Delayed effect of Arctic stratospheric ozone on tropical rainfall

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Abstract

The tropical precipitation has a wide effect on the tropical economics and social life. Many studies made efforts to improve the tropical precipitation forecast using tropical climate factors. This study, based on observations, found that Arctic stratospheric ozone (ASO) could exert a significant effect on the tropical precipitation, i.e. there is more (less) rainfall over the eastern Pacific and less (more) precipitation over the western Pacific when the ASO anomalies are lower (larger) than normal. It is because a decrease (increase) in ASO could affect El Niño (La Niña) events and lead to a weakened (enhanced) Walker circulation. Time-slice experiments confirmed that the ASO anomalies can force El Niño–Southern Oscillation-like anomalies of tropical sea surface temperature and subsequent tropical precipitation anomalies. In addition, the ASO variations could also change the occurrence probability of extreme precipitation in the tropics. During the anomalously low (high) ASO events, there are more occurrences of heavier precipitation over the eastern Pacific (western Pacific) and of lighter precipitation over the western Pacific (eastern Pacific). Furthermore, the ASO variations lead tropical rainfall by approximately 21 months, suggesting that the ASO can serve as a potentially effective predictor of tropical rainfall.

Keywords: Arctic stratospheric ozone (ASO); tropical rainfall; ENSO; Walker circulation

1. Introduction

Rainfall variability has an enormous impact on human society through modulating the availability of water resources and its effects on, e.g. agricultural yields, forage, and hydroelectric power generation. Which factors affect rainfall variability have been broadly studied, and this topic has been one of the hottest research topics. There are a large number of studies focusing on the effects of important climate change forcings, e.g. El Niño–Southern Oscillation (ENSO), North Atlantic Oscillation, carbon dioxide doubling, etc. on the global rainfall variability. Recently, there are more and more studies pointed out that the stratospheric ozone also can influence precipitation.

Stratospheric ozone is vital in protecting Earth’s life by absorbing harmful solar ultraviolet radiation (Lubin and Jensen, 2002; Chipperfield et al., 2015). It is also a key in controlling of the stratospheric temperature through atmospheric radiative heating that influences stratospheric circulation and chemical composition (Tung, 1986; Haigh, 1994; Ramaswamy et al., 1996; Forster and Shine, 1997), which can even affect tropospheric climate (Baldwin and Dunkerton, 2001; Graf and Walter, 2005; Cagnazzo and Manzini, 2009; Ineson and Scaife, 2009; Reichler et al., 2012; Karpechko et al., 2014; Kidston et al., 2015; Zhang et al., 2016).

There has been a strong decline in Antarctic stratospheric ozone in the past six decades (Solomon, 1990, 1999; Ravishankara et al., 1994, 2009; Dhomse et al., 2006, 2013), which can significantly influence the Southern Hemisphere high and middle latitudes circulation (Son et al., 2008, 2010; Thompson et al., 2011; Gerber and Son, 2014; Waugh et al., 2015), and even affect the Southern Hemisphere extratropical and tropical precipitation (Feldstein, 2011; Kang et al., 2011).

Although Arctic stratospheric ozone (ASO) has not shown such a strong decrease as Antarctic ozone, the amplitude of the inter-annual variability of ASO is comparable with, or even much larger than, that of Antarctic stratospheric ozone (Manney et al., 2011). Thus, the effects of ASO variations on Northern Hemisphere rainfall also deserve investigation. Smith and Polvani (2014) and Calvo et al. (2015) performed numerical experiments and revealed that the precipitation variability over some regions in the high–middle latitudes of the Northern Hemisphere significantly responds to extreme ASO anomaly events. More recently, Ivy et al. (2017) presented observational evidence for linkages between extreme ASO anomalies in March and Northern Hemisphere precipitation in spring (March–April), suggesting that March ASO is a useful indicator of spring-averaged (March–April) precipitation in specific regions of the
Northern Hemisphere high–middle latitudes. However, it is still unknown whether the effect of ASO on precipitation, also like that of Antarctic stratospheric ozone, can extend from high–middle latitudes to tropics. The tropical rainfall variability is significantly modulated by ENSO and subsequent variations in the Walker circulation (Lau and Sheu, 1988; Rasmusson and Arkin, 1993; Dai et al., 1997; Camberlin et al., 2004; Power et al., 2013; Chung and Power, 2014; Huang and Xie, 2015; Huang, 2016; Yim et al., 2016). The Walker circulation is a large overturning cell that spans the tropical Pacific Ocean, characterized by rising motion (lower sea level pressure; SLP) over Indonesia and sinking motion (higher SLP) over the eastern Pacific (Bjerknes, 1969; Gill, 1980). Fluctuations in the Walker circulation reflect changes in the location and strength of tropical upwelling and downwelling, which have important impacts on tropical climate (Horel and Wallace, 1981; Kousky et al., 1984; Wang et al., 2013, 2015; Hu et al., 2016). Based on observational analyses and transient simulations, Xie et al. (2016) recently found that the ASO anomalies possibly affect ENSO events. Based on this study, there is a possibility that the ASO anomalies can affect tropical rainfall variations.

