropical troposphere, and in the subtropical jet in particular, could also be communicated to lower latitudes, e.g. by changing the strength and frequency of PV intrusions into the subtropical upper troposphere and correspondingly the effect on tropical convection. (See Section 3.2 below.) The Subtropical Pathway could also be relevant for any tropical tropospheric response to SSWs, if the previously mentioned meridional circulation response first communicates the effect of the SSW to the subtropical lower stratosphere.

The Tropical Pathway (3) is directly from the tropical lower stratosphere to the tropical troposphere and requires a mechanism by which temperature or wind changes in the tropical lower stratosphere can be communicated to the troposphere. The vertical non-locality of dynamics associated with PV inversion (and its radiative modifications) is restricted to small vertical scales in the tropics, because the Coriolis parameter is small. Therefore if stratospheric effects are to penetrate significantly into the troposphere some other mechanism for vertical communication is required. The first suggestions for such a mechanism invoked the possibility that deep convection, in which air parcels move rapidly from the surface to the tropopause, might be affected by changes to near-tropopause conditions (temperature, stratification and wind) and communicate those changes effectively through the depth of the troposphere. Gray et al. (1992a) argued that convection was sensitive to tropopause-level vertical wind shear, with strong shear inhibiting convection. The effect of the QBO on convection would therefore be modulated by the background geographical variation in wind shear, since the QBO would in some locations reinforce the background shear and in some locations diminish it, with these locations varying according to the QBO phase. In a subsequent paper Gray et al. (1992b) argued that deep convection might be affected by the change in static stability around the tropical tropopause associated with the QBO effect on temperatures in the very lowest part of the tropical stratosphere, which are warm when tropical lower stratospheric winds are westerly (QBOW) and cold when they are easterly (QBOE). For example, reduced static stability around the tropopause in QBOE would allow convection to penetrate higher than in QBOW. A third mechanism suggested by Collimore et al. (2003) was that upper-tropospheric large-scale vorticity variations associated with the QBO might affect deep convection, through the effect of absolute vorticity on convective outflow, with more anticyclonic absolute vorticity, associated with QBOE, implying stronger convection.

All these proposed mechanisms, particularly the first two, for downward influence from the tropopause and lower stratosphere to the convectively active main body of the tropical troposphere, have been repeatedly mentioned in work on QBO connections to the tropical troposphere (Collimore et al. 1998, Giorgetta et al. 1999, Collimore et al. 2003, Liess and Geller 2012, Huang et al. 2012). However for none of these is there yet any accepted concrete physical model that might allow a quantitative estimate of the sensitivity. Furthermore, whilst evidence has been presented (e.g. by Collimore et al. 2003) that the effect of the QBO is strongest in regions where convection penetrates highest, this does not explain all aspects of the strong geographical variation in the apparent tropospheric QBO signal. Only very recently has a response of tropical deep convection to QBO-like tropopause level temperature changes been demonstrated in convection-permitting modelling studies (Nie and Sobel 2015, Yuan 2015). These studies will be described in more detail in Section 4. A further distinct mechanism for vertical communication might be through wave propagation, analogous to the vertical communication in the extratropics through Rossby wave propagation that seems very likely to be important for the Extratropical Pathway. A realisation of such a mechanism is provided by the idealized modelling studies of Nishimoto et al. (2016) and Bui et al. (2017, 2019), also described in more detail in Section 4.

Note that the distinction between the Subtropical Pathway and the Tropical Pathway might be questioned on the basis that variations in the subtropical jet are inextricably linked to variations in the tropical upper troposphere. However the two Pathways might also be distinguished on the basis of the physics of the relevant processes – the Subtropical Pathway as dominated by ‘balanced’ PV dynamics of the subtropical jet and the Tropical Pathway as dominated by a more direct effect (e.g. through the mechanisms mentioned above) on the dynamics and thermodynamics of tropical convective systems.

In practice, of course, for any particular stratospheric effect on the tropical troposphere identified in observational studies or in model simulations, a combination of the Pathways described above may be important and it may be difficult to identify a single Pathway which dominates. In particular an apparent tropical tropospheric response to the QBO or to SSWs may in principle arise through any of the Extratropical, Subtropical or Tropical Pathways. Gray et al. (2018) attempted to address this in their multiple regression study of QBO effects on the extratropical troposphere by including an extra regression variable which is a measure of polar vortex variation. They found that the QBO signals in subtropical and tropical tropospheric winds remain, suggesting that it is the Subtropical or Tropical Pathways that are responsible for these signals (i.e. not QBO induced variation of the polar vortex which is then transmitted to the troposphere via the Extratropical Pathway and then within the troposphere to low latitudes).

2.3 Tropospheric feedbacks

It was argued above that it is useful to consider separately communication from stratosphere to troposphere and feedbacks within the troposphere. For the extratropics (A in Figure 1) research has shown that an important feedback mechanism that shapes and potentially amplifies the response of the troposphere to stratospheric changes is the two-way interaction between the large-scale tropospheric flow and synoptic-scale eddies (i.e. weather systems) (Hartmann et al. 2000, Polvani and Kushner 2002, Kushner and Polvani 2004, Song and

Robinson 2004, Chen and Plumb 2009, Simpson et al. 2009, Hitchcock and Simpson 2014, 2016). This two-way interaction is also a key part of the mechanism for internal low-frequency variability, such as the North Atlantic Oscillation or the Northern Annular Mode (or the Southern Annular Mode), in the extratropical troposphere. It is also key to the general problem of the response of the extratropical tropospheric circulation to any ‘external’ forcing, including increases in greenhouse gases (e.g. Lu et al. 2008). Note that the ‘two-way’ character of this interaction is important. Therefore, whilst work such as Wittman et al. (2007) which considered only the effect of mean flow changes on the eddies via ‘baroclinic life-cycle experiments’ was a useful contribution to building understanding, a major part of the important feedback is missed (Hitchcock and Simpson 2016). Complete dynamical understanding of this interaction remains elusive, both of its role in determining variability and of its role in determining forced response. Nonetheless it is now widely accepted and has been exploited in seasonal weather forecasting, for example, that a large part of the signal of extratropical stratosphere-troposphere coupling appears as changes to the tropospheric flow that have similar spatial structure to the Northern or Southern Annular Mode.

In the tropics any mechanisms for feedbacks within the troposphere that might shape and amplify the response to changes in the stratosphere are likely to be completely different to those in the extratropics, but just as for that case, they are likely to be relevant also to broader phenomena of tropical low-frequency variability (e.g. Jiang et al. 2015) and of the tropical response to increasing greenhouse gases (e.g. Voigt and Shaw 2015). As was noted above for the Tropical Pathway, relevant mechanisms are likely to involve convective systems, but detailed investigation of the viability of such mechanisms has begun only recently.

3. Observational studies and data analyses

3.1 Influence of the QBO on the tropical troposphere

The QBO in tropical stratospheric winds (see Section 2.1 above) has well-established effects on the circulation in the extratropical stratosphere (Holton and Tan 1980, Dunkerton and Baldwin 1991, Naito and Hirota 1997, Anstey and Shepherd 2014). These effects are typically quantified in observations or in models by choosing different measures of the circulation, perhaps averaged over each individual month or over each year, then forming composites according the sign of the QBO winds at a particular reference level, and taking the difference between the two. A characteristic feature of the QBO is the downward phase propagation of the wind signal (recall Figures 2 and 3). For example, when QBO winds at 70 hPa (about 18km) are westerly, they are typically easterly at 10 hPa (about 30 km). Thus the choice of the reference level that is used to define QBOE and QBOW composites will significantly affect the deduced QBO signal in whatever measure of the tropospheric circulation is being considered. Different studies of the extratropical QBO signal have often chosen different reference levels which makes their results difficult to compare. The same potential difficulty applies to studies of possible QBO signals in the tropical troposphere and there is further uncertainty introduced by the fact that it may be the QBO temperature signal in the lower stratosphere that provides the main physical effect on the troposphere (see 2.1 above), and different measures of the QBO winds have been chosen to provide a representation of the temperature signal. More recently (e.g. Gray et al. 2018) it has become customary to quantify the state of the QBO by the coefficients of the two dominant principal components describing the height and time variation of equatorial winds (Wallace et al. 1993).

The possibility of a QBO effect on the extratropical troposphere was first suggested by Ebdon (1975) and has now been demonstrated more clearly by careful statistical work with a longer data record (e.g. Coughlin and Tung 2001, Thompson et al. 2002). The Extratropical Pathway discussed above provides a plausible mechanism for such an effect, with the equatorial QBO affecting the extratropical stratosphere and then being communicated downwards to the extratropical troposphere. The observed QBO signal in the NH extratropical stratosphere is clear only in the winter (see e.g. Fig 3 of Anstey and Shepherd 2014) and correspondingly any NH tropospheric QBO signal resulting from the Extratropical Pathway is expected to be confined to the winter. In the SH extratropical stratosphere any QBO signal seems to be confined to the late spring/early winter period of transition to summer easterlies and the Extratropical Pathway to the troposphere is therefore likely to be relevant to communication of a QBO signal primarily during this season.

The QBO signal in the extratropical troposphere is regarded as providing strong evidence for an effect of the stratosphere on the troposphere, i.e. for coupling from the stratosphere to the troposphere, because the basic ingredient of the QBO, the oscillation in tropical stratospheric winds, may be regarded, at leading order, as externally imposed on the extratropical circulation. Of course this is only a leading-order view and over the years different aspects of the possible effects of the extratropical circulation on the tropical QBO have been suggested and investigated. These have included effects on seasonal modulation of the QBO (Kinnersley and Pawson 1996, Hampson and Haynes 2004) and, very recently, demonstration that waves propagating from the extratropics played an important role in the unexpected QBO disruption in 2015/16 (Newman et al. 2016, Osprey et al. 2016).

Correspondingly if there is a signal of the QBO (as defined by stratospheric winds) in the tropical troposphere the view is taken here that this may be regarded as evidence for coupling from the stratosphere to the tropical troposphere. Justification for this view is that there is no suggestion from basic dynamical theory or from modelling studies that the stratospheric QBO requires organized variation on the same timescale in the troposphere. Indeed the basic mechanism, captured, for example by the simple model of Plumb (1977), is that stratospheric flow at any given level essentially varies as the time integral of the force due to dissipating waves, with that force varying in time through the effect of the flow at lower levels on the wave propagation and dissipation. However it has been suggested that the QBO is modulated by the El Nino/La Nina variation in the troposphere (Taguchi 2010) and such modulation has been reproduced in model studies (e.g. Kawatani et al 2019). So, again, a leading-order interpretation of a QBO signal in the tropical troposphere as evidence for stratospheric influence is justifiable, but care may be required in interpretation of details.

a. Annual and seasonal means

There are several papers, published over a period of 30 years, which have suggested or investigated the possibility of a QBO signal in seasonal or annual mean measures of the circulation in the tropical troposphere. Confidence in the reality of these signals has increased with the length of the QBO data record and, equally important, as data coverage across the tropics as a whole has improved; however for some quantities particular care is needed to remove the strong ENSO signal. (See further comment below.) The history of this work, including some new observational results, has been reviewed in a companion paper (Hitchman et al. 2021) to this review and the reader is referred to that paper for more detail. As noted in Section 2.1, there is a clear QBOE-QBOW signal in temperatures that extends down to the tropopause, with (corresponding to the vertical shear in the QBO winds) colder temperatures for QBOE relative to QBOW in the lower stratosphere. Within the stratosphere this QBO temperature signal is generally considered to be longitudinally independent at leading order. However, as with other dynamical features, the longitudinal variation becomes stronger as the tropopause is approached and appears to be modulated by regional variations in convection (Collimore et al. 2003). The current picture of the QBO signal in temperature at tropopause level (e.g. at 100hPa) is summarized by Hitchman et al. (2021, see e.g. Figs 17 and 18). The QBOE-QBOW difference at low latitudes is everywhere negative but, broadly speaking, largest in regions where convective activity is strongest, i.e. over South America, Africa and Indonesia, and shows significant seasonal variation. Alongside the colder tropical tropopause temperatures in QBOE relative to QBOW there is a corresponding increased frequency of tropical tropopause layer (TTL) cirrus (Davis et al. 2013, Tseng and Fu 2017, Son et al. 2017). As with temperatures, there is evidence of longitudinal variation in the difference, but the shorter data record for cirrus limits certainty on the detailed structure of that variation.

Within the troposphere itself QBO-related patterns have been found in different observational measures of tropical convective activity obtained from satellite datasets on outgoing long-wave radiation (OLR), precipitation and different types of cloud (Collimore et al. 2003, Liess and Geller 2012, Son et al. 2017, Gray et al. 2018, Lee et al. 2019). Some authors have made use of re-analysis data products alongside satellite data. These include upper tropospheric velocity potential (Liess and Geller 2012), precipitation estimates (Gray et al. 2018) and a range of convection/precipitation diagnostics (Lee et al. 2019). Whilst these products need to be treated with caution because of the possible effects of differences in model/analysis schemes, they are potentially a very useful way of combining information from a range of different data sources. The patterns identified in these papers are characterized by very strong longitudinal variation. It is difficult to be clear on the consistency between the patterns described in different papers, because different authors have used different measures of QBO phase and some authors (Collimore et al. 2003, Gray et al. 2018) have considered seasonal variation of any patterns, while others have not.

Considering first the annual averaged patterns, and taking QBOE and QBOW to be defined by the wind at 50 hPa, the common features that emerge are that convective activity (associated with larger values of precipitation and smaller values of OLR) in QBOE-QBOW is relatively enhanced in the tropical west Pacific, relatively suppressed in the equatorial central and east Pacific and enhanced in the annual average ITCZ region to the north of that and also in the corresponding ITCZ region in the Atlantic. This QBOE-QBOW pattern is illustrated in Figure 4 which shows the annual average of the monthly regression of precipitation onto minus the value of a QBO index based on winds at 50 hPa, i.e., this is the precipitation change associated with a one standard deviation decrease in QBO zonal wind. Precipitation data are from the Global Precipitation Climatology Project (GPCP; Adler et al., 2018). Results from Gray et al. (2018) and Lee et al. (2019) are consistent with those shown. The QBOE-QBOW pattern has been described as a strengthening of the Walker circulation, i.e. in the west-east difference in convective activity in the tropical Pacific, together with a westward shift across the tropical Pacific of the local Hadley circulation[2]. If the QBO is defined by the wind at 70 hPa or below (Liess and Geller 2012, Gray et al. 2018) then the patterns appear to be a little different, with reduced precipitation along the northern flank of the Maritime Continent and enhanced precipitation to the east of that, and with a difference in the central and eastern Pacific that is more a northward shift of ITCZ precipitation rather than an enhancement.

It should be noted that any identification of a QBO signal in the tropical troposphere is subject to statistical uncertainty, and indeed some studies of some quantities that are

[2] 'local Hadley circulation' is used to mean the local circulation in the meridional (latitude-height) plane, to be distinguished from the zonal mean meridional circulation.

potentially relevant, e.g. lightning (Dowdy 2016), have found no significant QBO signal. A particular difficulty is that any QBO signal has to be distinguished from the very strong El Nino signal. This has been addressed in various ways. For example Liess and Geller (2012) carefully tested the effect of excluding El Nino or La Nina years by different criteria, Gray et al. (2018) considered regression against a set of indices including QBO and ENSO as well as simple QBOE-QBOW differences. The patterns shown in Figure 4 have been calculated by regressing year-by-year time series of precipitation for each calendar month against the Nino3.4 index and then extracting the regression signal. (See Figure caption for further details.) Only very small parts of the patterns shown in Figure 4 can be justified as statistically significant at the 5% level and the test applied has not accounted for spatial correlations which reduce the effective degrees of freedom; nonetheless they are presented here, subject to that uncertainty, as a basis for further consideration and discussion.

Turning to the seasonal variation, any influence of the QBO is likely to be modulated by the strong climatological seasonal variation in the pattern of precipitation and related quantities (see e.g. Figure 1 of Lee et al. 2019). An interesting initial indication of seasonal differences was reported by Collimore et al. (2003) who found an opposite signed longitudinal QBOE-QBOW pattern in NH summer relative to NH winter with convective activity weaker in the west Pacific and stronger in the east Pacific. Gray et al. (2018), using a longer data record, showed QBOE-QBOW differences in precipitation to the north of the Maritime Continent that are strongest in NH summer (though present in all seasons). The calculations used to generate Figure 4 showed strong differences between the QBOE-QBOW patterns in NH summer and those in other seasons. However all these possible seasonal variations in QBOE-QBOW differences are subject to the increased statistical uncertainty that results from reduction in the effective length of the available time series due to decomposition by season.

b. Madden-Julian Oscillation and other intraseasonal and higher-frequency variability

The MJO is a major feature of tropical tropospheric variability on subseasonal timescales (e.g. Zhang 2005). A possible QBO modulation of the MJO was suggested many years ago (Kuma 1990), on the basis of analysis of upper tropospheric winds in radiosonde data. Interest in this topic has revived recently through the work of Yoo and Son (2016) and Son et al. (2017) who demonstrated a strong QBO signal in the NH winter (or SH summer) MJO, with the difference between QBOE and QBOW accounting for more than 50% of the interannual variance of NH winter MJO activity over 35 years (1979 to 2015). The MJO is larger amplitude and more persistent when the QBO wind in the lower stratosphere is easterly and smaller amplitude and less persistent when it is westerly. This work was based primarily on OLR-based measures of the MJO, but a similar signal is detected (Marshall et al. 2017, see in particular their Figure 6) with the RMM (Real-time Multivariate MJO) indices (Wheeler and Hendon 2004) that are dominated by the zonal wind component of the MJO. Again this signal is strong only in NH winter and is negligible in other seasons.

More geographical detail is given in Figure 5 taken from Son et al. (2017), which shows the climatological seasonal average NH winter distribution of low latitude OLR and its intraseasonal variance, and the corresponding El Nino-La Nina and QBOE-QBOW differences. The QBOE-QBOW signal in the seasonal average (Figure 5c) is consistent with the precipitation signal shown in Figure 4, with regions of negative OLR anomalies broadly corresponding to regions of positive precipitation anomalies, however it is weak compared to the El Nino – La Nina signal (Figure 5b). The typical magnitude of the QBOE-QBOW signal in the intraseasonal variance (Figure 5f), on the other hand, is of similar magnitude to that in the corresponding El Nino – La Nina signal (Figure 5e). The QBOE-QBOW signal is largely confined to the central and eastern Indian Ocean, the maritime continent and the western Pacific and to a narrow latitudinal band to the south of the equator. The El Nino – La Nina signal, on the other hand, is localized further to the east. Nishimoto and Yoden (2016) demonstrated a corresponding difference in spatial structure of MJO-associated convection. Zhang and Zhang (2018) examined further the MJO-QBO connection and argue that the MJO signal in QBOE is stronger in part because the MJO is active for a larger fraction of time.

They argued that this results from a longer duration of individual MJO events and in particular that in QBOE more MJO events propagate beyond the Maritime Continent into the West Pacific. The characterization of the MJO as active for a larger fraction of time requires a quantitative criterion which in Zhang and Zhang (2018) was chosen to be a threshold RMM amplitude. This tacitly neglected any changes associated with MJO events below threshold amplitude. On the other hand Lim et al. (2019) showed that the probability distribution of daily MJO amplitudes is shifted to higher amplitudes during QBOE across amplitudes from the smallest to the largest, suggesting there is a QBO effect regardless of MJO amplitude. Son et al. (2017) provided evidence that the QBO-MJO connection was strongest when winds at 50hPa were used to define the QBO phase and much of the work mentioned above has followed this, however Densmore et al. (2019) suggest on the basis of the principal component approach to defining the QBO that winds in the 20-50hPa layer give the strongest signal.

Hendon and Abhik (2018) presented a more detailed analysis of the significant difference in the structure and magnitude of the MJO temperature anomalies in the upper troposphere and lower stratosphere between QBOE and QBOW and suggested that these upper level differences were an important part of the mechanism for the enhancement of the MJO under QBOE. Sakaeda et al. (2020) demonstrated further that there is an increase of MJO high cloud fraction during QBO easterlies and a consequent strengthening of cloud-radiative feedback, as measured by the correlation between precipitation and OLR, which might be expected to enhance MJO activity (Adames and Kim 2016).

Abhik et al. (2019) and Sakaeda et al. (2020) recently investigated QBOE-QBOW differences across the many different components of temporal variability in the tropical troposphere. Sakaeda et al. (2020) concluded that there was no significant modulation by the QBO of convectively coupled equatorial Kelvin waves, Rossby waves, mixed Rossby-gravity waves and gravity waves (at least down to a period of 2 days) and Abhik et al. (2019) came to largely the same conclusion regarding all high-frequency (2–30-day period) variance and the non-MJO component of the intraseasonal (30–120-day period) convective variance. Abhik et al. (2019) argued that the unique sensitivity of the NH winter MJO might be due to the MJO vertical structure (deep and upright) as compared to other convectively coupled equatorial waves together with the very cold tropopause temperatures, across the Maritime Continent in particular, in NH winter.

Klotzbach et al. (2019) and Sakaeda et al. (2020) have presented evidence that the MJO-QBO connection as described above has emerged only since the early 1980s. Their analysis, notwithstanding some uncertainty in quantifying MJO activity in the pre-satellite era (i.e. pre-

1979), shows no discernible correlation between the QBO and the MJO strength during the 1950s to 1970s (a period when QBO wind measurements were available) and suggests that this was also true in the 1900s-1950s period (when there were no direct QBO measurements, but for which an estimated QBO time series is available, constructed from extratropical surface pressure measurements).

c. Tropical cyclones

Gray (1984) suggested a statistical connection between the QBO and Atlantic hurricane frequency, with a correlation coefficient r~0.4 between occurrence of QBOW at 30 hPa in a given year and the number of hurricanes in that year, significant at the 5% level. Camargo and Sobel (2010) later showed that neither this relation nor a relation based on a different QBO level holds when a longer data record is considered. They noted that this might be because the apparent earlier connection was a statistical fluke, or because a multidecadal change in the background state of the atmosphere has meant that the physical mechanism leading to the connection no longer operates so effectively, though they were ultimately unable to identify any specific change of this type. There has also been interest in possible connections between the QBO and other aspects of tropical cyclone behavior, such as tracks, though quantifying the statistical significance of any signal is not straightforward. For the

Western Pacific Ho et al. (2009) presented evidence of a connection between QBO phase and the tracks (not the frequency or intensity) of the tropical cyclones. Fadnavis et al. (2014) found a dependence of cyclones in the Bay of Bengal on the QBO, with cyclones occuring more often during QBOE conditions and changing their tracks depending on the QBO, moving westward and northwestward during QBOE and northward/northeastward during QBOW. Distinct from the above studies, which considered characteristics of observed cyclones, there has been consideration of ‘potential intensity’, which is a theoretical predictor of tropical cyclone intensity based on large-scale dynamic and thermodynamics variables. See Section 3.3 below for further details.

d. Monsoons

Another suggested QBO effect is on the Indian Summer Monsoon (ISM). Given the importance to human society of the latter it is not surprising that the possibility of using such an effect to aid prediction has received significant attention. Connections between the QBO and the ISM have been suggested by several authors including e.g. Mukherjee et al. (1985), Bhalme (1987) and Madhu (2014), though clear simple connections supported by strong statistical evidence have been hard to find. However Claud and Terray (2007) suggested that whilst the connection is weak in June-July it may be stronger, and potentially practically useful, in August-September.

e. Subtropics

Given the dynamical connections between subtropics and tropics, the QBO signal in the subtropics is briefly considered. Many studies based on re-analysis data have shown a QBO signal in the zonally averaged subtropical zonal winds (e.g. Crooks and Gray 2005, Inoue et al. 2011, Anstey and Shepherd 2014, Bronnimann et al. 2016, Gray et al. 2018), with the QBOE-QBOW (based on the lower stratosphere) signal broadly corresponding to a poleward shift of the subtropical jet. The signal is deeper than the subtropical jet itself and the latitudinal structure and magnitude vary significantly with season. There does not seem to have been any systematic study of seasonal variation (the results shown in the papers cited are either annual averages or are else shown for one or two selected seasons), though Gray et al. (2018) showed monthly variation from November to March. The most detailed studies have been provided by Inoue et al. (2011) and Inoue and Takahashi (2013), with the latter emphasizing the longitudinal structure in the QBO signal and focusing on the Asian region in northern autumn. Seo et al. (2013) showed, consistent with the results cited above for the zonally averaged flow, that there is a significant QBO signal in the latitude of the East Asian

Jet in northern spring and a corresponding signal in rainfall in the western North Pacific region (including in parts of China, Japan and Korea). Garfinkel and Hartmann (2011) identified a poleward shift in the NH winter subtropical jet in the Pacific sector in QBOE and an equatorward shift in QBOW and noted that the signal in the Atlantic sector is distinctly different. Similar features were noted by Wang et al. (2018a) who further discussed the implications for the storm tracks. None of the above studies have argued that the effect of the QBO on the subtropical jet has a significant influence on the tropical troposphere, but such an influence would be an example of the operation of the Subtropical Pathway.

