Surface ocean current variations in the North Pacific related to Arctic stratospheric ozone

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Abstract
Using reanalysis datasets and a coupled general circulation model, the relationship between springtime Arctic total column ozone (TCO) and surface (5 m) ocean currents in the North Pacific is investigated. We found that as March Arctic TCO decreases, a statistically significant northwestward ocean current anomaly occurs in the northern North Pacific in surface layer, but an anomalous southward ocean current appears in the central North Pacific in April, and vice versa. The decreased Arctic TCO favors an enhanced Arctic stratospheric circulation, which tends to induce the tropospheric positive Arctic Oscillation anomaly with easterly anomalies over the midlatitude eastern Asia in late March through stratosphere-troposphere dynamical coupling. The easterly anomaly over eastern Asia in late March further extends eastward and induces an easterly anomaly over the midlatitude North Pacific, which favors negative North Pacific Oscillation (–NPO)-like circulation anomaly via anomalous zonal wind shear and the interactions between synoptic scale eddies and the mean flow in early-middle April. The –NPO anomaly forces anomalous northwestward/southward surface ocean currents in the northern/central North Pacific through direct friction of wind and the Coriolis force. Our coupled numerical simulations with high- and low-ozone scenarios also support that the Arctic stratospheric ozone affects the North Pacific surface ocean currents by NPO anomalies. Moreover, the ozone-related ocean current anomalies contribute to Victoria mode-like sea surface temperature anomalies in the North Pacific by horizontal heat advection. These results imply that Arctic ozone signal could be a predictor for variations of the North Pacific surface ocean currents.

Keywords Arctic stratospheric ozone · Arctic stratospheric polar vortex · Stratosphere-troposphere dynamical coupling · North Pacific Oscillation · Surface ocean currents

1 Introduction

Stratospheric ozone not only protects life on the Earth by absorbing solar ultraviolet radiation (e.g., Kerr and Mcelroy 1993; Lubin and Jensen 1995) but also has important effects on the stratospheric temperature and circulation by radiative processes involving ozone (e.g., Ramaswamy et al. 1996; Labitzke and Naujokat 2000; Hu and Tung 2002, 2003; Tian et al. 2010). Moreover, the stratospheric circulation anomalies further affect tropospheric weather and climate via dynamic processes (e.g., Baldwin and Dunkerton 2001; Thompson et al. 2011; Zhang et al. 2016). Thus, the stratospheric ozone variations serve as an proxy for predictions of the tropospheric weather and climate (e.g., Garfinkel 2017; Ivy et al. 2017; Xie et al. 2019).

Due to the dramatic loss of Antarctic stratospheric ozone since the 1980s (e.g., Farman et al. 1985; Solomon 1990, 1999; Ravishankara et al. 1994, 2009; Son et al. 2009b),
many studies investigated the influences of Antarctic stratospheric ozone variations on the tropospheric climate (e.g., Son et al. 2008; Waugh et al. 2009; Hu et al. 2011; England et al. 2016; Xia et al. 2016). The Antarctic stratospheric ozone loss induces a cooler and stronger Antarctic stratospheric polar vortex through radiative cooling (e.g., Randel and Wu 1999), which further migrates to the troposphere through stratosphere-troposphere dynamical coupling (e.g., Song and Robinson 2004; Garfinkel et al. 2013) and contributes to a positive trend in Southern Annual Mode (SAM) in austral summer. This positive SAM trend associated with ozone has widespread effects on the tropospheric climate in the Southern Hemisphere (SH), e.g., temperature and precipitation anomalies over the Antarctic continent (Turner et al. 2005; Marshall et al. 2006; Lenaerts et al. 2018), a poleward shift in the extratropical jet, subtropical dry and precipitation zones (Son et al. 2009a; Son et al. 2010; Polvani et al. 2011; Feldstein 2011; Kang et al. 2011), an extension of the Hadley cell (Min and Son 2013; Gerber and Son 2014; Waugh et al. 2015) and even oceanic circulation in the SH (e.g., Sigmond and Fythe 2010; Bitz and Polvani 2012; Solomon et al. 2015).

Although the Arctic stratospheric ozone loss is smaller in magnitude than that in the Antarctic stratosphere (WMO, 2011), the interannual variability of the Arctic total column ozone (TCO) is large because of the stratospheric polar vortex (SPV) variability (e.g., Solomon et al. 2014). The March Arctic TCO experienced unexpected severe depletion in 1997, 2011 and 2020 (Coy et al. 1997; Lefèvre et al. 1998; Manney et al. 2011; Manney et al. 2020; Rao and Garfinkel 2020). Thus, the effects of Arctic stratospheric ozone changes on the tropospheric circulation, surface temperature and precipitation in the Northern Hemisphere (NH). Calvo et al. (2015) found significant responses in April–May tropospheric wind, temperature and precipitation in the NH to Arctic stratospheric ozone changes. Moreover, the Arctic stratospheric ozone variations in March favor Victoria mode (VM)-like sea surface temperature (SST) anomalies (SSTAs) in the North Pacific in April (Xie et al. 2017a) and thereby the El Niño-Southern Oscillation (Xie et al. 2016).