This work is to found the observed evidences that the tropical rainfall variations is related to the ASO changes by linking the changes in ASO, ENSO, the Walker circulation, and tropical rainfall. Furthermore, the second aim of this paper is to perform a series of sensitive experiments using state-of-the-art climate model to demonstrate that the ASO variability can force the ENSO-like sea surface temperature (SST) anomalies and relevant tropical rainfall variations. It not only further extend the implication of Xie et al. (2016) that ASO anomalies affect tropical climate, but also improve the understanding of the characteristic of tropical rainfall variations related to ASO changes.

2. Data, methods, and simulations

In the Northern Hemisphere high latitudes, the variability and depletion of ozone concentrations are most pronounced in the region 60–90°N, at an altitude of 150–50 hPa (Manney et al., 2011). The monthly anomaly of ozone concentration (after removing the climatological mean seasonal cycle) averaged over this region is used as the ASO index. This definition of the ASO index follows Xie et al. (2016). Note that hereafter –ASO index means the inverted ASO index. Ozone values are derived from the Stratospheric Water and OzOne Satellite Homogenized (SWOOSH, 1984–2013) dataset (Davis et al., 2016), which is in good agreement with ozone ($r = 0.89$) from the Global Ozone Chemistry and Related trace gas Data Records for the Stratosphere (GOZCARDS, 1979–2012) project (Froidevaux et al., 2015).

The NASA Global Precipitation Climatology Project (GPCP) monthly precipitation dataset from 1979 to the present, which combines observations and satellite precipitation data into a 2.5°×2.5° global grid (Huffman et al., 2009), is used in this study. SST and SLP are derived from the UK Met Office Hadley Centre for climate prediction and research SST (HadSST) and SLP (HadSLP) field datasets, respectively. Vertical velocity is from the National Centers for Environmental Prediction–Department of Energy (NCEP-DOE) dataset (version 2; NCEP2).

The calculation of the two-tailed Student’s $t$-test and the effective number ($N_{eff}$) of degrees of freedom (DOF) follows Xie et al. (2016). The $N_{eff}$ of DOF is determined by the following approximation:

$$\frac{1}{N_{eff}} \approx \frac{1}{N} + 2 \sum_{j=1}^{N} \frac{N - j}{N - \rho_{XX}(j) \rho_{YY}(j)},$$

where, $N$ is the sample size, and $\rho_{XX}$ and $\rho_{YY}$ are the autocorrelations of two sampled time series, $X$ and $Y$, respectively, at time lag $j$.

The National Center for Atmospheric Research’s (NCAR) Community Earth System Model (CESM, version 1.0.6) is used in this study. CESM includes ocean (POP2), land (CLM4), sea ice (CICE), and interactive atmosphere (CAM/WACCM) components, and is a fully coupled global climate model. The Whole Atmosphere Community Climate Model (WACCM, version 4; Marsh et al., 2013) is utilized for the atmospheric component. WACCM4 is a climate–chemistry model with 66 vertical levels extending from the surface to approximately 140 km, and with approximately 1 km vertical resolution in the tropical tropopause and lower stratosphere layers. For our study, we disabled the interactive chemistry of WACCM4. All simulations employed a horizontal resolution of 1.9°×2.5° (latitude × longitude) for the atmosphere and approximately the same for the ocean. All the forcing data employed in this study are available from the CESM model input data repository. Nine experiments were performed. An overview of all coupled experiments is given in Table 1.