3.2 Influence of Sudden Stratospheric Warmings and other extratropical stratospheric dynamics on the tropical troposphere

The wintertime stratospheric polar vortex, particularly in the NH, is intermittently disrupted through upward propagation of planetary-scale Rossby waves from the troposphere. The strongest such disruptions are known as Sudden Stratospheric Warmings (SSWs) (e.g. Butler et al. 2017). The dynamical effects of such mid-/high-latitude disruption, some associated with SSWs, some with dynamical disturbances that do not meet the criteria for SSWs, also extend horizontally within the stratosphere into the tropics and indeed into the opposite hemisphere, including into the tropical lower stratosphere (Dunkerton et al. 1981, Randel 1993, Taguchi 2011, Gomez-Escobar et al. 2014) where they lead to cooling. Li and Thompson (2013) have shown that these dynamically driven temperature variations in the tropical lower stratosphere are correlated with variations in tropopause level cloudiness and suggest this as a possible pathway for the influence of the stratosphere on the climate of the tropical troposphere.

A series of papers by Kodera and collaborators (e.g. Kodera 2006, Eguchi and Kodera 2007, 2010, Kodera et al. 2011a, 2015) have argued that significant effects of SSW-driven tropical lower stratospheric cooling extending downward into the tropical troposphere, lasting a period of two weeks or more, may be identified in observations. The identified effects vary from event to event, but for NH winter SSWs are typically associated with suppressed convection in the equatorial NH (i.e. the winter hemisphere) and enhanced convection in the equatorial SH (i.e. the summer hemisphere), manifested by changes in OLR and precipitation, and regional increases in high-level cloudiness. Bal et al. (2017) noted that this SH-enhancement/NH-suppression of convection is particularly strong for vortex-split SSWs. To the extent that the SH-enhancement/NH-suppression corresponds to enhancement of the geographical distribution of precipitation this signature has similarities with the QBOE signal in precipitation, also associated with cold temperatures in the tropical lower stratosphere, described in Section 3.1a. However it should be noted that the dynamically driven temperature anomaly associated with an SSW typically extends across a broad low-latitude region (~40S-40N) whereas the primary QBO temperature anomaly is much narrower (~15S-15N) (e.g. Randel and Wu 2015) and this might imply a significant difference between the two responses, for example the latitudinal width of the SSW signal might allow a more direct effect on convection and precipitation associated with Hadley Cell upwelling in the summer hemisphere. Eguchi and Kodera (2007) reported a study of tropical tropospheric changes accompanying the unusual SH SSW of September 2002. Cooling of the tropical lower stratosphere was apparent for 10 days or so after the high-latitude warming and was accompanied by changes in several different observational indicators of the tropical tropospheric circulation and convective activity. Other studies, including Kuroda (2008) and Kodera et al. (2017) have identified tropical tropospheric changes accompanying other types of dynamical events in the stratosphere such as 'vortex intensification' (VI) events, and with a strengthening of the upper-stratospheric subtropical jet, As with SSWs these events have a clear and well-understood effect on temperatures in the lower stratosphere.

The difficulty with these observational case studies (even when several events of the same type are considered) is in drawing confident conclusions that changes in tropical tropospheric circulation and convective activity are caused by stratospheric dynamical events, rather than simply being a manifestation of large week-to-week internal variability. A recent modelling study by Noguchi et al. (2020) that focuses on the strongly disturbed SH vortex of September 2019, gives more certainty over cause-and-effect, at least for that particular event. That work is discussed in Section 4.1b below and some results are shown in Figure 8.

Whilst the above has emphasised coupling of dynamical variability in the extratropical winter stratosphere to the tropical troposphere via the Tropical Pathway, other mechanisms are also possible. For example, (recall Section 2.2) Kuroda (2008) identified propagation of a dynamical signal from mid-latitudes to low latitudes within the troposphere as important in the later stages of SSW or VI events.

A different aspect of possible effects on the tropical troposphere associated with the dynamical changes in the stratosphere was provided by Sridharan and Sathiskumar (2011) who noted a significant increase in convection (indicated by decreased OLR) in the Maritime Continent region in the early stages of evolution towards an SSW and argued that this was associated with tropopause-level PV intrusions at similar longitudes. Such subtropical PV intrusions, manifested by equatorward extension of filaments with stratospheric PV values into the tropical upper troposphere, have a recognized connection with tropical convection (e.g. Kiladis 1998, Kiladis and Weickmann 1992) and therefore offer a potential route for SSWs to affect such convection. The association between SSWs and subtropical PV intrusions has been more widely demonstrated by Albers et al. (2016), who are cautious about assigning a causal relationship, but suggest that the mid-stratospheric distortion of the large-scale PV field associated with the SSW may, through the vertically non-local PV inversion operator, have a direct effect on the lower level circulation which favours the formation of intrusions. This possible effect of SSWs on the tropical troposphere via subtropical PV intrusions operates via the Subtropical Pathway shown in Fig.1b.

3.3 Influence of recent tropical stratospheric temperature trends on tropical cyclones.

Understanding the cause of observed recent trends in tropical cyclone intensity and projecting how tropical cyclone activity will differ under climate change is a topic of great interest and importance. Future projections indicate that anthropogenic warming will cause the globally averaged intensity of tropical cyclones to increase, shifting toward stronger storms (Knutson et al., 2010, and references therein). There is some evidence that tropical cyclone intensity has already changed, such as an increase in the estimated energy dissipated by tropical cyclones (Emanuel, 2005) and an increase in the intensities of the strongest tropical cyclones (Elsner et al., 2008; Kossin et al., 2013). Much of the previous work investigating the physical causes of these changes has focused on the sea surface temperature (either directly or indirectly), but several recent papers have addressed the role of upper tropospheric and lower stratospheric temperature changes in contributing to changes in tropical cyclone intensity.

Part of this work considers the 'potential intensity', defined as the square of the predicted maximum surface wind speed V_p,. The hurricane model of Emanuel (1986) and further developments of that model (see in particular Bister and Emanuel 2002) give the explicit prediction $V_p^2 = (C_k/C_D)\ (T_s/T_o - 1)\ (h^*_0 - h^*)$, where C_k is the non-dimensional surface exchange coefficient for enthalpy, C_D is the drag coefficient, T_s is the sea surface temperature, T_o is the 'outflow temperature', h^*_0 is the saturation moist static energy at the sea surface and h^* is the saturation moist static enthalpy in the troposphere. Each of the quantities appearing in this expression can be estimated from a combination of different atmospheric observations. Emanuel et al. (2013) (see also Wing et al. 2015) argued that there has been a systematic increase in potential intensity in the Atlantic region since 1990 (see Figure 6 for details) and concluded that a major part of this is due to a decrease in the outflow temperature, i.e. the temperature at tropopause level or in the lower stratosphere. (Some but not all of the datasets they considered, three from re-analysis and one from radiosondes, supported this conclusion.) More recent papers have debated this topic, including whether tropopause temperatures are the most relevant aspect of the temperature structure (Vecchi et al. 2013, Ferrara et al. 2017) or using satellite brightness temperatures of tropical cyclone outflow as an alternative to reanalysis temperatures (Kossin 2015) to conclude that there is no identifiable recent global trend in potential intensity.

4. Numerical model studies/mechanisms

Models, with a range of sophistication and complexity up to and including state-of-the-art climate models, have played an important role in research on extratropical stratosphere-troposphere coupling. A first important step was simply to establish that relationships between stratosphere and troposphere, indicated by time evolution of correlations for example, were causal. The lagged correlation between the tropospheric flow and the stratospheric flow 10-20 days earlier, for example, found by Baldwin and Dunkerton (2001), could imply a downward 'phase propagation' without any downward propagation of information (Plumb and Semeniuk 2002). But subsequent numerical model studies clearly demonstrated that artificially imposed changes in the stratosphere can have a significant tropospheric effect (Polvani and Kushner 2002, Gillett and Thompson 2003, Norton 2003, Kidston et al. 2015 and references therein). Model studies have also been used to good effect in clarifying the importance of different mechanisms for extratropical stratosphere-troposphere coupling (e.g. Kushner and Polvani 2004, Song and Robinson 2004, Hitchcock and Simpson 2016).

The response of deep convective systems in the tropical troposphere to perturbations originating in the stratosphere, particularly (see Fig. 1) via the Tropical Pathway but also by the Subtropical Pathway, is likely to be of major importance to tropical stratosphere-troposphere coupling. As noted above, it has been suggested in several previous studies that deep convective systems are sensitive to conditions in the tropical lower stratosphere. Perhaps the most concrete model which suggests, and potentially quantifies, sensitivity of tropical tropospheric circulations to upper level conditions is the hurricane model of Emanuel (1986) and its subsequent developments (e.g. Bister and Emanuel 2002) which, as noted in Section 3.3, give an explicit prediction of dependence of maximum surface wind speed V_p (and hence of other quantities such as minimum surface pressure) on tropopause temperature, which in many cases can be taken to be tropopause temperature. This model is the basis for the suggested effect of stratosphere-coupling on tropical cyclones in particular (recall Sections 3.1c, 3.3 above and see Section 4.2c below) but is often cited (e.g. by Liess and Geller 2012) as suggesting more general sensitivity of tropical circulations to upper level conditions. However this model relies very strongly on the coherent organization of dynamical and physical processes that is particular to tropical cyclones and its more general relevance, even in a qualititative sense, is not clear.

It is highly plausible that tropical circulations respond within the uppermost part of troposphere to externally imposed changes within the TTL or the tropical lower stratosphere. These responses might include the height to which deep convection penetrates, or in the amount of high-level cirrus (as noted in association with the QBO in Section 3.1a). However such upper-level responses do not by themselves necessarily imply a response that penetrates sufficiently deep into the troposphere to account, for example, for a significant change in precipitation. The interesting General Circulation Model (GCM) study by Thuburn and Craig (2000) in which a change in tropical lower stratospheric temperatures was imposed artificially noted an effect on convective heating that extended down to 12-13km, but the robustness of the effect or the mechanisms operating were not explored.

The remainder of this Section surveys the model studies that have been used to argue for, or to investigate possible mechanisms for, stratosphere-troposphere coupling in the tropics, including, in particular, those that might lead to effects extending through the depth of the troposphere. The survey is divided into two parts. The first (4.1) focuses on global models, which include free-running GCMs (the term GCM will be used only if the model is being used in a free-running mode), seasonal forecast models for which specific initial conditions are important and models that incorporate artificial nudging to constrain the circulation in certain regions. A common feature of these models is that all have convective parametrizations. The second part (4.2) of this section focuses on ‘regional’ models that, in contrast, are convection-resolving (or 'convection-permitting').

4.1 Global model studies

a. Global model studies on the QBO influence on the tropical troposphere

GCM studies of the effect of the QBO on the extratropical stratosphere and on the troposphere were first reported by Balachandran and Rind (1995) and Rind and Balachandran (1995). Successful GCM simulation of the QBO itself was at that time only just beginning (Takahashi 1996). However many early GCM studies of the wider effect of the QBO circumvented this problem by adding an artificial forcing of some kind on the tropical stratosphere, typically to force the model winds in this region to be either QBOE-like or QBOW-like and this was the approach taken in the Balachandran and Rind (1995) and Rind and Balachandran (1995) papers. They identified a relatively stronger Hadley circulation and increased tropical cloud cover in QBOE vs QBOW, but did not find any evidence of significant differences in the longitudinal structure. Interpretation of quantitative aspects of their results needs to take into account that the corresponding QBOE vs QBOW temperature difference in the tropical upper troposphere and lower stratosphere, whilst having the sign expected (cold in QBOE vs QBOW) penetrated further into the troposphere than appears to be the case in observations.

Giorgetta et al. (1999) subsequently demonstrated a QBO effect on the NH summer tropics by imposing different wind profiles in the model stratosphere and identifying a resulting signal in the troposphere (see Figure 7). The QBOE-QBOW signal was increased convective activity in a low-latitude band over the west Pacific and decreased convective activity to the north and south and to the east (over India), indicated by the signal in latent heating shown in the upper panel of Figure 7, There was increased upper tropospheric cloudiness in QBOE-QBOW over large regions of the tropics, but particularly co-located with regions of increased precipitation. Giorgetta et al. (1999) argued that geographical variation of the QBOE-QBOW signal in convective activity was caused the positive feedback effect of regional changes in cloud radiative forcing (lower panel of Figure 7), which was strongest where convection was deepest. Garfinkel and Hartmann (2011), as part of a broader study of the effect of the QBO on the troposphere, showed that for NH winter imposed QBOE conditions in the lower stratosphere led to increased convection in the tropical central Pacific and a larger region of increased high cloudiness, as measured by OLR. Since the Giorgetta et al. (1999) study is for NH summer conditions and the Garfinkel and Hartmann (2011) study is for NH winter conditions one would expect to find differences between their results. Certainly both show strong regional variation of the change in precipitation, consistent with a modulation of the Hadley and Walker circulations. Both also show to some extent that in QBOE convection is enhanced over the West Pacific region where convection is most active in the control state and in that sense are consistent with observed QBO differences shown in Figs. 4 and 5.

As noted previously, one of the most interesting suggested effects of the QBO is its modulation of the MJO. The possible connection between the QBO and the MJO was investigated in a GCM by Lee and Klingaman (2018). Whilst the model used, the UK Met Office Unified Model with a global ocean mixed layer, simulates to some extent both MJO and QBO, the QBO-MJO connection found in the model does not resemble that found in observations (see Section 3.1b). There is no significant correlation between the QBO phase and MJO amplitude and whilst there is some correlation between QBO phase and MJO activity in different geographical regions, this does not match that seen in observations. Lee and Klingaman (2018) noted that the lower stratospheric temperature differences between different QBO phases are significantly smaller in the model than in observations and have a different longitudinal structure. They also noted that GCM representations of the MJO often have significant differences in vertical structure from observations and the MJO simulation in this particular model exhibits other typical deficiencies including amplitude that is too weak, particularly to the east of the Maritime Continent. Any of these factors might diminish or otherwise alter the effect of the QBO on the MJO. More recent studies have sought QBO-MJO connections across wider sets of models. Lim and Son (2020) examined the four CMIP5 models with a realistic internally generated QBO and found that three substantially underpredicted MJO activity and the fourth did not show a robust QBO-MJO connection. Kim et al. (2020a) examined a much larger set of CMIP6 models and found that none exhibit the observed QBO-MJO connection. Both these studies noted that simulated QBO velocity and temperature anomalies in the lower stratosphere are generally weak relative to observations.

An alternative approach to examining the impact of the QBO on the MJO is to use seasonal forecast models initialised with observations. This ensures that the representation of the QBO and the MJO is realistic at least in the early stages of the simulation. Studies of this type can potentially give important information on relevant mechanisms as well as on specific implications for seasonal forecasting, Marshall et al. (2017) demonstrated using a global seasonal prediction model that in the NH winter season there is improved predictive skill for the MJO under QBOE conditions relative to QBOW for lead times of 5-30 days. This is an important demonstration, particularly in the current situation where no recognisable QBO-MJO connection can be reproduced in a free-running GCM. Marshall et al. (2017) further showed that this improvement does not simply stem from stronger MJO in initial conditions during QBOE, because the enhanced skill occurred for similar initial amplitude MJO events in both QBOE and QBOW.

The general result of enhanced predictive skill of the MJO during QBOE was confirmed by Lim et al. (2019) using models participating in the WCRP/WWRP subseasonal-to-seasonal (S2S) prediction project (Vitart et al. 2017). They too showed that the increase in skill was present over a range of initial MJO amplitudes. Kim et al. (2019), using a somewhat different set of models, also found enhanced skill during QBOE, but concluded that for most models the difference in skill is not statistically significant. However the Kim et al. (2019) conclusion might be affected by their consideration only of MJO with large initial amplitude (greater than 1.5 by the standard RMM measure). Abhik and Hendon (2019), who demonstrated a systematic difference in MJO forecast skill between QBOE and QBOW in two different models, also considered the simulated difference in vertical structure of the MJO at the tropopause between the QBOE and QBOW simulations and showed that these differences were consistent with those reported in observations by Hendon and Abhik (2018).

These seasonal forecast model studies have provided some important information on possible mechanisms for QBOE-QBOW differences in MJO evolution. Marshall et al. (2017) noted that the model used had low top and that the QBO signal in the lower stratosphere degrades during the simulation, losing more than half of its amplitude by day 30. This hints at the possibility that sustained representation of the QBO within the simulation is not important for the difference in the forecast evolution and indeed this is the conclusion reached by Kim et al. (2019), on the basis of comparison between high- and low-top versions of a particular model. Further support for this conclusion has come from the work of Martin et al. (2020) who considered seasonal forecast simulations in which for each initial condition defined by observations, additional simulations were performed where the initial condition in the troposphere was retained but that in the stratosphere was adjusted to either QBOE or QBOW. The finding was that whilst there was some evidence of an effect of the adjusted stratosphere, the dominant effect on QBOE-QBOW difference in simulated MJO evolution was determined by whether the tropospheric initial conditions were taken from QBOE or QBOW years.

A further very recent study that strictly speaking falls into the convection-resolving model category to be discussed in Section 4.2, but is very similar in spirit and methodology to the seasonal forecast studies reported above, is that by Back et al. (2020). This uses the WRF mesoscale model at a ‘convection-permitting’ resolution, on a limited geographical domain, with initial conditions and lateral boundary conditions specified by re-analysis data. A QBO-like perturbation is applied to a baseline MJO simulation via the initial and boundary conditions and some evidence of a QBO effect on the MJO is demonstrated.

b. Global model studies of SSW influence on the tropical troposphere

The effect of SSWs on the tropical troposphere proposed by Kodera and collaborators has been studied using model simulations reported in Kodera et al. (2011b). The technique used exploited previous modelling studies of SSWs (Mukougawa et al. 2005, Mukougawa et al. 2007) in which adding a certain set of predominantly high-latitude tropospheric anomalies to the initial conditions was shown to lead to SSWs. This allowed Kodera et al. (2011b) to generate SSW and non-SSW ensembles, each with 13 members, and to compare the tropical tropospheric evolution averaged over each of the ensembles. They noted statistically significant differences in the latitudinal structure of tropical precipitation between the two ensembles. During the early stages of development of the SSW, prior to a strong change at high latitudes, there is enhanced precipitation in the NH subtropics in the SSW ensemble. Then after the SSW there is enhanced precipitation in the SH tropics and suppressed precipitation in the NH tropics. Kodera et al. (2011b) interpreted the first stage as an effect of anomalous wave propagation within the troposphere and the second as an effect of cooling in the tropical lower stratosphere (recall Section 3.2). In establishing a difference between the SSW and non-SSW ensembles this study provided strong evidence of a genuine SSW effect.

Recent work is making further progress towards establishing reproducibility and examining cause-and-effect in more detail. Noguchi et al. (2020) have studied the evolution of the tropical troposphere in September 2019, when there was a significant SSW in the SH (which did not quite reach the standard criterion of a 'major' warming). They used an ensemble forecast approach in which a control ensemble was freely evolving and a nudged ensemble was constrained to the observed stratospheric evolution, following the approach of Hitchcock and Simpson (2014). Selected results from the Noguchi et al. (2020) paper are shown in Figure 8 and provide a clear picture of the co-evolution of different quantities, averaged across the simulation ensemble, as the SSW proceeded. Figure 8(a) shows the evolution of the actual high-latitude 10 hPa temperature in September 2019 together with the corresponding evolution in the freely evolving control ensemble and the nudged ensemble. Figures 8(b) and 8(c) show the differences between nudged and control ensembles in, respectively, tropical temperatures and tropical convective heating. Figures 8(d) and 8(e) show corresponding differences in meridional circulation, which are present both in stratosphere and troposphere. Figures 8(f) and 8(g) show differences in tropical precipitation. These results demonstrate that nudging towards the stratospheric evolution associated with the SH SSW has a systematic effect on the tropical troposphere. For example, the ensemble average difference in precipitation over a South/South-East Asian region over a two-week period is about 70% of the corresponding standard deviation within each ensemble. Many of the tropical stratospheric features seen in Figure 8 are similar to those identified in the observational case studies reported in Section 3.2. On the other hand the probability distributions of precipitation in a particular tropical region shown in Figure 8(g), if the variability within ensembles represented by this model is realistic, emphasise the difficulty of drawing conclusions on systematic effects on tropical precipitation from individual case studies.

In a different study Yoshida (2019), using a large ensemble of numerical model simulations including 6117 model-generated SSW events, has demonstrated a statistically significant relationship between SSWs and tropical precipitation (zonally averaged) with enhanced precipitation over a few days prior to and coincident with SSWs and reduced precipitation over a few days after SSWs. Whilst the signal is weak, typically about 10% in various relevant metrics, there is a substantial increase (30%) in the probability of extreme tropical cyclone events during a 10-day period after SSWs.

The study of Noguchi et al. (2020) also reports variation in the response of the tropical troposphere to nudging when the model convective parametrization is changed. This is a further important consideration for any global model study of stratospheric influence on the tropical troposphere. Investigation in models that do not rely on convective parametrization is of course desirable, and a first such case is reported by Eguchi et al. (2015) who considered tropical tropospheric change following an SSW as simulated in a 60-day integration of the NICAM (Nonhydrostatic ICosahedral Atmospheric Model) global convection-permitting model. However, as the authors acknowledge, only one integration was carried out and no direct causal effect of the SSW on the troposphere could be deduced from this alone.

c. Coupled chemistry-climate model studies of long-term change

The radiative effects of water vapour and ozone in the tropical lower stratosphere are potentially important both in determining the temperature distribution in the tropopause region and the upper troposphere and in determining the radiative balance of the tropical troposphere as a whole (e.g. Forster and Shine 1997, Solomon et al. 2010). Annual and interannual variations of ozone and water vapour are also potentially important in radiative-dynamical effects in the tropopause region, e.g. in determining annual variation (Fueglistaler et al. 2011, Gilford and Solomon 2017, Ming et al. 2017) and interannual variability (Gilford et al. 2016) in temperatures.

Chemistry-climate models, in which ozone and related chemical species are predicted rather than being specified from climatology, as is the case for most climate models, have been used to demonstrate that changes in ozone can lead, for example, to significantly different climate sensitivity to increased greenhouse gases. Nowack et al. (2015), for example, demonstrated a 20% reduction in the change in surface temperature resulting from 4 x CO2 (quadruple concentration of atmospheric carbon dioxide compared to the pre-industrial level) in a model with interactive ozone relative to fixed ozone, though it should be noted that not all chemistry-climate models demonstrate a percentage reduction that is as large as this. (See further discussion in Marsh et al. 2016, Chiodo et al. 2018, Nowack et al. 2018.) Nowack et al. (2015) demonstrate that the reduction results from a succession of feedbacks; firstly a strengthened Brewer-Dobson circulation results in reduced lower stratospheric ozone, then the resulting reduction in long-wave heating reduces tropical lower stratospheric and tropopause temperatures, resulting in reduced water vapour concentrations in the lower stratosphere, and finally there is a reduced greenhouse effect from that change in stratospheric water vapour. The reduced greenhouse effect is partially cancelled by the radiative effect of increased upper tropospheric and tropopause level cloudiness.

Nowack et al. (2017) have noted the implications of these feedbacks for possible changes in El Nino under global warming. One commonly predicted response to increased greenhouse gases is that the Walker Circulation (and to some extent the Hadley Circulation) is weakened as a result of stabilization of the troposphere (e.g. Ma et al. 2018). There is in turn weakening of the typical eastward surface wind stress and hence, with a coupled ocean, weakening of the east-west surface temperature gradient in the Pacific, leading to an increase in the frequency of El Nino events (e.g. Bayr et al. 2014). The effects of interactive ozone described above imply, relative to the case of fixed ozone, a reduced increase in surface temperatures, hence reduced stabilization of the troposphere and reduced weakening of the Walker circulation. Nowack et al. (2017) demonstrate these effects in model simulations, as shown in Figure 9, and further demonstrate that the result is to reduce the increase in the frequency of El Nino events, particularly the frequency of extreme El Nino events, relative to that predicted by models that neglect the ozone feedback (i.e., at least until recently, a large proportion of the models used for climate prediction).

d. GCM studies of geoengineering effects

Injection into the stratosphere of aerosols or aerosol forming compounds that absorb incoming solar radiation, analogous to the effects of naturally occurring volcanic eruptions, is one of the most commonly considered geoengineering methods to reduce future climate change. However it could result in unintended consequences such as changes in regional circulation and hydroclimate, particularly in the tropics. Interesting examples have been given of possible volcanic or geoengineering effects on Sahel rainfall (Haywood et al. 2013) and on El Nino (Khodri et al. 2017). There are a variety of pathways whereby increased stratospheric aerosol loading can impact on the troposphere. Commonly, the influence of the radiative effect of the aerosols on the surface energy balance is considered as an important driver of precipitation responses to this kind of forcing. But another pathway by which precipitation responses could occur is through the warming of the tropical lower stratosphere that arises from the increased absorption of radiation by the excess aerosols. This pathway is omitted in model simulations that represent the effect of aerosol injection simply by reducing incoming radiation (‘solar dimming’) (e,g, Kravitz et al. 2014) and even in model simulations in which aerosol is explicitly included the role of this pathway may be overlooked.

Ferraro et al. (2014) demonstrated using an intermediate complexity GCM that increases in stratospheric sulfate aerosols cause a weakening of the tropical tropospheric circulation through upper tropospheric heating arising from longwave radiation emitted by the aerosol and by the warmer lower stratosphere. Using a-state-of-the-art Earth System model Simpson et al. (2019) have investigated the influence of the warming of the lower stratosphere under geoengineering in isolation by assessing comprehensive GCM simulations under the RCP8.5 scenario for greenhouse gas increase, with geoengineering aerosols, and then extracting the aerosol heating of the lower stratosphere and adding this alone to the baseline climate integrations. Broadly speaking, the conclusion is that the aerosol heating of the lower stratosphere tends to reduce the strength of the tropical circulation and hence reduce geographical contrasts in precipitation, with precipitation reducing in previously wet regions and increasing in previously dry regions. These conclusions are also potentially relevant to the effects of volcanic eruptions that reach the tropical stratosphere. It is well-established that such eruptions lead to warming of the tropical lower stratosphere (e.g. Fujiwara et al. 2015) and it has also been argued that they lead to changes in precipitation, in particular to the distribution of tropical precipitation (Iles et al. 2013). The changes in precipitation are, as has previously been the case for geoengineering effects, conventionally explained in terms of changes in surface energy budget, but the results reported above suggest that the effect of aerosol heating in the tropical lower stratosphere may be an important part of the mechanism.

The modelled effects of aerosol heating also show some consistency with the previously suggested effect of the QBO (with the warmer lower stratosphere due to aerosol heating corresponding to QBOW) although the warming of the tropical lower stratosphere in these experiments is considerably larger (~10 K at 20 km) than the QBOE-QBOW signal (~4 K at

20 km). Simpson et al. (2019) also briefly discussed a simple ‘aquaplanet’ experiment, with imposed localized regions of relatively high and relatively low SST in the tropics, and showed that imposed stratospheric heating again tends to reduce precipitation in wet regions and increase precipitation in dry regions. These results from a study motivated by geoengineering are a useful complement to, and show many common features with, those from the QBO-motivated studies discussed in Section 4.1a.

e. GCM studies of solar tidal effects

A final distinct example of a GCM study of tropical stratosphere-troposphere coupling is that by Sakazaki et al. (2017) and Sakazaki and Hamilton (2017) of atmospheric tidal influences on the diurnal cycle of tropical rainfall. The focus of this work is on the semidiurnal (S2) tide, which is well known to be significantly excited by ozone heating in the stratosphere. The cited papers examine in a realistic general circulation model the individual contributions of tropospheric and stratospheric forcing, by artificially suppressing different forcing mechanisms in different experiments. These experiments confirmed the significant role for stratospheric forcing, accounting for about half of the S2 amplitude in the tropical troposphere.

Sakazaki et al. (2017) further considered the effect of the tide on the semidiurnal variation in tropical rainfall. In the experiments where different parts of the tidal forcing are suppressed, reducing the tidal amplitude, it is found that the semidiurnal variation in rainfall is also reduced. This supports the argument that the semidiurnal tide is a major forcing mechanism for the semidiurnal variation in rainfall and implies that about half of this variation is due to stratospheric effects. Sakazaki et al. (2017) also noted that the amplitude of semidiurnal variation in rainfall (but not the amplitude of the semidiurnal tide itself as measured by pressure variation) is quite sensitive to the convective parametrization in the model and suggest that this sensitivity is potentially very useful for evaluation of convective schemes. The sensitivity presumably indicates that the physical mechanisms required to convert a specified tropospheric pressure perturbation to a variation in convection are captured by some parametrization schemes and not by others. Therefore this has general relevance to the problem of stratosphere-troposphere coupling, though it should be noted that the tidal perturbation is relatively high-frequency and the corresponding mechanisms that operate on weekly and longer timescales might be very different.

4.2 Regional/ CRM studies on the QBO influence on the tropical troposphere

a. Convection-resolving models

Any simulated change in the tropical troposphere in global models, including the response to changes in the stratosphere, will depend strongly on the parametrization of convection. The number of global model studies that have carefully studied stratosphere-troposphere coupling in the tropics (see 4.1a-e above) is small and it would therefore be highly desirable to extend these studies to a broader set of models (and hence a broader set of convective parametrizations).

A different approach is offered by simulations in convection-resolving models (CRMs), or more strictly ‘convection-permitting’ models, with non-hydrostatic dynamics, high horizontal resolution (less than a few km) and appropriate representation of microphysical and radiative processes. The focus in the following is on CRM simulations under idealised or simplified conditions such as small horizontal domains. See Back et al. (2020), mentioned in Section 4.1a above, and references therein for information on relevant studies in convection-permitting mesoscale models,

Nie and Sobel (2015) made a pioneering study of the effect on convection of lower stratospheric QBO-like temperature perturbations, i.e. of the Tropical Pathway and the associated tropospheric feedback mechanisms, using a convection-permitting model on a limited horizontal domain, an approach which is relatively well established in the tropospheric convection community. The horizontal domain is taken to be square, with periodic boundary conditions. A key point is that rather than setting the domain average vertical mass transport to be zero, the domain-average temperature is relaxed towards a specified environmental temperature profile and domain average vertical mass transport is then deduced. This is motivated by the weak temperature gradient (hereafter WTG) approximation, which assumes that in the tropics, where the Coriolis parameter is small, horizontal temperature gradients are maintained as weak by horizontally propagating gravity waves. The domain for the numerical simulation is therefore envisaged as a small part of a large-scale convecting region, within an environment of non-convecting regions in which the temperature profile varies only slowly in time. The fact that the domain average vertical mass transport is not zero implies that the domain contains a source of mass and indeed of other quantities such as moisture. These sources are justified as being provided by horizontal fluxes into the domain from the environment. Therefore the WTG approximation represents some of the effects of horizontal transport, i.e. some aspects of the interaction between convection and large-scale circulation. However it does not allow two-way interaction between the convecting region and the environment, nor between neighbouring convecting regions with different properties.

Nie and Sobel (2015) first carried out a sequence of QBO-neutral simulations in which the the sea-surface temperature (specified as spatially uniform) took a sequence of different values. These values were characterized by the difference ∆SST between the sea-surface temperature and that in a radiative-convective control simulation used to specify the environmental temperature. Each of these simulations evolved to a state with a non-zero vertical velocity, with the profile depending on the value of ∆SST. Further QBOE-like and QBOW-like simulations were then carried out in which the environmental temperature was perturbed at upper levels with a simple representation of QBO temperature variations. QBOE-like cold perturbations increased vertical motion in the upper troposphere and reduced it in the lower troposphere, described as a more 'top-heavy' vertical motion, and increased upper-level cloudiness. (The effect of QBOW-like warm perturbations was simply the reverse of this.) The precipitation response was more complicated, increasing at low values of ∆SST and reducing at higher values. Figure 10 shows some of the features of these responses. Nie and Sobel (2015) explained this by considering the budget of moist static energy, showing that for small values of ∆SST the main driver of changes in precipitation was the increased radiative heating due to change in cloudiness leading to an increase in precipitation, but that at larger values of ∆SST this increase was overwhelmed by the effect of the increase in 'gross moist stability' (GMS) (see e.g. Raymond et al. 2009), associated with the increased top-heaviness of the vertical motion, which acted to reduce the size of the precipitation response to the QBO-like temperature perturbations. Nie and Sobel (2015) concluded that their results suggest a more complex overall mechanism than simply ‘QBOE implies more active convection’.

A separate convection-resolving study of the QBO convection interaction was carried out by Yuan (2015). The first part of this study used a 3-D simulation in limited horizontal domain, similar to the Nie and Sobel (2015) approach, except that the WTG approximation was not used and therefore there was no domain averaged convergence or divergence of horizontal fluxes. The response was much weaker than that found by Nie and Sobel (2015) suggesting that the physical/dynamical processes allowed by the WTG approximation were indeed important. A second part of the Yuan (2015) study considered a much larger horizontal domain, with imposed horizontal gradients of sea-surface temperature driving a Walker-type circulation, but only one horizontal space dimension was included, i.e. the calculation was two-dimensional. This part of Yuan’s study demonstrated a substantial effect of an imposed upper level QBO temperature on the convecting regions in the Walker circulation, with QBOE-like perturbations leading to a reduction in precipitation in these regions (and a slight increase in neighbouring regions, so that the total precipitation remained roughly constant). Therefore, on the basis that the central convecting region corresponds to large ΔSST, these results and those of Nie and Sobel (2015) are consistent, though the decomposition of the response in precipitation was different, with Yuan identifying the decrease as due in part to a reduction in evaporation and a part to an increase in GMS. Yuan's results need to be treated with caution, because they may have been significantly affected by the two-dimensionality (e.g. Wang and Sobel 2011) but it is worth noting that the simulations contained not only the 'one-way' circulation-convection interaction allowed by the WTG approximation, but also potentially the 'two-way' interaction between different horizontal regions allowed by horizontal advection of moisture and by the spreading of high clouds.

Martin et al. (2019) have extended the Nie and Sobel (2015) work, within the same limited-domain modelling framework, to simulations of MJO variations. The latter are incorporated by using time-varying environmental temperature profiles and domain-average humidity sources (representing varying horizontal transport) based on observations from an international Indian Ocean field campaign in 2011-2012 (Yoneyama et al. 2013). Simulations of this type (e.g. Sentic et al. 2015, Wang et al. 2016) address the question of whether, if large-scale MJO-like variations are imposed, convection in limited horizontal regions evolves as observed and whether it evolves in such a way as to reinforce (or reduce) the specified MJO variations. Martin et al. (2019) incorporate QBO-like temperature perturbations and show that the convective response to the imposed large-scale MJO variations is enhanced under QBOE conditions, with, for example, larger vertical velocities, larger cloud fractions and reduced OLR during periods of active convection. Martin et al. (2019) also varied the height at which the QBO-like temperature anomaly is imposed and the response rapidly reduced when this height is increased. (Recall the vertical variation shown in Figure 2.) There was clear enhancement of precipitation under QBOE conditions when the height of the perturbation was lowest, but no significant change in precipitation when the height took any other value (including the value that is arguably closest to realistic).

The work summarised above investigated the effect of a QBO-like temperature perturbation, without any accompanying perturbation to the vertical shear. The chosen conditions for such simulations, with no systematic latitudinal variation and zero background numerical simulations, means that perturbations to temperature and to vertical shear can be applied independently. Martin et al. (2019) also reported results with an imposed QBO-like wind perturbation. No detectable response was found, suggesting that the 'wind shear' mechanism proposed as one of the ways in which tropical convection could respond to the QBO is of minor importance.

Another set of convection-permitting simulations which provide some insight into potential mechanisms for tropical stratosphere-coupling are the idealised simulations reported by Nishimoto et al. (2016) and Bui et al. (2017). These are two-dimensional with periodicity in the horizontal, contain a resolved stratosphere and assume zero Coriolis parameter. The simulations showed the development of a QBO-like oscillation of the stratospheric winds together with, coherent with this oscillation, significant variation in tropospheric winds and in the space-time organization of precipitation. Bui et al. (2019) have recently described three-dimensional simulations which show broadly similar behavior. The coherence of the tropospheric variations with the QBO-like oscillation is suggestive of significant effect of the stratosphere on the troposphere, but as in other similar problems, more examination is needed to establish causality. Such examination was provided by the Bui et al. (2017) paper, which studied the dynamics of the tropospheric variations in more detail, exploiting in particular sets of numerical experiments in which the evolution of the zonal wind was constrained in specified layers of the atmosphere. Figure 11 shows selected results from this paper. From the numerical experiments it was demonstrated first that the low-level tropospheric shear, which varies during the oscillation in the control run, plays an important role in determining the precipitation strength (Fig. 11 a,c: light precipitation, b,d: heavy precipitation). Therefore the coherent variation of the precipitation and the zonal winds does not imply stratospheric control of the former, even though the amplitude of the zonal wind oscillation is much larger in the stratosphere. However, when the low-level tropospheric zonal wind was constrained, the organisation of the precipitation was shown to vary coherently with the shear in the 8-10km layer (but not with the shear in higher layers). Bui et al. (2017)

argued that this demonstrates the realizability of the Gray et al. (1992a) shear mechanism. However it should also be noted the 8-10 km layer for which sensitivity to wind shear was well within the upper troposphere rather than being tropopause-level or lower stratospheric, even taking into account that the configuration of the Nishimoto et al. (2016) and Bui et al. (2017) simulations had a tropopause that was artificially low, at about 13 km, Therefore, whilst this is an important concrete demonstration of an effect of upper-level shear on convection and precipitation, direct relevance to observed QBO signals has not yet been demonstrated and there is no current inconsistency with the Martin et al. (2019) results discussed above.

b. Tropical cyclone models

There have been several model studies of the dependence of tropical cyclone characteristics on the environment and in particular on changes in tropopause temperature. As noted previously in Section 3.3, this is one of the factors that determines the potential intensity (V_p^2) which has been argued to be relevant to the actual intensity of tropical cyclones (Bister and Emanuel 2002). The studies have been based on models of varying complexity, some axisymmetric and some three-dimensional. Many recent studies have had high enough horizontal resolution to be convection-permitting. The two-dimensional study of Ramsay (2013) used a horizontal grid spacing of 2 km and considers the effect of changing stratospheric temperatures, finding that the simulated maximum surface wind speed V_s increased by 1 m s^{-1} for each 1 K decrease in stratospheric temperature. The predicted surface wind speed V_p calculated from environmental conditions varied similarly. The three-dimensional study of Wang et al. (2014), used an interior computational domain with 4km resolution, again with a relatively simple environment that was varied from one simulation to another. The environmental and initial conditions imply sensitivity of V_p to tropopause temperature in the range -(0.4-1) m s^{-1} K^{-1}. The simulations themselves showed V_s values significantly larger than V_p estimates, and Wang et al. (2014) discussed the reasons for this, but the sensitivity of V_s to tropopause temperature was about -0.4 m s^{-1} K^{-1} (i.e. at the low end of the range estimated for V_p). Whilst there is quantitative disagreement by a factor of two in the sensitivity of V_s between the two-dimensional simulations of Ramsay (2013) and the three-dimensional simulations of Wang et al. (2014) these two investigations together support firstly the physical relevance of potential intensity, i.e. V_p^2 as an estimate for V_s^2, and secondly the sensitivity of tropical cyclone intensity to tropopause temperatures that potential intensity suggests. However, as has been noted in Section 3.3., there is ongoing debate on this topic and the recent two-dimensional model study by Takemi and Yamasaki (2020) provides evidence that tropical cyclone intensity is more sensitive to troposphere lapse rate than to tropopause temperature.

Note that none of these studies have addressed the question of whether the tropical cyclone frequency, which was the property originally considered by Gray (1984), is affected by tropopause temperatures and indeed it is not clear how effectively this could be addressed in these types of studies, e.g. because the formation and development of tropical cyclones is determined in part by a combination of large-scale or synoptic-scale processes. Indeed a recent comprehensive study (Vecchi et al. 2019), considering predictions by global models at different resolution of changes in tropical cyclone frequency and intensity under greenhouse warming, noted that both changes in the frequency of synoptic-scale tropical cyclone 'seeds' and changes in the probability of intensification of those seeds are needed to explain overall changes in frequency.

5. Practical implications

5.1 Seasonal and subseasonal forecasting

The coupling between the stratosphere and the extratropical troposphere is now being exploited in seasonal forecasting (e.g. Fereday et al. 2012, Domeisen et al. 2019a,b) and is leading to revised practice in climate modelling (e.g. Scaife et al. 2012, Manzini et al. 2014). It is also being recognized as an important component of extratropical forecasting on subseasonal time scales (e.g. Domeisen et al. 2019a,b). So far most of the exploitation of stratospheric effects in seasonal forecasting has been focused on the NH winter over the North Atlantic region where a significant connection between the state of the stratosphere and the North Atlantic Oscillation has been discovered. However corresponding gains for seasonal forecasting in the SH spring, e.g. of the Southern Annular Mode, have also been demonstrated (e.g. Seviour et al. 2014, Hendon et al. 2020). If influence of the stratosphere on the tropical troposphere could be better understood and established as robust then, just as has been the case for the extratropics, there might be significant practical gains. The MJO, for example, is the dominant feature in tropical variability on subseasonal time scales and improved forecasting of the MJO would be relevant not only to forecasting high-impact tropical weather events such as tropical cyclones (Vitart 2009, Vitart et al. 2017), but also to subseasonal and longer term forecasting in the extratropics where an important part of the variability is driven by tropical rainfall anomalies (Manola et al. 2013, Scaife et al. 2017, Dias and Kiladis 2019).

As noted in Section 4.1a above, some information on the implications of coupling for subseasonal forecasting in the tropics has already been presented by Marshall et al. (2017) who considered the skill of subseasonal forecasts of the MJO using the Australian Bureau of Meteorology POAMA (Predictive Ocean Atmosphere Model for Australia) system. They showed that the forecast skill is greater in QBOE vs QBOW years, with the same level of skill being achieved 8 days later in QBOE vs QBOW. Lim et al. (2019) and Wang et al. (2019) have demonstrated similar conclusions from subsequent studies across larger sets of forecast models. Lim et al. (2019) noted also that in QBOW years reduced forecast skill corresponds in part to the failure to reproduce the reduced duration of MJO events, relative to QBOE, that is observed. Further study is ongoing, for example, as noted previously, Kim et al. (2019) have recently concluded that whilst several models show larger subseasonal prediction skill of the MJO in QBOE relative to QBOW the difference is not statistically significant. More detail emerging from these various studies has already been given in Section 4.1a, but note in particular that they have all focused on NH winter, which is the season where the observations show significant correlation between the QBO and the MJO (Son et al. 2017, Marshall et al. 2017), and therefore the season where gain in seasonal prediction skill is likely to arise.

Gains in other geographic regions might also be possible, particularly given that the MJO plays a major role in subseasonal to seasonal forecasts in the extratropics. For example Wang et al. (2018b) have noted that the MJO signal in the North Pacific Storm Track is stronger in winter in QBOE years, as might be expected if the MJO signal in the tropics is stronger. Kim et al. (2020b) show that there is QBO modulation of the MJO signal in winter precipitation in East Asia. Mundhenk et al. (2018) (see also Baggett et al. 2017) have demonstrated that skillful subseasonal forecasts of ‘atmospheric river’ events, potentially associated with strong precipitation, on the west coast of North America may be based on a combined QBO-MJO index. The work of Inoue et al. (2011), who considered the effect of the QBO on precipitation in the tropical, subtropical and extratropical Asian region in NH autumn, and Seo et al. (2013) who showed a corresponding effect on precipitation patterns in the tropical, subtropical and extratropical west Pacific in NH spring, suggest that there may be significant gains from exploitation of stratospheric effects in seasonal forecasting in other seasons.

5.2 Model assessment and validation

A different potential exploitation for improved understanding of tropical stratosphere-troposphere coupling is in model assessment and validation. Such assessment is particularly important for models used for climate prediction, where there can be no direct assessment of predictive skill of long-term changes against observations. An indirect approach is to consider instead a model’s ability to simulate variations on shorter time scales, particularly variations which are well characterised in observations. If a model is able to reproduce variability consistent with observations then that builds confidence in model skill more generally, particularly if the physical processes playing a role in that variability are also potentially relevant to long-term change.

Sakazaki et al. (2017) have already noted that model simulations of semi-diurnal variation in rainfall, driven in part by ozone heating in the tropical stratosphere, which are relatively well characterised in observations, vary significantly between different convective parametrizations. They therefore suggested that this variation might be used as a basis for assessment for parametrizations. The apparent effect of the QBO on the MJO might provide a similar opportunity. Even if the QBO effect on the MJO were ‘weak’, in the sense that it could not be incorporated into subseasonal forecasts in a way that added significantly to predictive skill, it provides a component of deterministic time variation to the MJO that could be used for model assessment, noting, of course, that the current situation is that no free-running model reproduces the effect (Lee and Klingaman 2018, Kim et al. 2020a, Lim and Son 2020). This seems potentially valuable given the current wide range of simulated MJO behaviour in climate models (Jiang et al. 2015). There is an ongoing debate over the physical mechanisms that are most important for the MJO, with several candidate theoretical models and, as suggested by Zhang et al. (2020), whether and how such models reproduce an QBO-MJO connection may be a valuable criterion for selecting between them.

6. Outstanding questions and future challenges

The previous sections have summarised the evidence from observations and models that the stratosphere exerts a significant influence on the tropical troposphere, the various coupling mechanisms that have been proposed to account for this influence and the extent to which these proposed mechanisms have been tested or verified by theory or modelling. A comparison has been made with the development of evidence for and understanding of coupling from stratosphere to extratropical troposphere, where there has been much progress over the last two decades, noting the similarities and differences between the extratropical and tropical coupling problems. Figure 12 summarises the range of processes in the stratosphere that potentially couple to different aspects of the behaviour of the tropical troposphere.

6.1 Observations

Some of the suggested tropical tropospheric indications of influence from the stratosphere, particularly the possible QBO signal in Atlantic tropical cyclone frequency, have become less clear as the length of the available data record has increased. Whilst a coherent pattern of a QBO signal on the seasonal and annual mean tropical tropospheric circulation seems gradually to be emerging, as has been noted in Section 3, there is significant uncertainty over details of longitudinal structure and seasonal variation. The length of the data record, both for the QBO, which extends back to the 1950s, and for the tropical troposphere, particularly its variability, is a fundamental limitation. The scope of studies of the relation between QBO and tropical precipitation, for example, is limited by the availability of reliability of tropical precipitation data. The Global Precipitation Climatology Project (GPCP) has combined satellite, sounding and surface observations starting in 1979 and was used in the studies by Liess and Geller (2012), Gray et al. (2018) and Lee et al. (2019). Gray et al. (2018) have compared use of GPCP data against use of precipitation from ERA-40 reanalysis and show that using ERA-40/ERA-I reanalysis data on precipitation, which extends back to 1958, gives similar conclusions and improves statistical significance. In general reanalysis datasets for the late 1950s to late 1970s (the 'pre-satellite era') are regarded as being reliable for large-scale dynamical quantities in the NH extratropics (e.g. Gerber and Martieau 2018), but their reliability for the tropics is less clear. However there may be useful scope for including other quantities from such datasets, including some (e.g. precipitation) that are largely model-generated and only weakly related to direct observations, into these QBO studies. Additionally Hersbach et al. (2017) have demonstrated the value of using upper air data in reanalyses for the 1950s and earlier; this would potentially allow exploitation of reanalysis data for the entire period (1950 onwards) for which direct observations of the QBO are available.