3. Results

Figure 1 shows the correlation coefficients between the global field of rainfall and the –ASO index with the lead of the ASO index ranging from 0 to 21 months, at 3-month intervals. There are almost no significant correlation coefficients between ASO and rainfall variations over the globe while the time lag is less than 15 months (Figures 1(a)–(f)). When the ASO variations lead rainfall by approximately 18 months, however, significant negative correlation coefficients appear over Indonesia and northern South America, and positive correlation coefficients over the tropical eastern Pacific (Figure 1(g)). The patterns of correlation coefficients are most robust for ASO leading rainfall by approximately 21 months (Figure 1(h)). To further probe the relationship between the changes in ASO and tropical rainfall over the region 5°N–5°S and 120°E and the region 0°–10°S and 150°W–120°W
correlation coefficients between the South America, as suggested by the negative (positive) Pacific and downwelling over Indonesia and northern is anomalous upward motion over the tropical eastern anomalies over the western Pacific. Consequently, there are anomalies over the eastern Pacific and negative SST ENSO activity (Xie, 2015; Huang, 2016; Yim et al., 1993; Camberlin et al., 2004; Power et al., 2013; Chung and Power, 2014; Huang and Xie, 2015; Huang, 2016; Yim et al., 2016). An ASO decrease in prior may cause a warm phase of ENSO over Indonesia and northern South America when ASO decreases. All above analysis supports the hypothesis that, as described in Section 1, ASO depletion results in a warm ENSO phase (Xie et al., 2016) which is associated with a weakened Walker circulation (Heureux et al., 2013; Kociuba and Power, 2015). Note that a 35-month low-pass filter is applied in Figure 3. On one hand, the interannual variations of both ASO and tropical signals are affected by quasi-biennial-oscillation (QBO). The 35-month low-pass filter would favor to remove the effect of QBO on the correlation between ASO and tropical vertical velocity and SLP. On the other hand, according to the results in Xie et al. (2016), this connection is gradually achieved from extratropics to tropics via ocean–atmosphere dynamic interaction; i.e. North Pacific Oscillation (NPO) anomalies caused by ASO change force the Victoria Mode (VM) of the North Pacific, which further modulates the development of ENSO through the seasonal footprinting mechanism. Finally, the tropical vertical velocity and SLP is changed with regard to the ENSO development. Previous studies have demonstrated that only the low-frequency variations in VM are effective in modulating the development of ENSO through the seasonal footprinting mechanism (Vimont et al., 2001; Ding et al., 2015). This is why a 35-month low-pass filter is applied to tropical vertical velocity and SLP variations in Figure 3. It would highlight the low-frequency variations, which are related to ozone changes, in the vertical velocity and SLP but remove noise-like high-frequency variations.

An ensemble of nine time-slice experiments (E1x–E3x) is further performed to simulate the influence of ASO changes on the spatial patterns of tropical rainfall. The results show that decreased (increased) rainfall in the tropical eastern Pacific, and vice versa for high ASO.

It is well known that ENSO events can influence the Walker circulation across the equator, thereby affecting tropical precipitation (Horel and Wallace, 1981; Kousky et al., 1984; Lau and Sheu, 1988; Rasmussen and Arkin, 1993; Dai et al., 1997). The patterns of correlation between global rainfall and the ASO index (Figure 1(h)) resemble the ENSO-like anomalies in tropical rainfall that has been noted by previous studies (Lau and Sheu, 1988; Rasmussen and Arkin, 1993; Dai et al., 1997; Camberlin et al., 2004; Power et al., 2013; Chung and Power, 2014; Huang and Xie, 2015; Huang, 2016; Yim et al., 2016). An ASO decrease in prior may cause a warm phase of ENSO activity (Xie et al., 2016); i.e. positive SST anomalies over the eastern Pacific and negative SST anomalies over the western Pacific. Consequently, there is anomalous upward motion over the tropical eastern Pacific and downwelling over Indonesia and northern South America, as suggested by the negative (positive) correlation coefficients between the ASO index and pressure vertical velocity approximately 21 months later over the eastern Pacific (Indonesia and northern South America) (Figure 3(a)). The SLP change over the tropical Pacific is another indicator of change in the Walker circulation. There are negative correlation coefficients between the ASO index and SLP over the eastern Pacific and positive correlations over Indonesia and northern South America (Figure 3(b)), indicating falling SLP over the eastern Pacific and rising SLP over Indonesia and northern South America (Figure 3(a)). The SLP change over the eastern Pacific and rising SLP is larger than the region used to define the ASO index (60–90°E, 5–30°N, 150–50hPa).