The recent evidence for QBO-MJO connection has stimulated great interest. Whilst the length of observational record that has often been considered is limited, Kim et al. (2020a) have concluded, on the basis of the intrinsic interannual variability of the MJO simulated by models, that the connection is very unlikely to have arisen by chance. The conclusion of Klotzbach et al. (2019), using longer data records, that the connection has emerged only since the 1980s, perhaps because of changes in the temperature structure which have increased the sensitivity of the MJO, now also needs to be taken into account. A similar point is implicit in the separate Camargo and Sobel (2010) discussion of the apparent change in a statistical relation between the QBO and tropical cyclones. Perhaps there have been changes in the sensitivity to the QBO of the intraseasonal variations in the tropical tropospheric circulation and in tropical cyclone behaviour and perhaps the same applies to seasonal timescales as well? In the absence of a clear understanding of relevant mechanisms it is difficult to rule out any of these possibilities.

Turning to observational evidence for the effect of SSWs on the tropical troposphere, further work is clearly needed if the effects suggested on the basis of individual events are to be demonstrated to be systematic and robust. The limitations of the length of the currently available data record are almost certainly at least as great as they are for examining the effect of SSWs on the extratropical troposphere, particularly with regard to the latitude-longitude structure (e.g. Hitchcock and Simpson 2014). Statistical uncertainty in observational evidence for these effects can decrease only slowly in the future. As has been the case for the extratropics, complementing observational evidence with suitably designed modelling studies (see following Section) seems to offer the best route to progress in the near future.

6.2 Global models

The number of numerical modelling studies considering the effect of the stratosphere on the tropical troposphere is still remarkably small. For GCM studies there is a need to examine carefully the robustness of the tropical troposphere response to the QBO across a range of different models, particularly those with different cumulus and radiative parametrizations. There are now several models that simulate a QBO, and this is the focus of the SPARC QBO initiative (QBOi) activity (Anstey et al. 2020), The response of the tropical troposphere to the QBO is probably most efficiently studied, at least initially, by imposing a QBO artificially, as was done in the studies by Giorgetta et al. (1999) and Garfinkel and Hartmann (2011). This would allow, for example, examination of the sensitivity of any tropical tropospheric response to the structure of the QBO in the very lowest part of the stratosphere, where free-running models typically underpredict the amplitude of the QBO in both wind and temperature (e.g. Kim et al. 2020a). Robustness across models is also a key question regarding the effects of coupling to stratospheric chemistry noted by Nowack et al. (2015, 2017) and the effects of stratospheric heating on aerosol geoengineering response discussed by Simpson et al. (2019).

As described in Section 4.1a, seasonal forecasting models have been used to good effect to study the QBO-MJO connection (Marshall et al. 2017, Lim et al. 2019, Wang et al. 2019, Martin et al. 2020). The results from these models are particularly valuable in the absence of simulation of the QBO-MJO connection in free-running GCMs, and they offer further potential for clarifying the role of different processes. Simulations from these models also provide a valuable complement to observations which, as noted above, are limited by the length of the historical record. An approach similar to that taken in some of the seasonal forecasting studies has also been applied by Back et al. (2020) using the WRF mesoscale model at a ‘convection-permitting’ resolution.

GCM studies of the effect of SSWs on the tropical troposphere require that any identifed effect must be distinguished from natural model variability. The need to distinguish a hypothesised effect from natural variability is, of course, a generic requirement that applies also to proposed mechanisms for interannual variability, including the QBO, and for long-term changes, and has motivated ‘large-ensemble’ projects (e.g. Deser et al., 2020). The approach of Hitchcock and Simpson (2014, 2016) in which the stratospheric flow is ‘nudged’ towards a particular specified evolution for a large range of tropospheric initial conditions has been applied very fruitfully to studying the effect of SSWs on the extratropical troposphere. As discussed in Section 4.1b above, Noguchi et al. (2020) have recently applied a similar approach to demonstrate a causal influence of SSWs on the tropical troposphere.

6.3 Cloud-resolving models

The use of CRMs to study possible stratospheric effects on tropical convection has already provided some interesting insights, but again it is important to demonstrate robustness across models with regard to dynamical formulation, microphysical and radiative parametrizations. The Nie and Sobel (2015), Yuan (2015) and Martin et al. (2019) papers cited previously have all used the System for Atmospheric Modelling (Khairoutdinov and Randall 2003) with the radiation scheme from the NCAR Community Climate Model (Kiehl et al. 1998). There is already an ongoing project to make systematic comparison of several CRMs in a set of well defined experimental configurations (Wing et al. 2018) and it would be very interesting to include experiments that perturb lower stratospheric or tropopause level conditions in a multi-model comparison of this type.

The effects of the stratosphere on tropical convection that have been suggested by observational and modelling studies have been on the large-scale, e.g. in shifts in seasonal average patterns or in the amplitude and structure of the MJO. CRM simulations on domains large enough to address these effects directly are now possible (e.g. Satoh et al. 2019) but require enormous computational resources and the scope for long duration integrations or for sensitivity studies is very limited. The weak temperature gradient approach allows CRM simulations on limited spatial domains to be used to address certain questions regarding the large-scale distribution of convection, but what physical effects are missed by this approach and whether those effects might be important in stratosphere-troposphere coupling needs to be considered carefully. For example, this approach cannot capture the non-local coupling between the large-scale moisture field, the convection and the large-scale dynamics that is emphasised by 'moisture mode' theories of the MJO (e.g. Sobel and Maloney 2012, Adames and Kim 2016). Therefore, for example, the Nie and Sobel (2015) result that differing signs of precipitation change in QBO-E vs QBO-W are possible according to the magnitude of the

SST anomaly in the convecting region expresses, within the limited-domain CRM approach, a purely local relation between SST and precipitation change. Whether or not this provides a valid explanation of the spatial variation of the QBOE vs QBOW precipitation change suggested by observations or by GCM studies remains to be investigated.

6.4 Mechanisms

Section 2 has summarised principal pathways – Tropical, Subtropical and Extratropical -- by which the stratosphere may potentially affect the tropical troposphere. Some of these pathways, or components of them, depend on large-scale dynamics, within the troposphere or the stratosphere or both, are relevant to a broad class of climate-dynamics phenomena and might be expected to be captured by most GCMs, though establishing that one pathway or another is important in a particular model simulation is often non-trivial. Potentially a model can be adjusted so that one pathway is eliminated, but it is often difficult to be sure that this sort of adjustment has not had a wider effect on the model behaviour. Gray et al. (2018) have attempted to distinguish between the role of the different pathways in observations by including extra variables in their regression calculation and this kind of approach could be used in model simulations too. Note also that the identified pathways potentially form part of a larger set that control the two-way coupled behaviour of the troposphere-stratosphere system. For example, Yamazaki et al. (2020) have recently suggested that the Tropical Pathway may be important in the much studied connection between the QBO and the extratropical stratosphere, with the QBO effect on the tropical troposphere changing precipitation patterns and hence generation of planetary waves into the extratropical troposphere and stratosphere.

What is specific to tropical, compared to extratropical, stratosphere-troposphere coupling is the potential direct effect on tropospheric convective systems – from above as envisaged in the Tropical Pathway and via the subtropical jet as envisaged in the Subtropical Pathway and Extratropical Pathway together with any feedbacks within the troposphere in which convective systems play a role. As noted previously, three principal mechanisms that have been suggested for a tropospheric response to changes in the stratosphere have been: (i) the effect of changes in tropopause-level vertical wind shear on deep convective systems, (ii) the effect of changes in lower stratospheric temperatures and hence tropopause-level static stability on deep convective systems and (iii) the effect of changes in tropopause-level relative or absolute vorticity on the coupling between deep convective systems and the larger scale circulation in their environment. These mechanisms focus on effects felt directly at tropopause level. For there to be an effect felt through the depth of the troposphere, which is required if there is to be a change in the MJO, or a geographic change in the distribution of precipitation, then there must also be significant feedbacks within the troposphere itself. A clearer understanding of which, if any, of (i), (ii) or (iii) is most important, should give a clearer picture, for example, of which measure of the QBO phase, which as noted in Sections 3.1a and 3.1b, has been defined in different ways by different authors, gives the strongest link to the troposphere.

Given the central role of the detailed dynamics of convective systems, continued investigation of these mechanisms, particularly (i) and (ii), in CRMs is likely to be a productive approach. The Bui et al. (2017) results discussed above in Section 4.2a have shown an effect of shear, but only at levels well below the tropopause. The results recently reported by Martin et al. (2019) in CRM simulations designed to study certain aspects of MJO variability suggest that changes in tropopause level wind shear have only a weak effect (though as with many other aspects of stratosphere-troposphere coupling a wider range of simulations in a wider set of models is needed to confirm this). Therefore current evidence suggests that (ii) is more likely than (i) to be an effective mechanism by which changes at tropopause level or within the lower stratosphere might have a significant effect on the troposphere. This mechanism would be potentially relevant to both QBO effects and SSW effects.

Within (ii), with changes in tropopause-level temperature or static stability being key, different detailed mechanisms are possible. For example, Gray et al. (1992b) seem to envisage that there would be a direct effect on the dynamics of deep convective systems through a combination of the meridional circulation anomaly associated with the QBO, the associated change in the height of the tropopause and the change in static stability at tropopause level which might affect gravity wave dissipation processes. Giorgetta et al. (1999) in analysing the response to an imposed QBO in their GCM simulations emphasised the important role of cloud-radiative effects and these have also been identified as important in the CRM study of Nie and Sobel (2015). The effect of QBO modulation of temperatures near the tropopause on cirrus, and hence through radiative effects on the temperatures and circulation lower in the troposphere, was suggested by Son et al. (2017) for the observed MJO-QBO connection. Hendon and Abhik (2018) and Abhik and Hendon (2019) have noted, respectively in observations and in seasonal prediction model studies, a strong difference in the structure of the MJO upper-level temperature field in QBOE vs QBOW years and argued that the stronger upper-level cold temperature anomaly in QBOE years is suggestive that cirrus radiative feedbacks are important. However establishing that cirrus-radiative effects are playing an active role requires further investigation. Radiative calculations exploiting satellite data on clouds have demonstrated an effect of thin cirrus on the overall radiative balance of the troposphere (e.g. Choi and Ho 2006, Hong et al. 2016), as well as on the TTL (e.g. Fu et al.

2018), and cloud-radiative feedbacks have been invoked in MJO mechanisms (e.g. Raymond 2001, Sobel and Maloney 2012, Adames and Kim 2016), but whether or not tropopause level cirrus could play a significant role in such feedbacks is not yet clear.

The effect of the QBO on the MJO or any other aspect of the tropospheric circulation may be an example of the type of circulation-moisture-cloud-radiation interaction described by Voigt and Shaw (2015) in the context of response to increased greenhouse gases. The same might apply to the corresponding effect of any change in tropopause or lower stratospheric temperatures, induced for example by SSWs or by intraseasonal or interannual changes in the Brewer-Dobson circulation. However, again, many of the ‘high cloud’ changes identified by Voigt and Shaw (2015) are within the upper troposphere rather than being confined to the tropopause, and further work will be needed to establish whether or not the radiative effect of clouds at tropopause level is sufficiently strong to trigger deeper changes in the tropospheric circulation. One approach may be to use ‘mechanism denial’ experiments, in which a set of changes are made to the model representation of different processes and the consequences for the phenomenon of interest noted. This approach has been used effectively in other contexts, e.g. to investigate convective aggregation (e.g. Muller and Bony 2015) and the MJO (e.g. Khairoutdinov and Emanuel 2018). For the stratosphere-troposphere coupling problem it would be natural to investigate the effects of removing e.g. cloud-radiation feedbacks or restricting those feedbacks only to a limited range of levels.

The key insight from work on stratosphere-troposphere coupling in the extratropics is that a major part of the effect of the stratosphere on the troposphere has, as a result of the dynamical feedbacks operating within the troposphere, the spatial pattern of the Northern or Southern Annular Mode. The pattern describes the shape and latitudinal position of the midlatitude jet but also, particularly in the NH, has significant structure in longitude, with important implications for regional weather and climate, This characteristic spatial pattern is seen on timescales ranging from those on monthly (e.g. SSW perturbations) to interannual (e.g. QBO, volcanic perturbations), decadal (e.g. solar cycle) and centennial (e.g. response to changes in long-lived greenhouse gases) timescales (e.g. Kidston et al. 2015, Figure 2).

As reported in this review, and depicted schematically in Figure 12, there are several pieces of observational and modelling evidence for an effect of the stratosphere on the tropical tropospheric circulation, with the effect of perturbations to the tropical lower stratosphere being communicated downward through some combination of dynamical, radiative or cloud-radiative processes and altering the structure of tropospheric convection. These perturbations to the tropical lower stratosphere might be induced, proceeding from left to right in Figure 12, on timescales of days (tides driven by ozone heating), weeks (driven by SSWs and other variations in the extratropical stratospheric circulation), years (e.g. QBO, or variations in the BDC, or effect of volcanic eruptions), to decades and centuries. Some of these effects, indicated by orange arrows in Figure 12, are periodic (diurnal or annual) and others, indicated by blue arrows, are irregular. The amplitudes and geographical patterns of the tropospheric response on these different timescales are not yet fully characterized but there is evidence that the QBO response, for example, is marked by changes in the Walker circulation and the latitudinal distribution of convection in the central and east Pacific. As with the NAM/SAM pattern characteristic of stratosphere-troposphere coupling in the extratropics, this strong spatial variation is almost certainly determined by the feedback mechanisms operating within the troposphere.

Here the problem of understanding stratosphere-troposphere coupling has much in common with the problem of understanding changes in circulation and precipitation that arise as a response to increased greenhouse gases. Mechanisms such as ‘wet get wetter’ or ‘rich get richer’ resulting from internal tropospheric feedbacks have been proposed by e.g. Chou and Neelin (2004) and Held and Soden (2006) and further examined by e.g. Chou et al. (2009). Ma et al. (2018) provide a recent review. Bony et al. (2013) distinguish between ‘thermodynamic’ and ‘dynamical’ changes and argue that the latter play a significant role in differences in predicted changes between different models. There may be similar differences in the predicted response of the tropical troposphere to the QBO, for example, and the fact that this has been examined only in a very small number of models is a further limit on understanding.

6.5 What is the role of the MJO?

The apparent MJO response to the QBO is, unlike other examples of stratospheric influence, specifically a change in intraseasonal variability rather than a change in circulation averaged over the timescale of whatever stratospheric effect is being considered. An emerging debate is between an ‘MJO-centric’ view where the QBO effect on the MJO is the fundamental phenomenon which leads as a consequence to an apparent QBO effect on longer time scales (e.g. anomalies in the seasonal mean state may simply be a result of changes in the strength and frequency of MJO events within that season) or the alternative view where there is an effect of the QBO on the seasonal or longer term state in the troposphere which then leads as a consequence to a change in the strength and frequency of the MJO. The first, ‘MJO-centric’, view is being argued on the basis that the MJO may be particularly sensitive, e.g. through radiative feedbacks, to the temperatures at tropopause level and may therefore feel the QBO directly. This would potentially explain why there seems to be a clear QBO-MJO signal but a much less clear QBO signal in seasonal averages. However the results from seasonal forecast models, most recently that of Martin et al. (2020), suggest that simulated MJO differences between QBOE and QBOW are determined more by some signature of the QBO in the initial conditions than by a sustained effect of the stratospheric QBO state within the simulation. What is not yet clear is whether this is due to 'pre-MJO' structures in the initial state, which would support the MJO-centric view, or due to large-scale properties of the initial state, which would support the alternative view.

This kind of debate is familiar in discussion of the extratropical circulation – is the strength and position of the seasonal mean westerly jet simply a consequence of the relative frequency of high-index vs low-index events, or vice-versa? As always the question is whether the distinction is simply a matter of taste or whether one or the other possibility can be excluded by a careful combination of observation, modelling and theory.

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References

Abhik, S., H. H. Hendon, and M. C. Wheeler, 2019: On the Sensitivity of Convectively Coupled Equatorial Waves to the Quasi-Biennial Oscillation. *J. Climate*, **32**, 5833–5847. https://doi.org/10.1175/JCLI-D-19-0010.1

Abhik, S., and H. H. Hendon, 2019: Influence of the QBO on the MJO during coupled model multiweek forecasts. *Geophys. Res. Lett.*, **46**, 9213– 9221. https://doi.org/10.1029/2019GL083152

Adames, A. F. and D. Kim, D, 2016: The MJO as a dispersive, convectively coupled moisture wave: Theory and observations. *J. Atmos. Sci.*, **73**, 913–941.

Adler, R., M. Sapiano, G. Huffman, J. -J. Wang, G. Gu, D. Bolvin, L. Chiu, U. Schneider, A. Becker, E. Nelkin, P. Xie, R. Ferraro, and D. -B. Shin, 2018: The Global Precipitation Climatology Project (GPCP) Monthly Analysis (New Version 2.3) and a Review of 2017 Global Precipitation. *Atmosphere*, **9**, 138, https://doi.org/10.3390/atmos9040138.

Albers, J. R., G. N. Kiladis, T. Birner, and J. Dias, 2016: Tropical upper-tropospheric potential vorticity intrusions during sudden stratospheric warmings. *J. Atmos. Sci.*, **73**, 2361–2384.

Alexander, M.J., and L. A. Holt, 2019: The Quasi-Biennial Oscillation and its influences at the surface. In *U.S. Clivar Variations*, **17**(1), *Stratosphere-troposphere coupling across timescales.*, ed. A.H. Butler, pp20-26. doi:10.5065/q3jb-9642.

Anstey, J. A., and T. G. Shepherd, 2014: High-latitude influence of the quasi-biennial oscillation. *Quart. J. Roy. Meteorol. Soc.*, **140**,1–21. doi:10.1002/qj.2132.

Anstey J. A., N. Butchart, K. Hamilton, and S. M. Osprey, 2020: The SPARC Quasi-Biennial Oscillation initiative. *Quart. J. Roy. Meteorol. Soc*., https://doi.org/10.1002/qj.3820.

Back, S. -Y., Han, J. -Y., and S. -W. Son, 2020: Modeling evidence of QBO-MJO connection: A case study. *Geophys. Res. Lett.*, **47**, e2020GL089480. https://doi.org/10.1029/2020GL089480

Baggett, C. F., E. A. Barnes, E. D. Maloney, and B. D. Mundhenk, 2017: Advancing atmospheric river forecasts into subseasonal-to-seasonal time scales. *Geophys. Res. Lett.*,**44**, 7528–7536, doi:10.1002/2017GL074434.

Bal, S., S. Schimanke, T. Spangehl, and U. Cubasch, 2017: Variable influence on the equatorial troposphere associated with SSW using ERA-Interim. *J. Earth Syst. Sci.*, **126**, 19, 1-13.

Balachandran, N. K., and D. Rind, 1995: Modeling the effects of UV variability and the QBO on the troposphere/stratosphere system. Part I: The middle atmosphere. *J. Climate*, **8**, 2058-2079, doi:10.1175/1520-0442(1995)008<2058:MTEOUV>2.0.CO;2.

Baldwin, M.P., and T. J. Dunkerton, 2001: Stratospheric harbingers of anomalous weather regimes. *Science*, **294**, 581–584.

Baldwin, M.P., L. J. Gray, T. J. Dunkerton, K. Hamilton, P. H. Haynes, W. J. Randel, J. R. Holton, M. J. Alexander, I. Hirota, T. Horinouchi, D. B. A. Jones, J. S. Kinnersley, C. Marquardt, K. Sato, and M. Takahashi, 2001: The quasi-biennial oscillation. *Rev. Geophys.*, **39**,179–229.

Baldwin, M. P., T. Birner, G. Brasseur, J. Burrows, N. Butchart, R. Garcia, M. Geller, L. Gray, K. Hamilton, N. Harnik, M. L. Hegglin, U. Langematzh, A. Robock, K. Sato, and Scaife, A. A., 2018: 100 Years of Progress in Understanding the Stratosphere and Mesosphere. *Meteor. Monogr.*, **59**, 27.1–27.62. doi: https://doi.org/10.1175/AMSMONOGRAPHS-D-19-0003.1.

Bayr, T., D. Dommenget, T. Martin, and S. B. Power, 2014: The eastward shift of the Walker circulation in response to global warming and its relationship to ENSO variability. *Clim. Dyn.*, **43**, 2747–2763, doi:10.1007/s00382-014-2091-y.

Bhalme, H. N., S. S. Rahalkar, and A. B. Sikder, 1987: Tropical Quasi-Biennial Oscillation of the 10mb Wind and Indian Monsoon Rainfall—Implications for Forecasting. *Journal of Climatology*, **7**, 345-353. http://dx.doi.org/10.1002/joc.3370070403.

Bister, M., and K. A. Emanuel, 2002: Low frequency variability of tropical cyclone potential intensity. 1. Interannual to interdecadal variability. *J. Geophys. Res.*, **107**, 4801, doi:10.1029/2001JD000776.

Bony, S., G. Bellon, D. Klocke, S. Sherwood, S. Fermepin, and S. Denvil, 2013: Robust direct effect of carbon dioxide on tropical circulation and regional precipitation. *Nat. Geosc.*, **6**, 447–451.

Brönnimann, S., A. Malik, A. Stickler, M. Wegmann, C. C. Raible, S. Muthers, J. Anet, E. Rozanov, and W. Schmutz, 2016: Multidecadal variations of the effects of the Quasi-Biennial Oscillation on the climate system. *Atmos. Chem. Phys.*, **16**, 15529-15543, https://doi.org/10.5194/acp-16-15529-2016, 2016.

Bui, H.-H, E. Nishimoto, and S. Yoden, 2017: Downward Influence of QBO-Like Oscillation on Moist Convection in a Two-Dimensional Minimal Model Framework. *J. Atmos. Sci.*, **74**, 3635-3655.

Bui, H., S. Yoden, and E. Nishimoto, 2019: QBO-like oscillation in a three-dimensional minimal model framework of the stratosphere–troposphere coupled system. *SOLA*, **15**, 62−67, doi:10.2151/sola.2019-013.

Butler, A. H., J. P. Sjoberg, D. J. Seidel, and K. H. Rosenlof, 2017: A sudden stratospheric warming compendium, *Earth Syst. Sci. Data*, **9**, 63–76, https://doi.org/10.5194/essd-9-63-2017.

Butler, A., A. Charlton-Perez, D. I. V. Domeisen, C. Garfinkel, E. P. Gerber, P. Hitchcock, A. Y. Karpechko, A. C. Maycock, M. Sigmond, M., I. Simpson and S. -W. Son, 2019: Sub-seasonal predictability and the stratosphere. In: *Sub-Seasonal to Seasonal Prediction: The Gap between Weather and Climate Forecasting*, Robertson, A. W. and Vitart, F. (eds.), Elsevier, pp. 223-241. ISBN 9780128117149

Camargo, S.J., and A. H. Sobel, 2010: Revisiting the influence of the quasi-biennial oscillation on tropical cyclone activity. *J. Climate*, **23**, 5810-5825.

Charlton, A.J., A. O'Neill, P. Berrisford, and W. A. Lahoz, 2005: Can the dynamical impact of the stratosphere on the troposphere be described by large-scale adjustment to the stratospheric PV distribution? *Quart. J. Roy. Meteorol. Soc.*, **131**, 525–543.

Chen, G., and R. A. Plumb, 2009: Quantifying the eddy feedback and the persistence of the zonal index in an idealized atmospheric model. *J. Atmos. Sci.*, **66**, 3707–3720, doi:10.1175/2009JAS3165.1.

Chiodo G, L. Polvani, D. Marsh, A. Stenke, W. Ball, E. Rozanov, S. Muthers, and K. Tsigaridis, 2018: The response of the ozone layer to quadrupled CO_2 concentrations. *J. Clim.*, **31**, 3893–3907.

Choi, Y.-S., and C.-H. Ho, 2006: Radiative effect of cirrus with different optical properties over the tropics in MODIS and CERES observations. *Geophys. Res. Lett.*, **33**, L21811, doi:10.1029/2006GL027403.

Chou, C., and J. D. Neelin, 2004: Mechanisms of global warming impacts on regional tropical precipitation. *J. Climate*, **17**, 2688–2701.

Chou, C., J. D. Neelin, C. -A. Chen, and J. -Y. Tu, 2009: Evaluating the ‘Rich-Get-Richer’ mechanism in tropical precipitation change under global warming. *J. Clim.*, **22**, 1982–2005.

Claud, C., and P. Terray, 2007: Revisiting the Possible Links between the Quasi-Biennial Oscillation and the Indian Summer Monsoon Using NCEP R-2 and CMAP Fields. *J. Climate*, **20**, 773-787.

Collimore, C. C., M. H. Hitchman, and D. W. Martin, 1998: Is there a quasi-biennial oscillation in tropical deep convection? *Geophys. Res. Lett.*, **25**, 333–336.

Collimore, C. C., D. W. Martin, M. H. Hitchman, A. Huesmann, and D. Waliser D., 2003: On the relationship between the QBO and tropical deep convection. *J. Climate*, **16**, 2552-2568.

Coughlin, K., and K. -K. Tung, 2001: QBO signal found at the extratropical surface through northern annular modes. *Geophys. Res. Lett.*, **28**, 4563–4566.

Crooks, S. A., and L. J. Gray, 2005: Characterization of the 11-year solar signal using a multiple regression analysis of the ERA-40 dataset. *J. Climate*, **18**, 996–1015.

Davis, S. M., C. -K. Liang, and K. H. Rosenlof, 2013: Interannual variability of tropical tropopause layer clouds. *Geophys. Res. Lett.*, **40**, 2862– 2866, doi:10.1002/grl.50512.

Densmore, C.R., E. R. Sanabia, and B. S. Barrett, B.S., 2019: QBO Influence on MJO Amplitude over the Maritime Continent: Physical Mechanisms and Seasonality. *Mon. Wea. Rev.,***147**, 389-406.

Deser, C., F. Lehner, K. B. Rodgers, K.B., and Coauthors*,* 2020*:* Insights from Earth system model initial-condition large ensembles and future prospects. *Nat. Clim. Chang.* **10,** 277–286, https://doi.org/10.1038/s41558-020-0731-2.

Dias, J. and G. N. Kiladis, 2019: The Influence of Tropical Forecast Errors on Higher Latitude Predictions. *Geophys. Res. Lett.*, **46**, 4450-4459.

Ding, Q., and Q. Fu, 2018: A warming tropical central Pacific dries the lower stratosphere. *Clim. Dyn.,* **50**, 2813–2827, DOI 10.1007/s00382-017-3774-y.

Domeisen, D. I. V., A. H. Butler, A. J. Charlton-Perez, B. Ayarzagüena, M. P. Baldwin, E. Dunn-Sigouin, J. C. Furtado, C. I. Garfinkel, P. Hitchcock, A. Y. Karpechko, H. Kim, J. Knight, A. L. Lang, E. -P. Lim, A. Marshall, G. Roff, C. Schwartz, I. R. Simpson, S. -W. Son, and M. Taguchi, 2019a: The role of the stratosphere in subseasonal to seasonal prediction. Part I: Predictability of the stratosphere. *J. Geophys. Res.*, **125**, 2019JD030920–

38. http://doi.org/10.1029/2019JD030920.

Domeisen, D. I. V., A. H. Butler, A. J. Charlton-Perez, B. Ayarzagüena, M. P. Baldwin, E. Dunn-Sigouin, J. C. Furtado, C. I. Garfinkel, P. Hitchcock, A. Y. Karpechko, H. Kim, J. Knight, A. L. Lang, E. -P. Lim, A. Marshall, G. Roff, C. Schwartz, I. R. Simpson, S. -W. Son, and M. Taguchi, 2019b: The role of the stratosphere in subseasonal to seasonal prediction. Part II: Predictability arising from stratosphere - troposphere coupling, *J. Geophys. Res.*, **125**, 2019JD030923. http://doi.org/10.1029/2019JD030923

Dowdy, A. J., 2016. Seasonal forecasting of lightning and thunderstorm activity in tropical and temperate regions of the world. *Scientific Reports*, **6**, 20874. https://doi.org/10.1038/srep20874.

Dunkerton, T.J., and M. P. Baldwin, 1991: Quasi-biennial modulation of planetary-wave fluxes in the northern hemisphere winter. *J. Atmos. Sci.,* **48**,1043–1061.

Dunkerton, T.J., C. P. F. Hsu, , and M. E. McIntyre, 1981: Some Eulerian and Lagrangian diagnostics for a model stratospheric warming. *J. Atmos. Sci.*, **38**, 819–843.

Ebdon, R. A., 1975: The quasi-biennial oscillation and its association with tropospheric circulation patterns. *Meteorol. Mag.,***104**, 282–297.

Eguchi, N., and K. Kodera, 2007: Impact of the 2002, Southern Hemisphere, stratospheric warming on the tropical cirrus clouds and convective activity. *Geophys. Res. Lett.*, **34**, L05819, doi:10.1029/2006GL028744.

Eguchi, N., and K. Kodera, 2010: Impacts of stratospheric sudden warming event on tropical clouds and moisture fields in the TTL: A case study. *SOLA*, **6**, doi: 10.2151/sola.2010-035.

Eguchi, N., K. Kodera, and T. Nasuno, 2015: A global non-hydrostatic model study of a downward coupling through the tropical tropopause layer during a stratospheric sudden warming. *Atmos. Chem. Phys.* **15**, 297–304.

Elsner, J. B., J. P. Kossin, and T. H. Jagger, 2008: The increasing intensity of the strongest tropical cyclones. *Nature*, **455**, 92–95.

Emanuel, K. A., 1986: An air-sea interaction theory for tropical cyclones, 1, Steady-state maintenance. J*. Atmos. Sci.*, **43**, 585–604.

Emanuel, K., 2005: Increasing destructiveness of tropical cyclones over the past 30 years. *Nature*, **436**, 686–688.

Emanuel, K., S. Solomon, D. Folini, S. Davis, and C. Cagnazzo, 2013: Influence of tropical tropopause layer cooling on Atlantic hurricane activity. *J. Clim.*, **6**, 2288–2301, doi:10.1175/JCLI-D-12-00242.1.

Fadnavis, S., P. Ernest Raj, P. Buchunde, and B. N. Goswami, 2014: In search of influence of stratospheric Quasi-Biennial Oscillation on tropical cyclones tracks over the Bay of Bengal region. *Int. J. Climatol.*, **34**, 567–580, DOI: 10.1002/joc.3706.

Fereday, D. R., A. Maidens, A. Arribas, A. A. Scaife, and J. R. Knight, 2012: Seasonal forecasts of Northern Hemisphere winter 2009/10. *Environ. Res. Lett.,* **7**, 034031.

Ferraro, A. J., E. J. Highwood, and A. J., Charlton-Perez, 2014: Weakened tropical circulation and reduced precipitation in response to geoengineering. *Env. Res. Lett.*, **9**, 014001.

Ferrara, M., F. Groff, Z. Moon, K. Keshavamurthy, S. M. Robeson, and C. Kieu, 2017: Large-scale control of the lower stratosphere on variability of tropical cyclone intensity. *Geophys. Res. Lett.*, **44**, 4313–4323, doi:10.1002/ 2017GL073327.

Forster, P. M. de F., and K. P. Shine, 1997: Radiative forcing and temperature trends from stratospheric ozone changes. *J. Geophys. Res.*, **102**, 10841–10855, doi:10.1029/96JD03510.

Forster, P. M. D. F., and K. P. Shine, 2002: Assessing the climate impact of trends in stratospheric water vapor. *Geophys. Res. Lett.,* **29**, 1086, doi:10.1029/2001GL013909.

Forster, P. M., G. Bodeker, R. Schofield, S. Solomon, and D. Thompson, 2007: Effects of ozone cooling in the tropical lower stratosphere and upper troposphere. *Geophys. Res. Lett.*, **34**, L23813, doi:10.1029/2007GL031994.

Fu, Q., M. Smith, and Q. Yang, 2018: The Impact of Cloud Radiative Effects on the Tropical Tropopause Layer Temperatures. *Atmosphere*, **9**, 377, doi:10.3390/atmos9100377.

Fuchs, Z. and D. J. Raymond, 2017: A simple model of intraseasonal oscillations. *J. Adv. Model. Earth Syst.*, **9**, 1195–1211, doi:10.1002/ 2017MS000963.

Fueglistaler, S., P. H. Haynes, and P. M. Forster, 2011: The annual cycle in lower stratospheric temperatures revisited. *Atmos. Chem. Phys.*, **11**, 3701-3711.

Fujiwara, M., T. Hibino, S. K. Mehta, L. Gray, D. Mitchell, and J. Anstey, 2015: Global temperature response to the major volcanic eruptions in multiple reanalysis data sets, Atmos. Chem. Phys., 15, 13507–13518, https://doi.org/10.5194/acp-15-13507-2015.

Garfinkel, C. I., and D. L. Hartmann, 2011: The influence of the quasi-biennial oscillation on the troposphere in wintertime in a hierarchy of models. Part I: Simplified dry GCMs. *J. Atmos. Sci.,* **68**, 1273–1289.

Garfinkel, C. I., T. A. Shaw, D. L. Hartmann, and D. W. Waugh, 2012: Does the Holton Tan Mechanism explain how the Quasi-Biennial Oscillation modulates the Arctic Polar Vortex?, *J. Atmos. Sci*., **69**, 1713–1733.

Gerber, E.P., and Coauthors, 2010: Stratosphere–troposphere coupling and annular mode variability in chemistry–climate models. *J. Geophys. Res.*,**115**, DOI: 10.1029/2009JD013770.

Gerber, E. P., and P. Martineau, 2018: Quantifying the variability of the annular modes: reanalysis uncertainty vs. sampling uncertainty. *Atmos. Chem. Phys.*, **18**, 17099–17117, https://doi.org/10.5194/acp-18-17099.

Gilford, D. M., and S. Solomon, 2017: Radiative effects of stratospheric seasonal cycles in the tropical upper troposphere and lower stratosphere, *J. Climate*, **30**, 2769–2783, doi:10.1175/JCLI-D- 16-0633.1.

Gilford, D. M., S. Solomon, and R. W. Portmann, 2016.: Radiative Impacts of the 2011 Abrupt Drops in Water Vapor and Ozone in the Tropical Tropopause Layer. *J. Climate*, **29**, 595–612, doi:10.1175/JCLI-D-15-0167.1..

Gillett, N. P., and D. W. J. Thompson, 2003: Simulation of Recent Southern Hemisphere Climate Change. *Science*, **302**, 273–275.

Giorgetta, M. A., L. Bengtsson, L., and K. Arpe, 1999: An investigation of QBO signals in the east Asian and Indian monsoon in GCM experiments. *Clim. Dyn.*, **15**, 435-450.

Gomez-Escolar, M., N. Calvo, D. Barriopedro, and S. Fueglistaler, 2014: Tropical response to stratospheric sudden warmings and its modulation by the QBO. *J. Geophys. Res.*, **119**, 7382–7395, doi:10.1002/2013JD020560.

Gray, L. J., J. A. Anstey, Y. Kawatani, H. Lu, S. Osprey, and V. Schenzinger, 2018: Surface impacts of the Quasi Biennial Oscillation. *Atmos. Chem. Phys.*, **18**, 8227–8247.

Gray, W. M., 1984: Atlantic seasonal hurricane frequency. Part I: El Niño and 30 mb quasi-biennial oscillation influences. *Mon. Wea. Rev.*, **112**, 1649-1668.

Gray, W. M., J. D. Sheaffer and J. A. Knaff, 1992a: Hypothesized mechanism for stratospheric QBO influence on ENSO variability. *Geophys. Res. Lett.*, **19**, 107-110.

Gray, W. M., J. D. Scheaffer, and J. A. Knaff, 1992b: Influence of the stratospheric QBO on ENSO variability. *J. Meteor. Soc. Japan*, **70**, 975–995.

Hampson, J.E., and P. H. Haynes, 2004: Phase Alignment of the Tropical Stratospheric QBO in the Annual Cycle. *J. Atmos. Sci.*, **21**, 2627-2637.

Hartmann, D. L., J. M. Wallace, V. Limpasuvan, D. W. J. Thompson, and J. R. Holton, 2000: Can ozone depletion and global warming interact to produce rapid climate change?, *Proc. Nat. Acad. Sci. USA*, **97**, 1412– 1417.

Haynes, P. H., C. J. Marks, M. E. McIntyre, T. G. Shepherd, K. P. Shine, 1991: On the ``downward control'' of extratropical diabatic circulations by eddy-induced mean zonal forces. *J. Atmos. Sci.*, **48**, 651--678.

Haywood, J. M., A. Jones, N. Bellouin, and D. Stephenson, 2013: Asymmetric forcing from stratospheric aerosols impacts Sahelian rainfall. *Nature Climate Change*, **3**, 660-665, DOI:10.1038/NCLIMATE1857.

Held, I. M., and B. J. Soden, 2006: Robust responses of the hydrological cycle to global warming. *J. Climate*, **19**, 5686–5699.

Hendon, H. H., and S. Abhik, 2018: Differences in vertical structure of the Madden-Julian Oscillation associated with the quasi-biennial oscillation. *Geophys. Res. Lett.*, **45**, 4419–4428. https://doi.org/10.1029/ 2018GL077207.

Hendon, H. H., E. P. Lim, and S. Abhik, 2020. Impact of Interannual Ozone Variations on the Downward Coupling of the 2002 Southern Hemisphere Stratospheric Warming. *Journal of Geophysical Research: Atmospheres*, **125,** e2020JD032952.

Hersbach, H., S. Brönnimann, L. Haimberger, M. Mayer, L. Villiger, J. Comeaux, A. Simmons, D. Dee, S. Jourdain, C. Peubey, P. Poli, N. Rayner, A. M. Sterin, A. Stickler, M. A. Valente, and S. J. Worley, 2017: The potential value of early (1939–1967) upper-air data in atmospheric climate reanalysis, *Quart. J. Roy. Meteorol. Soc.*, **143**, 1197–1210, https://doi.org/10.1002/qj.3040.

Hitchcock, P., and P. H. Haynes, 2016: Stratospheric control of planetary waves. *Geophys. Res. Lett.*, **43**, 11884–11892, doi:10.1002/2016GL071372.

Hitchcock, P., and I. R. Simpson, 2014: The downward influence of stratospheric sudden warmings. *J. Atmos. Sci.*, **71**, 3856–3876, doi:10.1175/JAS-D-14-0012.1.

Hitchcock, P., and I. R. Simpson, 2016: Quantifying forcings and feedbacks following stratospheric sudden warmings. *J. Atmos. Sci.*, **73**, 3641–3657, doi:10.1175/JAS-D-16-0056.1.

Hitchman, M. H., S. Yoden, P. H. Haynes, V. Kumar, and S. Tegtmeier, 2021: An Observational History of the Direct Influence of the Stratospheric QBO on the Tropical and Subtropical UTLS. *J. Meteor. Soc. Japan*, **99**, https://doi.org/10.2151/jmsj.2021-012.

Ho, C.-H., H. -S. Kim, , J. -H. Jeong, and S. -W. Son, 2009: Influence of stratospheric quasi-biennial oscillation on tropical cyclone tracks in western North Pacific. *Geophys. Res. Lett.*, **36**, L06702, doi:10.1029/2009GL037163.

Hong, Y., G. Liu, and J. -L. F. Li, 2016: Assessing the radiative effects of global ice clouds based on CloudSat and CALIPSO measurements. *J. Climate*, **29**, 7651–7674, doi.org/10.1175/JCLI-D-15-0799.1.

Holton, J. R., and H. C. Tan, 1980: The influence of the equatorial quasi-biennial oscillation on the global circulation at 50mb. *J. Atmos. Sci.*, **37**, 2200–2207.

Holton J. R., and H. C. Tan, 1982. The quasi-biennial oscillation in the Northern Hemisphere lower stratosphere. *J. Meteor. Soc. Japan*, **60**, 140–147.

Holton, J. R., Haynes, P. H., McIntyre, M. E., Douglass, A. R., Rood, R. B., and L. Pfister, 1995: Stratosphere--troposphere exchange. *Revs. Geophys.*, **33**, 403--439.

Huang, B., Z. -Z. Hu, J. L. Kinter III, Z. Wu, and A. Kumar, 2012: Connection of stratospheric QBO with global atmospheric general circulation and tropical SST. Part I: Methodology and composite life cycle. *Clim. Dynam.*, **38**, 1–23, doi:10.1007/s00382-011-1250-7.

Huesmann, A. S., and M. H. Hitchman, 2001: The stratospheric quasibiennial oscillation in the NCEP reanalyses: Climatological structures. *J. Geophys. Res.*, **106**, 11859–11874.

Iles, C. E., G. C. Hegerl, A. P. Schurer, and X. Zhang, 2013: The effect of volcanic eruptions on global precipitation, *J. Geophys. Res*., **118**, 8770–8786, doi:10.1002/jgrd.50678.

Inoue, M., M. Takahashi, and H. Naoe, 2011: Relationship between the stratospheric quasi-biennial oscillation and tropospheric circulation in northern autumn. *J. Geophys. Res.*, **116**, D24115,

Inoue, M., and M. Takahashi, 2013: Connections between the stratospheric quasi-biennial oscillation and tropospheric circulation over Asia in northern autumn, *J. Geophys. Res.*,**118**, 10740–10753, doi:10.1002/jgrd.50827.

Jiang, X., D. E. Waliser, P. K. Xavier, J. Petch, N. P. Klingaman, S. J. Woolnough, B. Guan, G. Bellon, T. Crueger, C. DeMott, C. Hannay, H. Lin, W. Hu, D. Kim, C. -L. Lappen, M. -M. Lu, H.-Y. Ma, T. Miyakawa, J. A. Ridout, S. D. Schubert, J. Scinocca, K. -H. Seo, E. Shindo, X. Song, C. Stan, W. -L. Tseng, W. Wang, T. Wu, X. Wu, K. Wyser. G. J. Zhang, and Z. Hongyan, 2015: Vertical structure and physical processes of the Madden-Julian oscillation: Exploring key model physics in climate simulations. *J. Geophys. Res.*, **120**, 4718–4748, https://doi.org/10.1002/2014JD022375.

Kawatani, Y., K. Hamilton, K. Sato, T. J. Dunkerton, S. Watanabe and K. Kikuchi, 2019: ENSO modulation of the QBO: Results from MIROC models with and without non-orographic gravity wave parameterization. *J. Atmos. Sci.*, **76**, 3893–3897. doi:10.1175/JAS-D-19-0163.1.

Khairoutdinov, M. F., and K. Emanuel, 2018: Intraseasonal variability in a cloud-permitting near-global equatorial aquaplanet model. *J. Atmos. Sci*., **75**, 4337-4355, doi:10.1175/jas-d-18-0152.1.

Khairoutdinov, M. F., and D.A. Randall, 2003: Cloud-resolving modeling of the ARM summer 1997 IOP: Model formulation, results, uncertainties and sensitivities. *J. Atmos. Sci*., 60, 607-625.

Khodri, M., T. Izumo, J. Vialard, S. Janicot, C. Cassou, M. Lengaigne, J. Mignot, G. Gastineau, E. Guilyardi, N. Lebas, A. Robock, and M. J. McPhaden, 2017: Tropical explosive volcanic eruptions can trigger El Niño by cooling tropical Africa. *Nat Commun,* **8,** 778, https://doi.org/10.1038/s41467-017-00755-6.

Kidston, J., A. A. Scaife, S. C. Hardiman, D. M. Mitchell, N. Butchart, M. P. Baldwin, and L. J. Gray, 2015: Stratospheric influence on tropospheric jet streams, storm tracks and surface weather. *Nat. Geosci.*, **8**, 433–440, doi:10.1038/ngeo2424.

Kiehl, J. T., J. J. Hack, G. B. Bonan, B. A. Boville, D. L. Williamson, and P. J. Rasch, 1998: The National Center for Atmospheric Research Community Climate Model: CCM3. J. Climate, 11, 1131–1149, doi:10.1175/1520-0442(1998)011,1131: TNCFAR.2.0.CO;2.

Kiladis, G. N., 1998: Observations of Rossby waves linked to convection over the eastern tropical Pacific. *J. Atmos. Sci.*, **55**, 321– 339.

Kiladis, G. N. and K. M. Weickmann, 1992: Extratropical forcing of tropical Pacific convection during northern winter. Mon. Wea. Rev., 120, 1924–1939, doi:10.1175/1520-0493(1992)120,1924: EFOTPC.2.0.CO;2.

Kim, H., J. H. Richter, and Z. Martin, 2019: Insignificant QBO-MJO prediction skill relationship in the SubX and S2S subseasonal reforecasts. *J. Geophys. Res.*, **124**, doi.org/ 10.1029/2019JD031416.

Kim, H., J. M. Caron, J. H. Richter, and I. R. Simpson, 2020a: The lack of QBO-MJO connection in CMIP6 models. *Geophys. Res. Lett.*, **47**, e2020GL087295. doi.org/ 10.1029/2020GL08729.

Kim, H., S. -W. Son, and C. Yoo, 2020b: QBO modulation of the MJO-related precipitation in East Asia. *J. Geophys. Res.*, **125,** e2019JD031929. https://doi.org/ 10.1029/2019JD031929.

Kinnersley, J. S., and S. Pawson, 1996: The descent rates of the shear zones of the equatorial QBO. *J. Atmos. Sci.,* **53,** 1937–1949.

Klotzbach, P. J., S. Abhik, H. H. Hendon, M. M. Bell, C. Lucas, A. G. Marshall, and E. C. J. Oliver, 2019: On the emerging relationship between the stratospheric quasi-biennial oscillation and the Madden-Julian oscillation. *Scientific Reports,* **9**, 2981, doi: 10.1038/s41598-019-40034-6.

Knutson, T.R. and Coauthors, 2010: Tropical cyclones and climate change. *Nat. Geosci.*, **3**, 157–163.

Kodera, K., 2006: Influence of stratospheric sudden warming on the equatorial troposphere, *Geophys. Res. Lett.*, **33**, L06804, doi:10.1029/2005GL024510.

Kodera, K., N. Eguchi, J. N. Lee, Y. Kuroda, S. Yukimoto, 2011a: Sudden Changes in the Tropical Stratospheric and Tropospheric Circulation during January 2009. *J. Meteor. Soc. Japan*, **89**, 283–290, doi:10.2151/jmsj.2011-308.

Kodera, K., H. Mukougawa, Y. Kuroda, 2011b: A general circulation model study of the impact of a stratospheric sudden warming event on tropical convection, *SOLA*, **7**, 197–200, doi: 10.2151/sola.2011-05010.2151/sola.2011-050.

Kodera, K., B. M. Funatsu, C. M. Claud, and N. Eguchi, 2015: The role of convective overshooting clouds in tropical stratosphere– troposphere dynamical coupling. *Atmos. Chem. Phys.*, **15**, 6767–6774, doi:10.5194/acp-15-6767-2015.

Kodera, K., N. Eguchi, H. Mukougawa, T. Nasuno, and T. Hirooka, T., 2017: Stratospheric tropical warming event and its impact on the polar and tropical troposphere. *Atmos. Chem. Phys.,***17**, 615-625, https://doi.org/10.5194/acp-17-615-2017.

Kossin, J. P., T. L. Olander, and K. R. Knapp, 2013: Trend Analysis with a New Global Record of Tropical Cyclone Intensity. *J. Climate*, **26**, 9960–9976, https://doi.org/10.1175/JCLI-D-13-00262.1.

Kossin, J. P., 2015: Validating atmospheric reanalysis data using tropical cyclones as thermometers, *Bull. Am. Meteorol. Soc.,* **96**, 1089–1096, doi:10.1175/BAMS-D-14-00180.1.

Kravitz, B., D. G. MacMartin, A. Robock, P. J. Rasch, K. L. Ricke, J. N. S. Cole, C. L. Curry, P. J. Irvine, D. Ji, D. W. Keith, J. E. Kristjánsson, J. C. Moore, H. Muri, B. Singh, S. Tilmes, S. Watanabe, S. Yang, S., and J. – H. Yoon, 2014.: A multi-model assessment of regional climate disparities caused by solar geo- engineering. E*nviron. Res. Lett.*, **9**, 074013, doi:10.1088/1748- 9326/9/7/074013.