Table 1. Fully coupled CESM-WAOCM4 experiments with various specified ozone forcing.

| Exp* | Specified ozone forcing | Other forcing |
|------|-------------------------|---------------|
| E11  | Time-slice run as the control experiment. The specified ozone forcing is a 12-month cycle of monthly ozone averaged from 1980 to 2015. Three ensemble simulations using slightly different initial conditions. | Fixed solar constant, fixed greenhouse gas (GHG) values (averages of emissions scenario A2 of the Intergovernmental Panel on Climate Change (WMO, 2003) over the period 1980–2015), volcanic aerosols (from the stratospheric processes and their role in climate (SPARC) chemistry–climate model validation (CCMVal) REF-B2 scenario recommendations), and QBO phase signals with a 28-month zonal wind fixed cycle. |
| E12  | Same as E1x, except that the ozone in the region 30–90°N at 300–30hPa is decreased by 15% compared with E1x. | Same as E1x. |
| E13  | Same as E1x, except that ozone in the region 30–90°N at 300–30hPa is increased by 15% compared with E1x. | Same as E1x. |
| E21  | Three ensemble simulations using slightly different initial conditions. | |
| E22  | Three ensemble simulations using slightly different initial conditions. | |
| E23  | Three ensemble simulations using slightly different initial conditions. | |
| E31  | Same as E1x. | |
| E32  | Same as E1x, except that ozone in the region 30–90°N at 300–30hPa is increased by 15% compared with E1x. | |
| E33  | Three ensemble simulations using slightly different initial conditions. | |

*Integration time for E1x–3x is 30 years.

To produce different initial conditions, the parameter $p_{pertlim}$ is used in the CESM model, which produces an initial temperature perturbation. The magnitude is about $e^{-14}$. To avoid the effect of the boundary of ozone change on the Arctic stratospheric circulation simulation, the replaced region (30–90°N, 300–30hPa) was larger than the region used to define the ASO index (60–90°N, 150–50hPa).
Figure 1. Correlation coefficients between the ASO index and rainfall, for the ASO index leading rainfall by (a) 0 months, (b) 3 months, (c) 6 months, (d) 9 months, (e) 12 months, (f) 15 months, (g) 18 months, and (h) 21 months. Ozone is based on SWOOSH data. Rainfall is derived from the GPCP data. Only regions with correlations significant at the 95% confidence level are shaded (see Section 2 for details of the statistical significance test).

Figure 2. Lead–lag correlation between the monthly ASO and tropical rainfall variations; the ASO is a time series of ozone averaged over the region 60–90°N at 150–50 hPa based on SWOOSH data; tropical rainfall is averaged in the region 5°N–5°S and 120–150°E for (a) and in the region 0–10°S and 150–120°W for (b) from the GPCP data. The positive months on the x-axis refer to the ASO leading tropical rainfall and negative months refer to tropical rainfall leading the ASO; the dashed lines denote the 90% confidence level.
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Figure 3. Correlation coefficients between the −ASO index and tropical mean (20°S–20°N) pressure vertical velocity (a) and SLP (b), with the ASO index variations leading by 21 months. Ozone is based on SWOOSH data. Pressure vertical velocity is from the NCEP2 reanalysis data. SLP is from Hadley Center. Only regions with correlations significant at the 95% confidence level are shaded (see Section 2 for details of the statistical significance test). All values are detrended, and a 35-month low-pass filter is applied (see Xie et al., 2016).