Kuma, K., 1990.: A quasi-biennial oscillation in the intensity of the intraseasonal oscillation, *Int. J. Climatol*., **10**, 263-278.

Kuroda, Y., 2008: Effect of stratospheric sudden warming and vortex intensification on the tropospheric climate. *J. Geophys. Res.,* **113**, D15110, doi:10.1029/2007JD009550.

Kushner, P. J., and L. M. Polvani, 2004: Stratosphere-troposphere coupling in a relatively simple AGCM: the role of eddies. *J. Climate*, **17**, 629– 639.

Lee, J.-H., M. -J. Kang, and H. -Y. Chun, 2019: Differences in the Tropical Convective Activities at the Opposite Phases of the Quasi-Biennial Oscillation. *Asia-Pac. J. Atmos. Sci.*, **55**, 317-336.

Lee J.C.K., and N. P. Klingaman, 2018: The effect of the quasi-biennial oscillation on the Madden–Julian oscillation in the Met Office Unified Model Global Ocean Mixed Layer configuration. *Atmos. Sci. Lett.*, **19**:e816.

Liess, S., and M. A. Geller, 2012: On the relationship between QBO and distribution of tropical deep convection. *J. Geophys. Res.,* **117**, D03108, doi: 10.1029/2011JD016317.

Li, Y., and D. W. J. Thompson, 2013: The signature of the stratospheric Brewer-Dobson circulation in tropospheric clouds, *J. Geophys. Res.*, **118**, 3486–3494, doi:10.1002/jgrd.50339.

Lim, Y., and S. -W. Son, A. G. Marshall, H. H. Hendon, and K. -H. Seo, K.-H., 2019: Influence of the QBO on MJO prediction skill in the subseasonal- to-seasonal prediction models. *Climate Dynamics*, **53**, 1–15.

Lim, Y., and S. -W. Son, 2020: QBO-MJO connection in CMIP5 models. *J. Geophys. Res.*, **125**, e2019JD032157, doi.org/ 10.1029/2019JD032157.

Lu, J., G. Chen, D. M. W. Frierson, 2008: Response of the Zonal Mean Atmospheric Circulation to El Nino versus Global Warming. *J. Climate*, **21**, 5835-5851, doi: 10.1175/2008JCLI2200.1.

Ma, J., R. Chadwick, K. -H. Seo, C. Dong, G. Huang, G. R. Foltz, and J. H. Jiang, 2018: Responses of the Tropical Atmospheric Circulation to Climate Change and Connection to the Hydrological Cycle. *Annu. Rev. Earth Planet. Sci.*, **46**, 549–580, doi: 10.1146/annurev-earth- 082517- 010102

Madhu, V., 2014: Variation of Zonal Winds in the Upper Troposphere and Lower Stratosphere in Association with Deficient and Excess Indian Summer Monsoon Scenario. *Atmospheric and Climate Sciences*, **4**, 685-695,. doi;10.4236/acs.2014.44062 698.

Manola, I., R. J. Haarsma, and W. Hazeleger, 2013: Drivers of North Atlantic Oscillation events. *Tellus A*, **65**, 1-13.

Manzini, E., and Coauthors, 2014: Northern winter climate change: Assessment of uncertainty in CMIP5 projections related to stratosphere-troposphere coupling, *J. Geophys. Res.,* **119**, 7979–7998, doi:10.1002/2013JD021403.

Marsh, D.R., J. F. Lamarque, A. J. Conley, and L.M. Polvani, 2016: Stratospheric ozone chemistry feedbacks are not critical for the determination of climate sensitivity in CESM1 (WACCM). *Geophys. Res. Lett.*, **43**, 3928-3934.

Marshall, A.G., H. H. Hendon, S. -W. Son, and Y. Lim, 2017: Impact of the Quasi-Biennial Oscillation on predictability of the Madden-Julian Oscillation. *Clim. Dyn.*, **49**, 1365–1377 doi:10.1007/s00382-016-3392-0.

Martin, Z., S. Wang, J. Nie, and A. Sobel, 2019: The Impact of the QBO on MJO Convection in Cloud-Resolving Simulations. *J. Atmos. Sci.*, **76**, 669–688, https://doi.org/10.1175/JAS-D-18-0179.1.

Martin, Z., F. Vitart, S. Wang, and A. Sobel, 2020: The impact of the stratosphere on the MJO in a forecast model. *J. Geophys. Res.*, **125**, e2019JD032106.

Martineau, P., and Son, S.-W., 2015: Onset of circulation anomalies during stratospheric vortex weakening events: The role of planetary-scale waves. *Journal of Climate*, **28**,, 7347– 7370.

Ming, A., A. C. Maycock, P. Hitchcock, and P. Haynes, P., 2017: The radiative role of ozone and water vapour in the annual temperature cycle in the tropical tropopause layer. *Atmos. Chem. Phys.*, **17**, 5677–5701, https://doi.org/10.5194/acp-17-5677-2017.

Monks, P. S., A. T. Archibald, A. Colette, O. Cooper, M. Coyle, R. Derwent, D. Fowler, C. Granier, K. S. Law, G. E. Mills, D. S. Stevenson, O. Tarasova, V. Thouret, E. von Schneidemesser, R. Sommariva, O. Wild, and M. L. Williams, 2015: Tropospheric ozone and its precursors from the urban to the global scale from air quality to short-lived climate forcer. *Atmos. Chem. Phys.,* **15**, 8889–8973, https://doi.org/10.5194/acp-15-8889-2015.

Mukherjee, B.K., K. Indira, R. S. Reddy, and Bh. V. Ramana Murthy, 1985: Quasi-Biennial Oscillation in Stratospheric Zonal Wind and Indian Summer Monsoon. *Monthly Weather Review*, **113**,1421-1424.

Mukougawa, H., H. Sakai, and T. Hirooka, 2005: High sensitivity to the initial condition for the prediction of stratospheric sudden warming. *Geophys. Res. Lett.*, **32**, L17806, doi:10.1029/2005GL022909.

Mukougawa, H., T. Hirooka, T. Ichimaru, and Y. Kuroda, 2007: Hindcast AGCM experiments on the predictability of stratospheric sudden warming. In *Nonlinear Dynamics in Geosciences*, A. A. Tsonis and J. B. Elsner, Eds., Springer-Verlag, New York, 221−233, 604 pp.

Muller, C., and S. Bony, 2015: What favors convective aggregation and why?, *Geophys. Res. Lett.*, **42**, 5626 – 5634, doi:10.1002/2015GL064260.

Mundhenk, B. D., E. A. Barnes, E. D. Maloney, and C. F. Baggett, 2018: Skillful Empirical Subseasonal Prediction of Landfalling Atmospheric River Activity using the Madden-Julian

Oscillation and Quasi-Biennial Oscillation. *npj Climate and Atmospheric Science*, **1**, doi: 10.1038/s41612-017-0008-2.

Naito Y, and I. Hirota, 1997. Interannual variability of the northern winter stratospheric circulation related to the QBO and the solar cycle. *J. Meteor. Soc. Japan*, **75**, 925–937.

Newman, P. A., L. Coy, S. Pawson, and L. R. Lait, 2016: The anomalous change in the QBO in 2015–2016. *Geophys. Res. Lett*., **43**, 8791–8797, https://doi.org/10.1002/2016GL070373.

Nie, J., and A. H. Sobel, 2015: Responses of tropical deep convection to the QBO: Cloud-resolving simulations. *J. Atmos. Sci.,* **72**, 3625-3638.

Nishimoto, E., and S. Yoden, 2016: Influence of the Stratospheric Quasi-Biennial Oscillation on the Madden–Julian Oscillation during Austral Summer. *J. Atmos. Sci.*, **73**, 1105-1124.

Nishimoto, E., S. Yoden, and H. -H. Bui, 2016: Vertical momentum transports associated with moist convection and gravity waves in a minimal model of QBO-like oscillation. *J. Atmos. Sci.*, **73**, 2935- 2957.

Noguchi, S., Y. Kuroda, K. Kodera, and S. Watanabe, 2020: Robust Enhancement of Tropical Convective Activity by the 2019 Antarctic Sudden Stratospheric Warming. *Geophys. Res. Lett.,* **47**, e2020GL088743. https://doi.org/10.1029/2020GL088743

Norton, W. A., 2003. Sensitivity of northern hemisphere surface climate to simulation of the stratospheric polar vortex. *Geophys. Res. Lett.*, **30**,1627, doi:10.1029/2003GL016958.

Nowack, P. J., N. L. Abraham, A. C. Maycock, P. Braesicke, J. M. Gregory, M. M. Joshi, and Coauthors, 2015: A large ozone-circulation feedback and its implications for global warming assessments. *Nature Climate Change*, **5**, 41–45. https://doi.org/10.1038/nclimate2451

Nowack, P. J., P. Braesicke, P., N. L. Abraham, and J. A. Pyle, 2017: On the role of ozone feedback in the ENSO amplitude response under global warming, *Geophys. Res. Lett.*, 44, 3858–3866, doi:10.1002/2016GL072418.

Nowack, P. J., N. L. Abraham, P. Braesicke, and J. A. Pyle, 2018. The impact of stratospheric ozone feedbacks on climate sensitivity estimates. *J. Geophys. Res.,***123**, 4630–4641, https://doi.org/10.1002/2017JD027943

Osprey, S. M., N. Butchart, J. R. Knight, A. A. Scaife, K. Hamilton, J. A. Anstey, V. Schenzinger, and C. Zhang, 2016: An unexpected disruption of the atmospheric quasi-biennial oscillation. *Science*, **353**, 1424–1427, https://doi.org/10.1126/.

Perlwitz, J., and N. Harnik, 2004: Downward coupling between the stratosphere and troposphere. The relative roles of wave and zonal mean processes. *J. Climate.*, **17**, 4902-4909.

Polvani, L. M., and P. J. Kushner, 2002: Tropospheric response to stratospheric perturbations in a relatively simple general circulation model, *Geophys. Res. Lett.*, **29**, 1114, doi:10.1029/2001GL014284.

Plumb, R. A., 1977: The interaction of two internal waves with the mean flow: Implications for the theory of the quasi-biennial oscillation, *J. Atmos. Sci*., **34**, 1847–1858.

Plumb, R. A., 1982: Zonally Symmetric Hough Modes and Meridional Circulations in the Middle Atmosphere. *J. Atmos. Sci.*, **39**, 983–991, https://doi.org/10.1175/1520-0469(1982)039<0983:ZSHMAM>2.0.CO;2.

Plumb, R. A., and R. C. Bell, 1982: A model of the quasi- biennial oscillation on an equatorial beta-plane. *Quart. J. Roy. Meteorol. Soc.*, **108**, 335–352, https://doi.org/10.1002/qj.49710845604.

Plumb, R. A., and K. Semeniuk, 2003: Downward migration of extratropical zonal wind anomalies, *J. Geophys. Res.*, **108**, 4223, doi:10.1029/ 2002JD002773.

Ramsay, H. A., 2013: The effects of imposed stratospheric cooling on the maximum intensity of tropical cyclones in axisymmetric radiative-convective equilibrium, *J. Clim.*, **26**, 9977–9985, doi:10.1175/JCLI-D-13-00195.1.

Randel, W. J., 1993: Global variations of zonal mean ozone during stratospheric warming events. *J. Atmos. Sci.*, **50**, 3308-3321.

Randel, W.J., and F. Wu, 2015: Variability of Zonal Mean Tropical Temperatures Derived from a Decade of GPS Radio Occultation Data. *J. Atmos. Sci.*, **72**, 1261-1275, DOI: 10.1175/JAS-D-14-0216.1

Raymond, D. J., 2001: A New Model of the Madden–Julian Oscillation. *J. Atmos. Sci.*, **58**, 2807–2819, https://doi.org/10.1175/1520-0469(2001)058<2807:ANMOTM>2.0.CO;2.

Raymond, D.J., S. L. Sessions, A. H. Sobel, and Z. Fuchs, 2009: The mechanics of gross moist stability. *J. Adv. Model. Earth Syst.*, **1**, , doi:10.3894/JAMES.2009.1.9.

Rind, D., and N. K. Balachandran, N. K., 1995: Modelling the effects of UV variability and the QBO on the troposphere–stratosphere system. Part II: The troposphere. *J. Climate*, **8**, 2080–2095. 727 doi:10.1175/1520-0442(1995)008, 2080: MTEOUV.2.0.CO;2.

Sakaeda, N., J. Dias, and G. N. Kiladis, 2020: The Unique Characteristics and Potential Mechanisms of the MJO-QBO Relationship. *J. Geophys. Res.*, **125**, doi.org/10.1029/2020JD033196.

Sakazaki, T., and K. Hamilton, 2017: Physical Processes Controlling the Tide in the Tropical Lower Atmosphere Investigated Using a Comprehensive Numerical Model. *J. Atmos. Sci.*, **74**, 2467-2487.

Sakazaki, T., K. Hamilton, C. Zhang, and Y. Wang, 2017: Is there a stratospheric pacemaker controlling the daily cycle of tropical rainfall?, *Geophys. Res. Lett.*, **44**, 1998–2006, doi:10.1002/2017GL072549.

Satoh, M., B. Stevens, F. Judt, M. Khairoutdinov, S. -J. Lin, W. M. Putman, and P. Düben, 2019: Global Cloud-Resolving Models. *Curr Clim Change Rep* **5,** 172–184. https://doi.org/10.1007/s40641-019-00131-0

Scaife, A.A. and Coauthors, 2012: Climate change projections and stratosphere-troposphere interaction. *Clim. Dyn.*, **38**, 2089–2097.

Scaife, A. A., R. E. Comer, N. J. Dunstone, J. R. Knight, D. M. Smith, C. MacLachlan, N. Martin, K. A. Peterson, D. Rowlands, E. B. Carroll, S. Belcher, and J. Slingo, 2017: Tropical rainfall, Rossby waves and regional winter climate predictions. *Quart. J. Roy. Meteorol. Soc.,* **143**, 1–11. doi:10.1002/qj.2910.

Scott, R.K., and L. M. Polvani, 2004. Stratospheric control of upward wave flux near the tropopause. *Geophys. Res. Lett.,* **31**, L02115, DOI:10.1029/2003GL017965.

Sentic, S., S. L. Sessions, Z. Fuchs, 2015: Diagnosing convection with weak temperature gradient simulations of DYNAMO. *J. Adv. Model. Earth Syst.*, **7**, doi:10.1002/2015MS000531.

Seo, J., W. Choi, D. Youn, D. -S. R. Park, and J. Y. Kim, 2013: Relationship between the stratospheric quasi-biennial oscillation and the spring rainfall in the western North Pacific. *Geophys. Res. Lett.*, **40**, 5949–5953, doi:10.1002/2013GL058266.

Seviour, W. J. M., S. C. Hardiman, L. J. Gray, N. Butchart, C. Maclachlan, and A. A. Scaife, 2014: Skillful seasonal prediction of the southern annular mode and Antarctic Ozone. *Journal of Climate*, **27***,* 7462-7474. https://doi.org/10.1175/JCLI-D-14-00264.1

Simpson, I. R., M. Blackburn, and J. D. Haigh, 2009. The Role of Eddies in Driving the Tropospheric Response to Stratospheric Heating Perturbations. *J. Atmos. Sci.*, **66**, 1347–1365.

Simpson, I. R., P. Hitchcock, R. Seager, R., and Y. Wu, 2018: The downward influence of uncertainty in the Northern Hemisphere stratospheric polar vortex response to climate change. *J. Clim.,* **31**, 6371–6391, https://doi.org/10.1175/JCLI-D-18-0041.1.

Simpson, I. R., S. Tilmes, J. H. Richter, B. Kravitz, D. G. MacMartin, M. J. Mills, J. T. Fasullo, and A. G. Pendergrass, 2019. The regional hydroclimate response to stratospheric sulfate geoengineering and the role of stratospheric heating. *J. Geophys. Res.*, **124**, 12587 –12616, doi: 10.1029/2019JD031093

Sobel, A.H., and E. Maloney, 2012: An idealized semi-empirical framework for modeling the Madden–Julian oscillation. *J. Atmos. Sci.*, **69**, 1691–1705, doi:10.1175/JAS-D-11-0118.1.

Solomon, S., K. H. Rosenlof, R. W. Portmann, J. S. Daniel, S. M. Davis, T. J. Sanford, and G. -K. Plattner, 2010: Contributions of Stratospheric Water Vapor to Decadal Changes in the Rate of Global Warming. *Science,* **327**, 1219, doi:10.1126/science.1182488.

Son, S.-W., Y. Lim, C. Yoo, H. H. Hendon, and J. Kim, 2017: Stratospheric control of Madden–Julian oscillation. *J. Climate*, **30**, 1909–1922, doi:10.1175/JCLI-D-16-0620.1.

Song, Y., and W. A. Robinson, 2004: Dynamical mechanisms for stratospheric influences on the troposphere. *J. Atmos. Sci.,* **61**, 1711–1725.

Sridharan, S., and S. Sathishkumar, 2011: Observational evidence of deep convection over Indonesian sector in relation with major stratospheric warming events of 2003–04 and 2005–06. *J. Atmos. Sol.-Terr. Phys.*, **73**, 2453–2461, doi:10.1016/j.jastp.2011.09.007.

Stohl, A., and Co-authors, 2003: Stratosphere-troposphere exchange: A review, and what we have learned from STACCATO. *J. Geophys. Res.*, **108**(D12), 8516, doi:10.1029/2002JD002490.

Taguchi, M., 2011: Latitudinal extension of cooling and upwelling signals associated with stratospheric sudden warmings, *J. Meteor. Soc. Japan.*, **89**, 571–580.

Taguchi, M., 2010: Observed connection of the stratospheric quasi-biennial oscillation with El Niño–Southern Oscillation in radiosonde data. *J. Geophys. Res.*, **115**, D18120, doi:10.1029/2010JD014325.

Takahashi, M., 1996: Simulation of the stratospheric quasi-biennial oscillation using a general circulation model, *Geophys. Res. Lett.*, **23**, 661–664.

Takemi, T., and S. Yamasaki, 2020: Sensitivity of the Intensity and Structure of Tropical Cyclones to Tropospheric Stability Conditions. *Atmosphere*, **11**, 411. https://doi.org/10.3390/atmos11040411

Thompson, D.W.J., M. P. Baldwin, and J. M. Wallace, 2002: Stratospheric connection to northern hemisphere weather: implications for prediction. *J. Climate*, **15**, 1421–1428.

Thuburn, J., and G. C. Craig, 2000: Stratospheric influence on tropopause height: The radiative constraint. *J. Atmos. Sci.*, **57**, 17–28.

Tseng, H.-H. and Fu, Q., 2017: Temperature control of the variability of tropical tropopause layer cirrus clouds. *J. Geophys. Res.*, **122**, 11,062–11,075. https://doi.org/10.1002/2017JD027093.

Vecchi, G. A., S. Fueglistaler, I. M. Held, T. R. Knutson, and M. Zhao, 2013: Impacts of atmospheric temperature trends on tropical cyclone activity, *J. Clim.,* **26**, 3877–3891, doi:10.1175/JCLI-D-12-00503.1.

Vecchi, G. A., T. L. Delworth, H. Murakami, Seth. D. Underwood, A. T. Wittenberg, F. Zeng, W. Zhang, J. W. Baldwin, K. T. Bhatia, W. Cooke, J. He, S. B. Kapnick, T. R. Knutson, G. Villarini, K. van der Wiel, W. Anderson, V. Balaji, J. –H. Chen, K. W. Dixon, R. Gudge, L. M. Harris, L. Jia, N. C. Johnson, S. -J. Lin, M. Liu, C. Ho, J. Ng, A. Rosati, J. A. Smith, X. Yang, 2019: Tropical cyclone sensitivities to CO_2 doubling: roles of atmospheric resolution, synoptic variability and background climate changes. *Climate Dynamics*, **53**, 5999–6033 https://doi.org/10.1007/s00382-019-04913-y.

Vitart, F., 2009: Impact of the Madden Julian Oscillation on tropical storms and risk of landfall in the ECMWF forecast system. *Geophys. Res. Lett.*, **36**, L15802, doi:10.1029/2009GL039089.

Vitart, F., and Coauthors, 2017: The Subseasonal to Seasonal (S2S) Prediction Project Database. *Bull. Amer. Met. Soc.*, **98**, 163-73.

Voigt, A., and T. A. Shaw, 2015: Radiative changes of clouds and water vapor shape circulation response to global warming. *Nature Geoscience.*, **8**, 102-106, 10.1038/ngeo2345.

Wallace, J. M., R. L. Panetta, and J. Estberg, 1993: Representation of the equatorial stratospheric Quasi-Biennial Oscillation in EOF phase space. J. Atmos. Sci., 50 (12), 1751–1762, doi:10.778 1175/1520-0469(1993)050h1751:ROTESQi2.0.CO;2.

Wang, S., S. J. Camargo, A. H. Sobel, L. M. Polvani, 2014: Impact of the tropopause temperature on the intensity of tropical cyclones—An idealized study using a mesoscale model, *J. Atmos. Sci.*, **71**, 4333–4348, doi:10.1175/JAS-D-14-0029.1.

Wang, J., H. -M. Kim, and E. K. M. Chang, 2018a: Interannual modulation of Northern Hemisphere winter storm tracks by the QBO. *Geophys. Res. Lett.*, **45**, 2786–2794. https://doi.org/ 10.1002/2017GL076929.

Wang, J., H. -M. Kim, E. K. M. Chang, S. W. Son, 2018b: Modulation of the MJO and North Pacific storm track relationship by the QBO. *J. Geophys. Res.*, **123**, 3976–3992, doi: 10.1029/2017JD027977.

Wang, S., and A. H. Sobel, 2011: Response of convection to relative sea surface temperature: Cloud-resolving simulations in two and three dimensions. *J. Geophys. Res.*,, **116**, D11119, doi:10.1029/2010JD015347.

Wang, S., A. H. Sobel, and J. Nie 2016: Modeling the MJO in a cloud-resolving model with parameterized large-scale dynamics: Vertical structure, radiation, and horizontal advection of dry air. *J. Adv. Model. Earth Syst.*, **8**, 121–139, doi:10.1002/2015MS000529.

Wang, S., M. Tippett, A. H. Sobel, Z. Martin, and F. Vitart, 2019: Impact of the QBO on prediction and predictability of the MJO convection. *J. Geophys. Res.*, **124**, https://doi.org/10.1029/2019JD030575.

Wheeler M.C., and H. H. Hendon, 2004: An all-season real-time multivariate MJO index: development of an index for monitoring and prediction. *Mon. Weather Rev.,* **132**,1917–1932

Wing, A. A., K. Emanuel, K., and S. Solomon, 2015: On the factors affecting trends and variability in tropical cyclone potential intensity, *Geophys. Res. Lett.*, **42**, 8669–8677, doi:10.1002/2015GL066145.

Wing, A. A., K. A. Reed, M. Satoh, B. Stevens, S. Bony, and T. Ohno, 2018: Radiative-Convective Equilibrium Model Intercomparison Project, *Geosci. Model Dev.*, **11**, 793-813, doi:10.5194/gmd-11-793-2018.

Wittman, M. A. H., A. J. Charlton, and L. M. Polvani, 2007: The effect of lower stratospheric shear on baroclinic instability. *J. Atmos. Sci.*, **64**, 479–496, doi:10.1175/JAS3828.1.

Yamazaki, K., T. Nakamura, J. Ukita, and K. Hoshi, 2020: A tropospheric pathway of the stratospheric quasi-biennial oscillation (QBO) impact on the boreal winter polar vortex. *Atmos. Chem. Phys.*, **20**, 5111–5127, https://doi.org/10.5194/acp-20-5111-2020.

Yamashita, Y., H. Akiyoshi, M. Takahashi, 2011: Dynamical response in the Northern Hemisphere midlatitude and high-latitude winter to the QBO simulated by CCSR/NIES CCM, *J. Geophys. Res.*, **116**, D06118, doi:10.1029/2010JD015016.

Yoneyama, K., C. Zhang, and C. N. Long, 2013: Tracking pulses of the Madden-Julian oscillation. *Bull. Am. Meteorol. Soc.*, **94**, 1871–1891, doi: 10.1175/BAMS-D-12-00157.1.

Yoo, C., and S. -W. Son, 2016: Modulation of the boreal wintertime Madden-Julian oscillation by the stratospheric quasibiennial oscillation. *Geophys. Res. Lett.*, **43**, 1392-1398.

Yoshida, K., 2019: Do sudden stratospheric warmings boost convective activity in the tropics? Presented at workshop: *Stratospheric predictability and impact on the troposphere.* ECMWF, Reading, 18-21 November 2019, available from: https://events.ecmwf.int/event/129/contributions/987/attachments/322/585/Stratospheric-WS-Yoshida.pdf

Yuan, W., 2015: ENSO modulation of the QBO, and QBO influence on tropical convection. Ph.D. thesis, Stony Brook University (downloadable from http://hdl.handle.net/11401/76129).

Zhang, C., 2005: Madden-Julian Oscillation, *Rev. Geophys.*, **43**, RG2003, doi:10.1029/2004RG000158.

Zhang, C., and B. Zhang, 2018: QBO-MJO connection. *J. Geophys. Res.*, **123**, 2957–2967. https://doi.org/10.1002/2017JD028171.

Zhang, C., Á. F. Adames, , B. Khouider, B. Wang, and D. Yang, 2020: Four theories of the Madden Julian Oscillation. *Rev. Geophysics*, **58**, e2019RG000685. https://doi.org/10.1029/2019RG000685.

Zhou, X.-L., M. A. Geller, and M. -H. Zhang, 2001: Tropical cold point tropopause characteristics derived from ECMWF reanalyses and soundings. *J. Climate*, **14**, 1823-1838.

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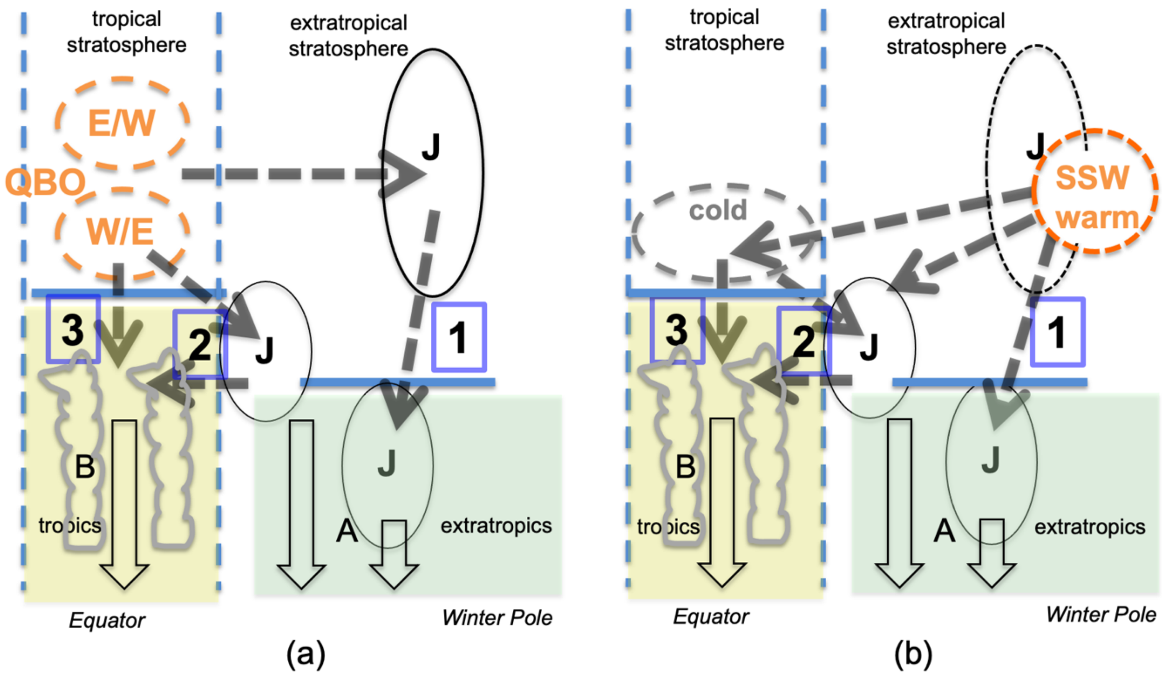

Figure 1: Schematic of pathways for coupling from stratosphere to troposphere for (a) QBO-type (starting in tropical stratosphere) and (b) SSW-type (starting in extratropical stratosphere). (Yellow is the tropical troposphere and green the extratropical troposphere.) The horizontal blue lines indicate the tropopause, higher (at around 15km) in the tropics and lower (at around 10km) in the extratropics. 'J' indicates a jet – stratospheric, subtropical or midlatitude.) Possible pathways for communication are (1) from the extratropical stratosphere to the midlatitude tropospheric jet, (2) from the tropical lower stratosphere to the subtropical jet and (3) from the tropical lower stratosphere directly to tropical upper troposphere. Possible pathways for tropospheric internal communication and feedback are (A) via extratropical dynamics and (B) via tropical dynamics. 1A is an accepted pathway (and 2A has also been demonstrated as a pathway for the effect of the tropical QBO on the extratropical troposphere). 3B and 2B have been suggested, but the mechanisms that might account for these pathways and their importance in the real atmosphere and in models remains uncertain. Note that other pathways not shown in this Figure may be relevant to the coupled

behaviour of the troposphere, rather than that to coupling from the stratosphere to the troposphere. See further comment in Section 6.

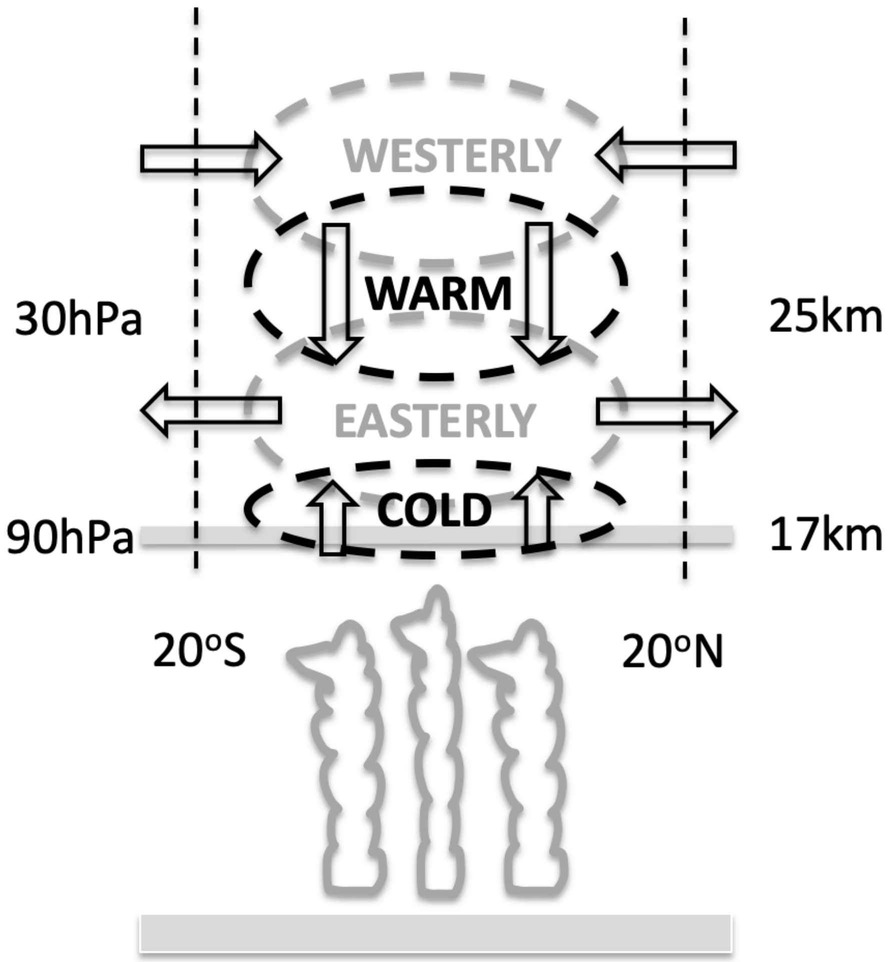

Figure 2: Schematic diagram showing the relation between QBO winds and the corresponding variation in temperatures and meridional circulation, adapted from a similar diagram in Plumb and Bell (1982). The gray line at about 17km indicates the tropical tropopause. The phase of the QBO shown is with easterly winds in the lower stratosphere and westerly winds in the upper stratosphere. Corresponding temperature variations are required from latitudinal integration of the thermal wind equation, given that the QBO wind signal is equatorially confined. For the phase of the QBO shown there are cold temperatures in the lowest part of the stratosphere and warm temperatures in mid-stratosphere. (In the opposite phase of the QBO the signs of all wind and temperature anomalies are reversed.) The wind anomaly in the lower stratosphere is typically -20 m s^{-1} in this phase and 10 m s^{-1} in the opposite phase. The temperature anomaly is typically about -0.5 K at the tropical tropopause increasing to about -2 K above 20 km (see Figure 3). In the phase shown the temperature anomaly is small at about 22 km and then becomes positive above, typically about 3 K at 25 km. Given the long time scale of the QBO, the temperature anomalies must

be maintained against radiative relaxation by the dynamical heating and cooling effects of the meridional circulation. The meridional circulation closes implying opposite signed vertical velocity anomalies and hence opposite signed temperature anomalies away from the equator. In the real atmosphere there are further forces associated with dissipation of planetary and synoptic-scale waves in the subtropics and these appear to be modulated by the QBO, therefore giving a signature in the meridional circulation which extends further poleward than suggested by the schematic. Furthermore the seasonal variation of these waves implies a strong seasonally varying component to the QBO signal in meridional circulation.

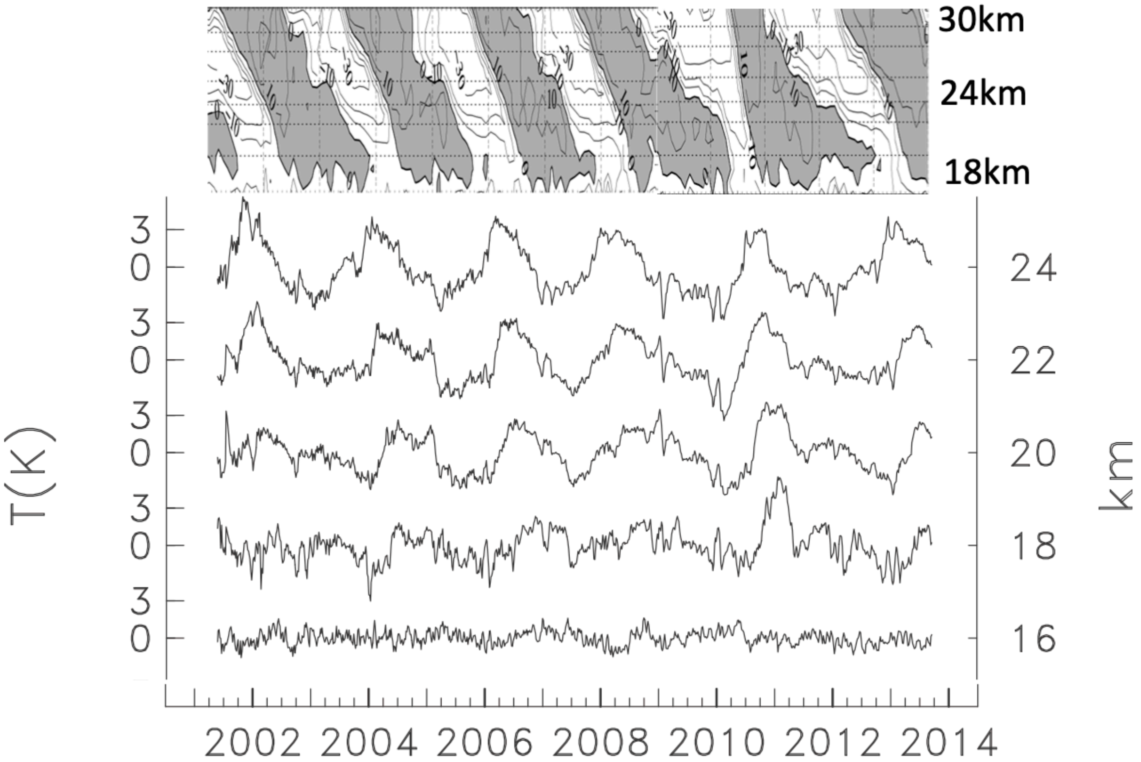

Figure 3: Upper panel: Wind variation in the lower stratosphere from FUB data: https://www.geo.fu-berlin.de/en/met/ag/strat/produkte/qbo/index.html. Westerly (eastward) winds are shaded. Lower panel: (Adapted from Randel and Wu 2015. © American Meteorological Society. Used with permission.). De-seasonalised temperature variations in the tropics at various levels over the period 2001-2013. The temperatures have been calculated from GPS radio occultation data (see Randel and Wu 2015 for further details). There is a clear correspondence between the QBO winds and the interannual temperature variations at 20km and above. There is significant interannual variation of temperatures at 18km but the correspondence with the overall pattern of QBO winds is less clear. At 16km (and below, not shown) interannual variation in temperatures is weak. Other studies, e.g. Randel and Wu (2015) have more systematically extracted a QBO signal in temperatures, using e.g. QBO wind at 50 hPa (about 21 km) or 70 hPa (about 18 km) or using a Principal Component based approach that takes account of the variation in wind at all levels. However the irregular nature of the QBO wind signal in the lower stratosphere (apparent from the Figure) means that some of these approaches may underestimate the strength of the relation between winds and temperatures.

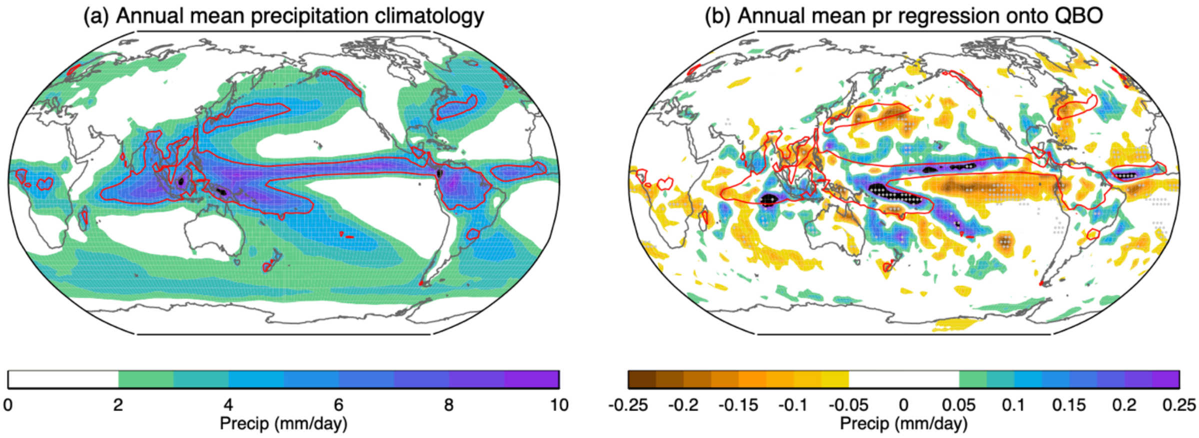

Figure 4: (a) Annual mean precipitation, calculated from fields from the Global Precipitation Climatology Project (GPCP; Adler et al., 2018) dataset at 2.5° latitude–longitude resolution for the period 1979–2019 (http://gpcp.umd.edu/). Red contours correspond to 5 mm/day. (b) The annual average regression of precipitation onto the standardized QBO winds at 50 hPa *multiplied by -1* (to give an estimate of QBOE-QBOW), with 5 mm/day contours for climatological distribution superimposed. This was calculated as follows. The year-by-year time series for each month was regressed against the Nino3.4 index and the variation explained by the regression was removed from the precipitation time series. The resulting time series for each month of the year were then regressed against the standardized QBO index at 50 hPa for that month. Panel (b) then shows minus the annual mean of these monthly regression coefficients. Gray stippled points indicate locations where the regression coefficient is significantly different from zero at the 95% level. This was calculated using a bootstrapping approach with 1000 samples where individual years in the observational record were re-sampled with replacement and the regression analysis performed on the resulting bootstrapped time series. Regions where the 2.5th to 97.5th percentile range of these bootstrapped samples do not encompass zero are considered significant at the 5% level by a two-sided test.

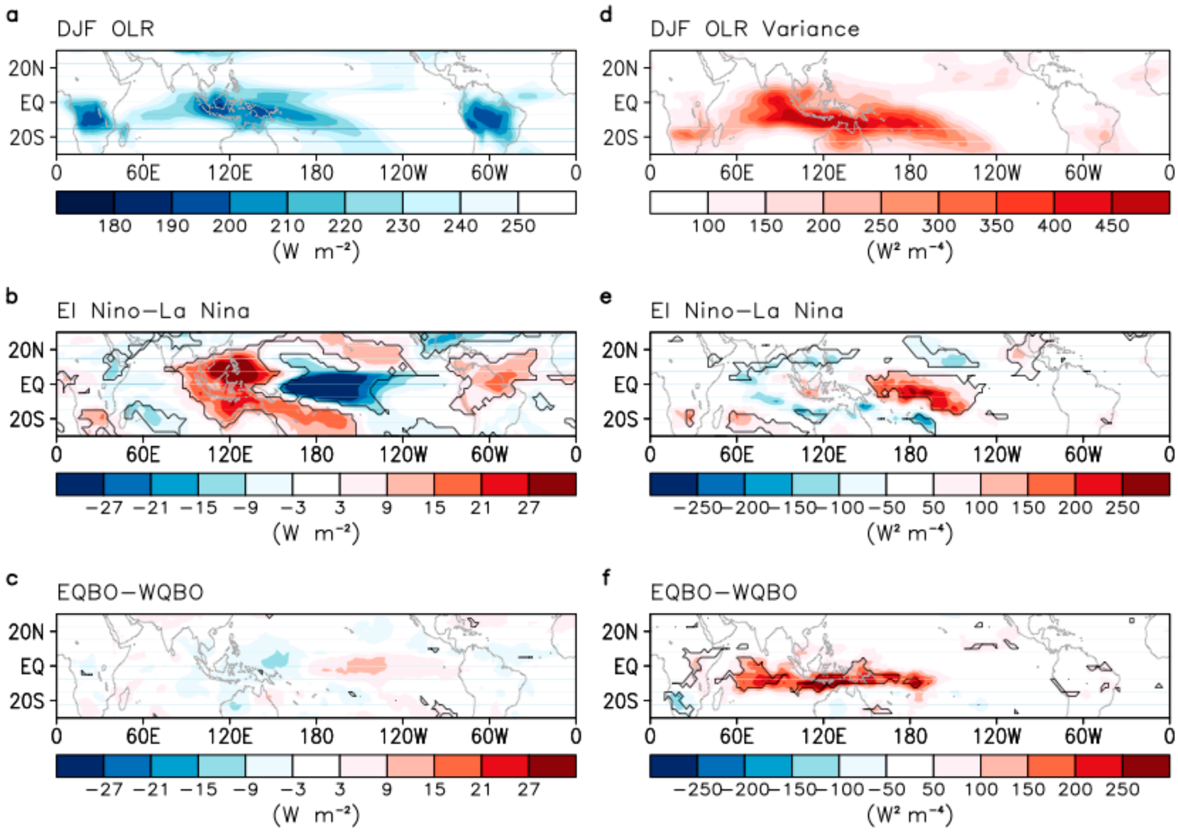

Figure 5. (Son et al. 2017. © American Meteorological Society. Used with permission.) (left) DJF-mean OLR and (right) bandpass-filtered (20–100 days) OLR variance: (a), (d) long-term climatology, (b),(e) interannual difference between El Niño and La Niña winters, and (c), (f) difference between QBOE and QBOW winters. In (b), (c), (e), (f), statistically significant values at the 95% confidence level are contoured.

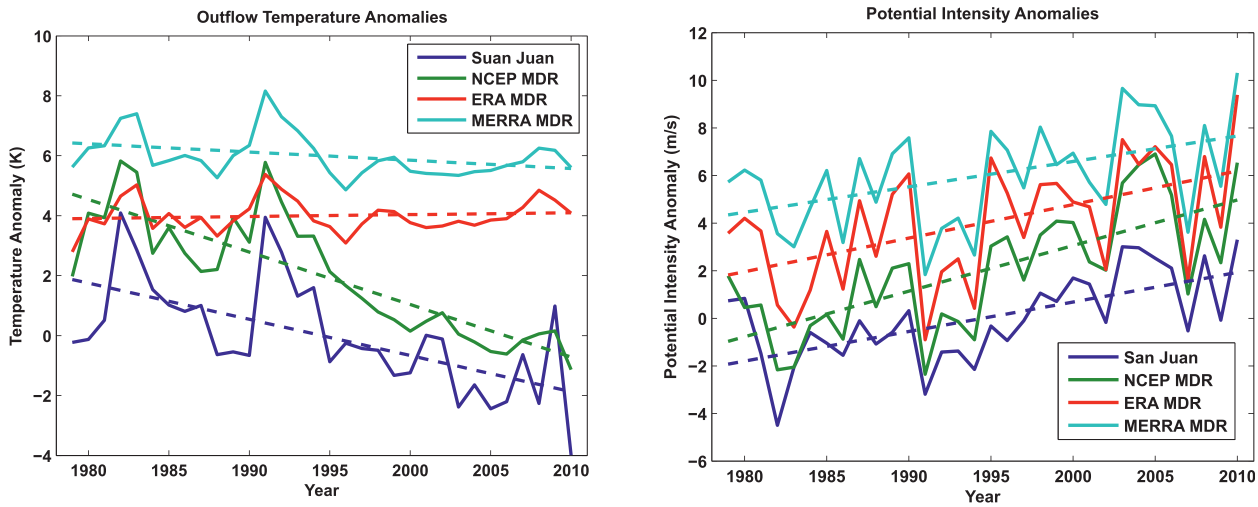

Figure 6: (Emanuel et al. 2013. © American Meteorological Society. Used with permission.) (left) Averaged outflow temperature (T_o) anomalies for the period 1979– 2010: RATPAC (radisonde) station data at San Juan, Puerto Rico (blue); NCEP–NCAR reanalysis data (green); ERA-Interim reanalysis data (red); and MERRA reanalysis data (aqua) with the re-analysis data averaged over the region 6–18N, 20–60W. Dashed lines show the linear regression slopes. The temperature anomalies are with respect to their respective means over the period of record, and 2 K has been added successively to each series for clarity. (right) Corresponding potential intensity (V_p) anomalies, calculated using T_o as displayed in the left-hand panel together with Hadley Centre Global Sea Ice and Sea Surface Temperature. In the left panel, 2 K has been added successively to each timeseries for clarity; in the right panel, 2 m s^{-1} has been added.