ASO events can force positive (negative) SST anomalies in the tropical eastern Pacific representing warm (cold) ENSO events (Figures 4(a) and (b)) and force decreased (increased) convective rainfall in Indonesia and northern South America but increased (decreased) convective rainfall in the tropical eastern Pacific (Figures 4(c) and (d)). This result is corresponding to above statistical analysis. The time-slice numerical experiments fully support the statistical conclusions based on observations. Figure 5 shows the probability density distribution versus precipitation over the region 5°N–5°S and 120–150°E and the region 0–10°S and 150–120°W from the experiments E1x–E3x. It is found that there is a larger occurrence probability of heavier precipitation over the region 0–10°S and 150–120°W (5°N–5°S and 120–150°E) and of lighter precipitation over the region 5°N–5°S and 120–150°E (0–10°S and 150–120°W) during the ASO decrease (increase) events. This result further confirms the ASO impacts on tropical rainfall. Furthermore, this figure implies that the ASO decrease (increase) can also improve (suppress) the occurrence of extreme precipitation events over the West Pacific region and suppress (improve) the occurrence over Middle and East Pacific Ocean.

Figure 4. (a) Simulated SST anomalies (°C) caused by a 15% decrease in ASO [(E21 + E22 + E23) − (E11 + E12 + E13)]. (b) Same as (a), but caused by a 15% increase in the ASO [(E31 + E32 + E33) − (E11 + E12 + E13)]. (c) and (d) Same as (a) and (b), but for simulated convective precipitation rate anomalies (m s⁻¹) × 1.0e8.

4. Conclusions and discussions

Observations were used to analyze the influence of ASO changes on tropical rainfall. The atmospheric bridge connecting the stratospheric ozone variations over the polar region, the Walker circulation, and tropical precipitation should be linked to the mechanism by which ASO variations modulate ENSO variability revealed by Xie et al. (2016). Twenty-one months after ASO depletion, the warm ENSO anomaly caused by ASO depletion leads to a weakened tropical Walker circulation, and vice versa after an increase in ASO. The anomalous Walker circulation changes the vertical motion in the tropics, thereby affecting tropical
rainfall. The transient and ensemble time-slice experiments fully support the results from the statistical analysis of the observations. In particular, the ensemble time-slice experiments further confirm that ASO depletion (increase) indeed forces a warm (cold) ENSO-like SST anomalies as suggested by Xie et al. (2016). The simulation also show that more precipitation than normal over the eastern Pacific and less precipitation over Indonesia and northern South America are significantly associated with depleted ASO. The situation is reversed for increased ASO.

An interesting feature in Figure 4 is that the intensities of ENSO-like signs of SST and rainfall anomalies forced by the ASO decrease (Figures 4(a) and (c)) are asymmetry with that forced by the ASO increase (Figures 4(b) and (d)). Figure 6 depicts the sum of SST (rainfall) anomalies caused by the 15% decrease and increase in ASO. It illustrates that the positive SST and rainfall anomalies over center and eastern Pacific (corresponding to El Niño activity) forced by the ASO decrease are larger than the negative anomalies (corresponding to La Niña activity) forced by the ASO increase. The results suggest that the influences of ASO decrease on tropical ENSO-like SST and precipitation anomalies appears to be stronger than those of ASO increase. This may be because the effects of the equal decrease and increase in ASO on Arctic stratospheric circulation are asymmetry (Xie et al., 2008; Hu et al., 2015). There exists an observed phenomenon that the magnitude of El Niño on average tends to be larger than the magnitude of La Niña, and the strongest El Niño is stronger than the strongest La Niña (Burgers and Stephenson, 1999; Kessler, 2002). Factors determining the asymmetry of ENSO have been investigated by many studies (Kang and Kug, 2002; Jin et al., 2003; An and Jin, 2004; Liang et al., 2017). The time-slice experiments in this study (Figures 4 and 6) show a hint of the asymmetry of ENSO related to the ASO changes, which deserves further investigation.

Figure 5. The probability density distribution versus precipitation over the region 5°N–5°S and 120–150°E for (a) and the region 0–10°S and 150–120°W for (b). The black, red and blue lines are calculated from $E_{1x}$, $E_{2x}$, and $E_{3x}$, respectively.

Figure 6. (a) The sum of SST anomalies of Figures 4(a) and (b). (b) The sum of rainfall anomalies of Figures 4(c) and (d).

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