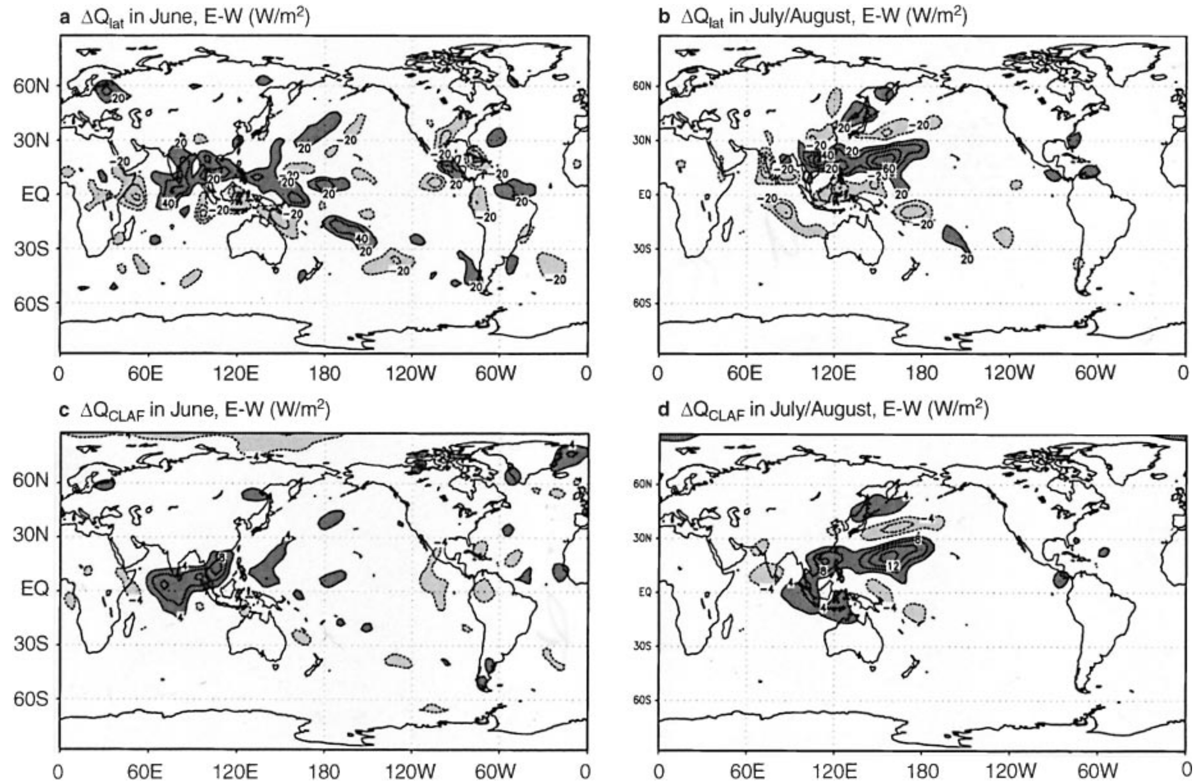

Figure 7: (from Giorgetta et al. 1999. Reproduced by permission of Springer Nature: Climate Dynamics © 1999) Results from numerical simulations in which the tropical stratospheric flow is relaxed to a perpetual QBOE or QBOW state. QBOE has easterly winds in the layer 70-30 hPa and westerly above that. (Signs reversed for QBOW.) (top) QBOE-QBOW latent heating difference for (a) June. (b) corresponding difference for July/August. Shading indicates sign with dark shading positive. (bottom) cloud long-wave atmospheric forcing difference QBOE-QBOW in (c) June and (d) July/August. Giorgetta et al. argue that in QBOE relative to QBOW changes in clouds act to warm the troposphere and cool the tropopause thereby enhancing the tropopause temperature anomaly associated with the QBO.

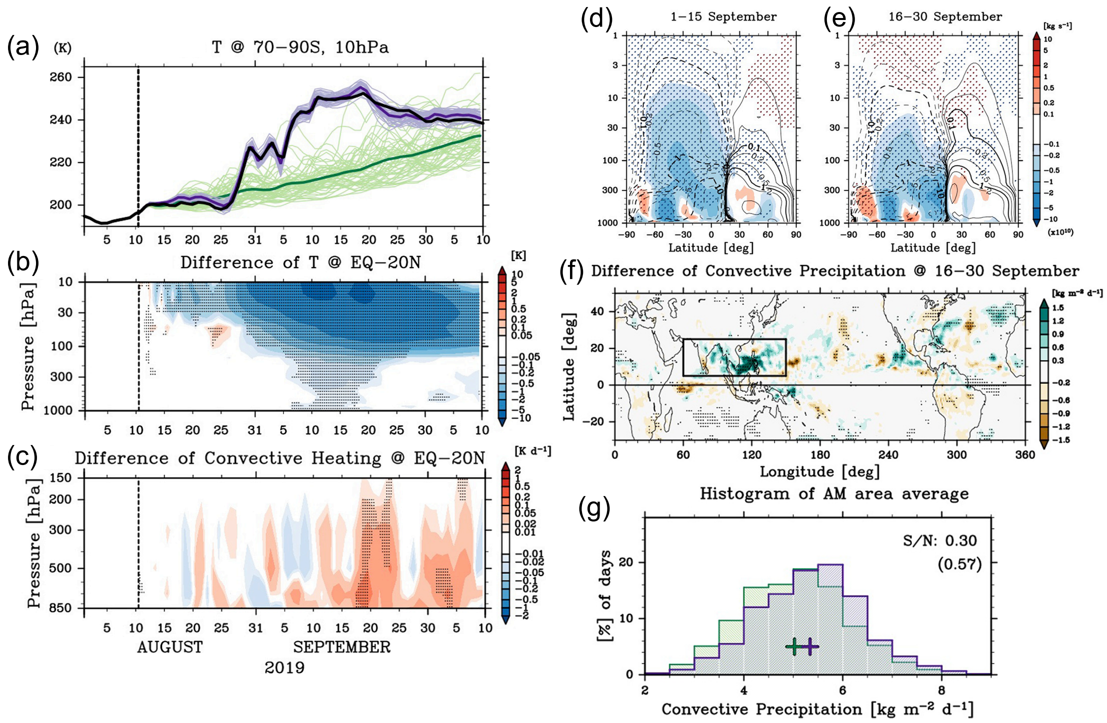

Figure 8: Results from Noguchi et al. (2020). (a) Time series of (a) 10 hPa polar cap (70°S to 90°S) temperature. The thick black line indicates the analysis (JRA-55). Purple lines show ensemble members of the NUDGE forecast from 10 August 2019. Green lines show corresponding for the FREE forecast. Ensemble means are indicated by thick lines. (b) and (c) Time evolution of ensemble mean differences of the NUDGE forecast from the FREE forecast shown as time-height cross sections of (b) the temperature and (c) the heating rate by cumulus convections averaged over the near-equatorial region of the Northern Hemisphere (0–20°N). The regions where the difference is significant at 90% confidence (estimated by Welch's t test) are stippled. (d) and (e) Latitude-height cross sections of the TEM residual mass stream function for (d) 1-15 September 2019 and (e) 16-30 September 2019. The ensemble mean of the NUDGE forecast is shown by contours with a logarithmic interval. The ensemble mean difference of the NUDGE forecast from the FREE forecast is shown by colors. The regions where the positive (negative) difference is significant at 90% confidence (estimated by Welch's t test) are stippled by red (blue) points. (f) Longitude-latitude cross section of the ensemble mean difference between NUDGE and FREE of convective precipitation averaged over 16-30 September 2019. The regions where the difference is significant at 90% confidence (estimated by Welch's t test) are stippled. The black box defines the Asian Monsoon region. (g) Histogram of the daily values of convective precipitation averaged over the Asian Monsoon region for 16-30 September 2019. The purple histogram indicates the NUDGE forecast, the green histogram the FREE forecast. The ensemble and time mean values are shown by crosses. The signal-to-noise ratio (number in brackets) is calculated as the ensemble mean difference divided by the spread of the area-

averaged (and period-averaged) value, which is the mean of the NUDGE and FREE runs. The non-bracketed number is the corresponding value calculated from individual days.

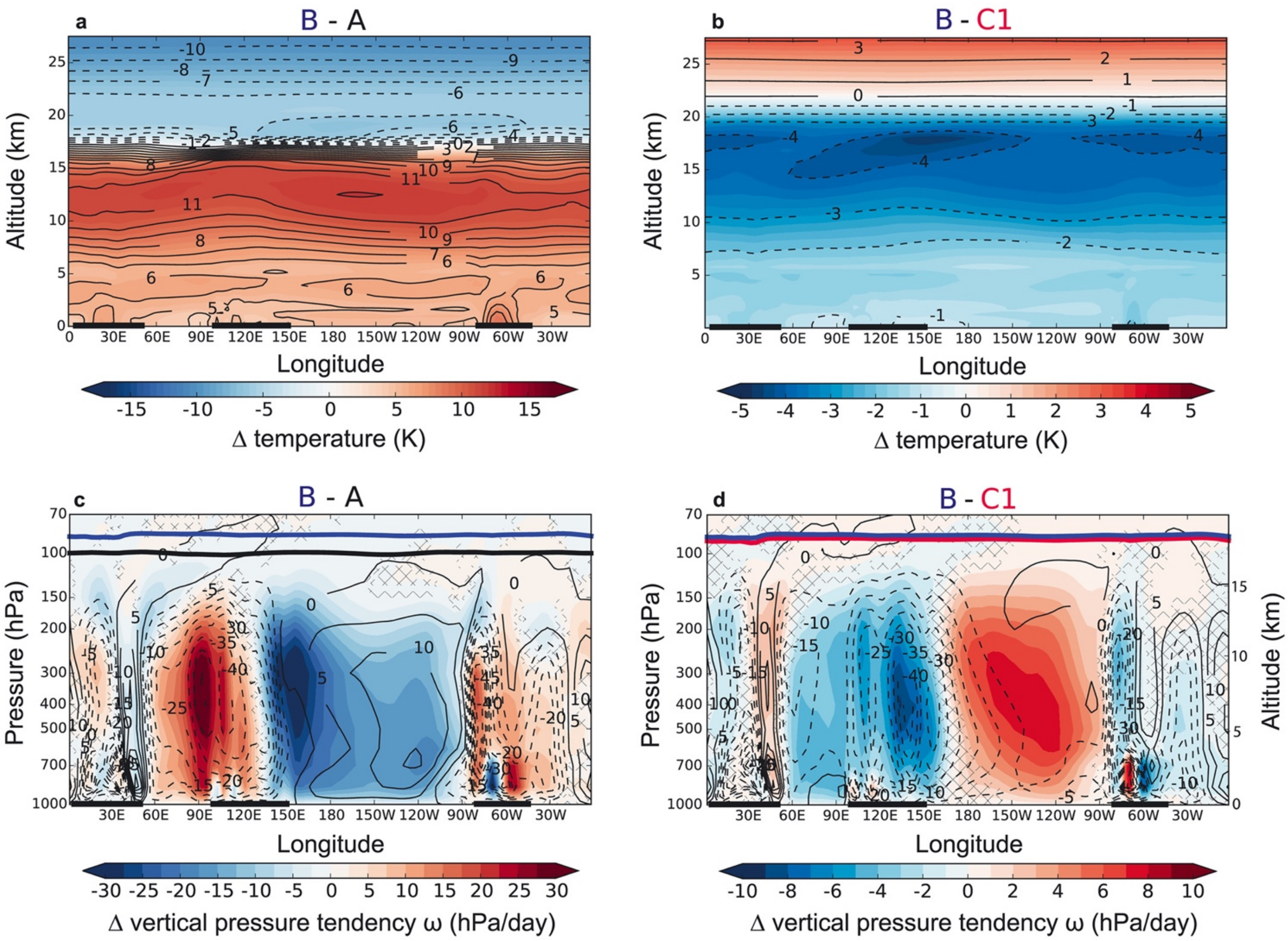

Figure 9: (from Nowack et al. 2017) Results from chemistry-climate model integrations. ‘A’ is a control simulation with interactive ozone, ‘B’ has 4xCO_2 relative to ‘A’ again with interactive ozone, ‘C1’ has 4xCO_2 relative to ‘A’ but the ozone distribution from ‘A’ is imposed. Therefore ‘B-A’ shows the effect of 4xCO_2 including the effect of changed ozone, ‘B-C1’ shows the effect of the changed ozone in ‘B’ relative to that in ‘A’. All quantities shown are 5S-5N averages. (a) and (b) show changes in temperature. (a) shows stronger warming in the upper troposphere relative to the lower troposphere, i.e. a decreased tropospheric lapse rate. (b) shows that changes in ozone play a significant part in this feature and that the decrease in lapse rate in ‘B’ is less than it would have been without ozone feedbacks. The reason is that the changes in ozone tend to cool the upper troposphere, diminishing the warming of the upper troposphere that is expected from increasing CO_2. (c) shows omega (positive values implying descent) for ‘B’ (contours) and the ‘B-A’ difference (shading). (d) shows the omega distribution for ‘B’ contours and the ‘B-C1’ difference (shading). (c) shows a reduction in the strength of the Walker circulation and an eastward shift of the strongest upwelling. (d) shows that the Walker circulation is stronger in ‘B’ relative to ‘C1’, i.e. the effect

of the ozone changes is to lessen the weakening of the Walker circulation that would be driven by increasing CO2 alone.

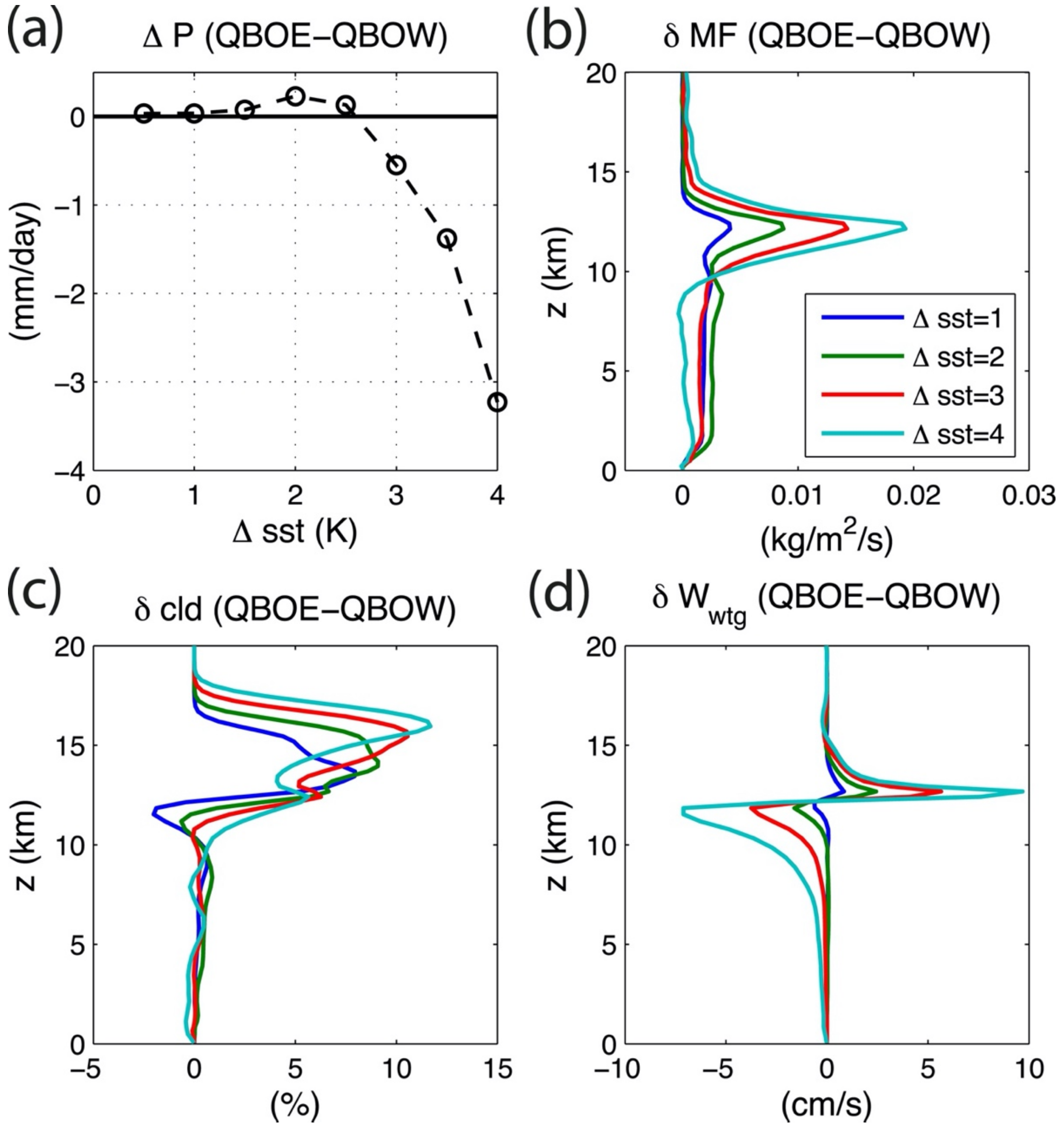

Figure 10 (from Nie and Sobel 2015. © American Meteorological Society. Used with permission.). Difference in various quantities between QBOE vs QBOW simulations (cold vs warm temperature anomalies at tropopause level). Each set of simulations with control (not shown), QBOE and QBOW, has domain averaged temperature set by a radiative-convective equilibrium simulation at a fixed sea surface temperature (SST), to which a uniform SST perturbation ΔSST is added. Quantities displayed are QBOE vs QBOW differences in domain-averaged (a) precipitation, (b) cloud mass flux, (c) cloud fraction and (d) vertical velocity. The weak temperature gradient approach is applied, with the domain-averaged temperature being specified and correspondingly, no mass constraint, with any local mass flux imbalance envisaged as being taken up by mass-exchange with the far-field environment.

Nie and Sobel (2015) discuss the change in sign of the QBOE vs QBOW precipitation response, positive for small ΔSST, negative for larger ΔSST.

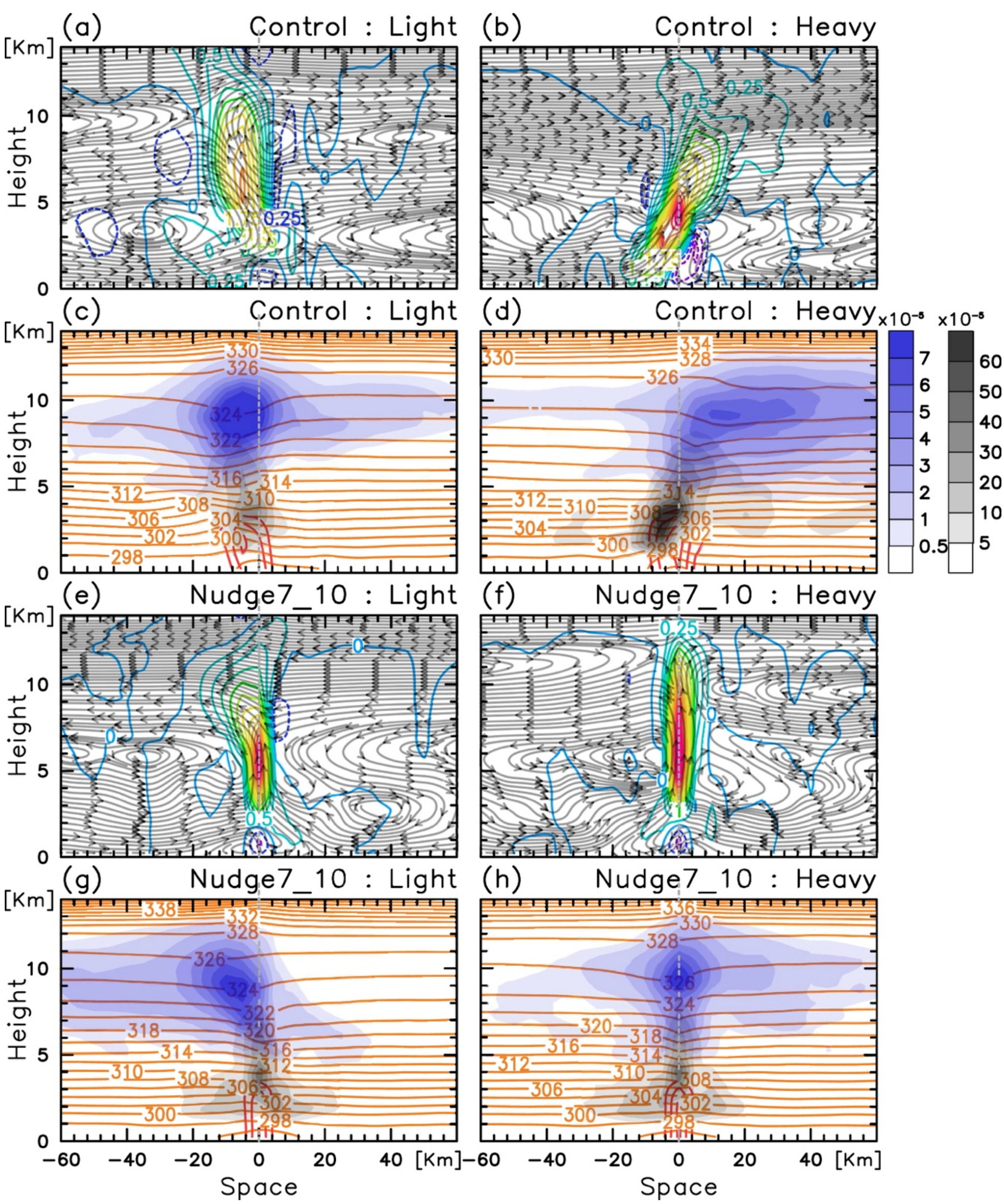

Figure 11: (From Bui et al. 2017. © American Meteorological Society. Used with permission.) Results from two-dimensional cloud-resolving numerical simulations without rotation (a),(b),(e),(f) Cross section of composite vertical speed (rainbow contours; m s^{-1}) and streamline of zonal wind relative to the propagation speed. (c),(d),(g),(h) Cross sections of composite water cloud (gray shades; x10^{-2} g kg^{-1}), ice cloud (blue shades; x10^{-2} g kg^{-1}), rainwater (red contours; x10^{-1} g kg^{-1}), and potential temperature (orange contours; K). (a)-(d) are for a control simulation, (e)-(h) are for a 'nudged simulation' in which the wind in the range 0-8.5 km (note Bui et al. 2017; Eqs. 3 and 4) is highly constrained. (left column)

composite for light precipitation condition and (right column) heavy one. These simulations show how upper level shear can reduce the strength/penetration height of convection, which is one of the mechanisms suggested for QBO influence on the tropical troposphere. (However the relevant level of shear in this case is well below the tropopause.)

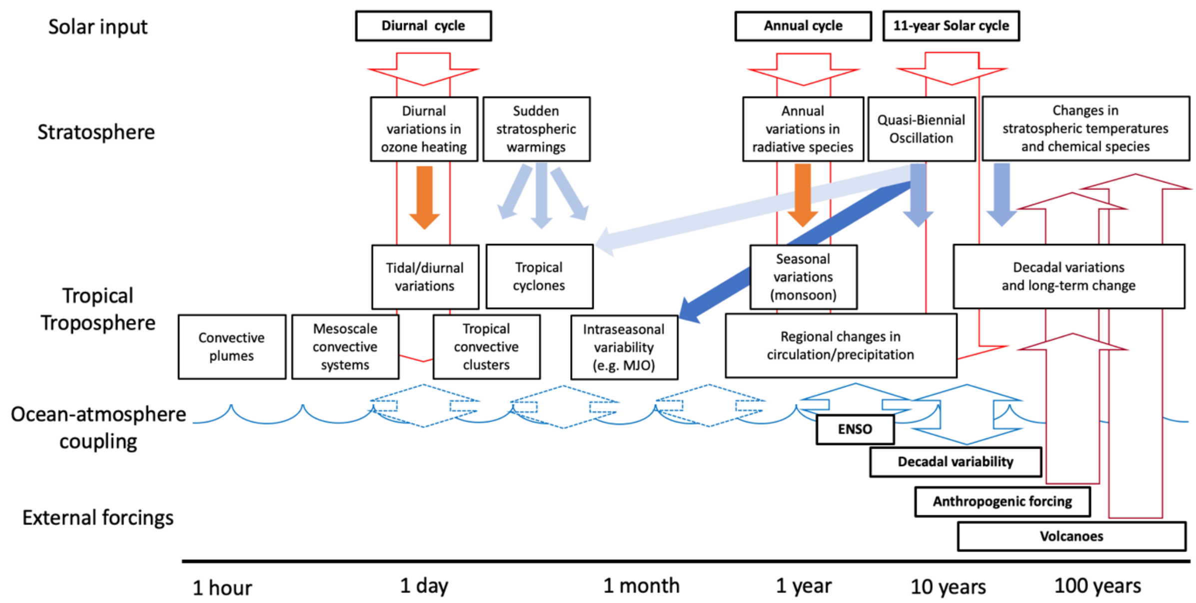

Figure 12. Stratospheric and tropical tropospheric processes on different timescales and possible couplings between them indicated by orange (periodic response to solar forcings) and blue (responses on other timescales) arrows. Darker blue indicates coupling that has been clearly identified from either observations or models, lighter shades indicate coupling for which some evidence exists but which are still subject to uncertainty.