As mentioned above, previous studies mainly reported the effects of Arctic stratospheric ozone on the tropospheric atmosphere. However, the effects of Arctic stratospheric ozone on the oceanic circulation are not clear at present, although the influences of the Antarctic stratospheric ozone on the oceanic circulation and SST in the SH were widely reported (e.g., Previdi and Polvani 2014; Sigmond and Fythe 2014; Ferreira et al. 2015; Sevior et al. 2016, 2017, 2019). Additionally, the Arctic SPV affects the stratospheric ozone via chemical processes and transport processes (e.g., Liu et al. 2020; Manney et al. 2020). For instance, the severe ozone depletion in 1997, 2011 and 2020 is mainly caused by extreme cooling inside the strengthened SPV (e.g., Coy et al. 1997; Manney et al. 2011; Rao and Garfinkel 2020). The Arctic SPV also affects the troposphere directly. Many studies reported the impacts of Arctic SPV on the tropospheric atmosphere, such as the Arctic Oscillation (AO)/northern annular mode (NAM) (e.g., Baldwin and Dunkerton 2001; Limpasuivan et al. 2004; Ineson and Scaife 2009), tropospheric jet streams (e.g., Sigmond et al. 2008; Kidston et al. 2015), surface temperature anomalies and cold air outbreaks (e.g., Garfinkel et al. 2017; Kretschmer et al. 2018). However, the influence of Arctic SPV on the ocean gets relatively less attentions. Although some studies found that the Arctic stratospheric circulation anomalies significantly affect the Atlantic SST and the Atlantic Meridional Overturning Circulation (e.g., Reichler et al. 2012; Scaife et al. 2013; O’Callaghan et al. 2014), the effects of Arctic stratospheric anomalies on the North Pacific oceanic circulation remain unclear.

The ocean currents have important effects on the climate system and ocean system, such as the global redistribution of heat from tropical to polar regions (e.g., Bigg et al. 2003; Marshall et al. 2001), global surface temperature (e.g., Vellinga and Wood 2002), and hydrography, plankton and fish species (e.g., Hatun et al. 2005; Hatun et al. 2009). Thus, it is necessary to examine the relationship between Arctic stratospheric ozone changes and surface ocean current variations. Additionally, although Xie et al. (2017a) found that the Arctic stratospheric ozone in March affects the North Pacific SSTs in April, details of the mechanism are not clear, i.e., whether the Arctic stratospheric ozone affects the North Pacific SSTs via ocean currents. In this study, we attempt to explore the relationship between Arctic stratospheric ozone changes and the surface ocean current variations in the North Pacific and the mechanism by which the Arctic ozone influences the North Pacific SSTs. The remainder of this paper is organized as follows. Section 2 describes data, model and methods. Section 3 analyzes the connections between the Arctic TCO and the North Pacific surface ocean currents.
and underlying mechanisms. Section 4 presents the results of numerical simulations with high- and low-ozone scenarios. Section 5 discusses the mechanism by which the Arctic TCO influences the North Pacific SSTs. Section 6 gives conclusions and discussions.

2 Data, model and methods

2.1 Reanalysis datasets

The TCO data are from the multisensor reanalysis (MSR) dataset with a horizontal resolution of $0.5^\circ \times 0.5^\circ$ (latitude × longitude) (van der et al. 2010, 2015). Geopotential height (GH), temperature and horizontal wind data are obtained from the European Centre for Medium-Range Weather Forecasts (ECMWF) Reanalysis-Interim (ERA-Interim) dataset (Dee et al. 2011). The surface sensible heat flux, latent heat flux, surface solar radiative flux and longwave radiative flux data are also from ERA-Interim dataset. The oceanic horizontal and vertical velocity data are from the National Centers for Environmental Prediction (NCEP) global ocean data assimilation system (GODAS). The GODAS was developed using the Geophysical Fluid Dynamics Laboratory Modular Ocean Model version 3 (MOM.v3) and a three-dimensional variational data assimilation scheme (Behringer and Xue 2004). It has a horizontal resolution of $0.33^\circ \times 1.0^\circ$ (latitude × longitude) and 40 levels in the vertical direction. The SST data are from the Hadley Centre Sea-ice and Sea-surface Temperature Data Set Version 1 (HadISST; Rayner et al. 2003) dataset with a horizontal resolution of $1^\circ \times 1^\circ$. Data used in this paper cover the period from 1980 to 2018. In this study, anomalies in the reanalysis data are defined as the differences between negative and positive March Arctic TCO events, and the seasonal cycle of monthly data is removed by subtracting the climatological value in each month. Anomalies in modeling results refer to differences between low-ozone and high-ozone experiments.

2.2 General circulation model and numerical experiments

The NCAR Community Earth System Model (CESM) version 1.2.2 (Vertenstein et al. 2012) is used in this study. CESM is a coupled system, including components of atmosphere (CAM/WACCM), ocean, land, and sea ice. The atmospheric component used is the Whole Atmosphere Community Climate Model version 4 (WACCM4) (Marsh et al. 2013). The WACCM4 has detailed middle-atmosphere chemistry and a finite volume dynamical core, with 66 vertical levels extending from the surface to approximately 140 km. Its vertical resolution is approximately 1 km in the tropical tropopause and lower stratosphere layer. In this study, we disabled the interactive chemistry to study the effects of stratospheric ozone. The simulations presented in this study were performed at a horizontal resolution of $1.9^\circ \times 2.5^\circ$ (latitude × longitude) for the atmosphere and approximately the same for the ocean.

The original ozone data are ensemble mean ozone output from phase 5 of the Coupled Model Intercomparison (CMIP5) for the period from 1955 to 2005 (Taylor et al. 2012), and these data represent zonal mean ozone field. Experiments R1 and R2 with prescribed low- and high-ozone scenarios are performed to explore the influence of Arctic stratospheric ozone on the North Pacific surface ocean currents. The ozone forcing field is constructed from the CMIP5 ensemble mean ozone output, and the case of CESM1.2.2 used in this study is “B_1955-2005_WACCM_SC_CN”. In experiment R1/R2, March Arctic (60°–90° N) ozone concentrations of this forcing field are decreased/increased by 15%, and this ozone change is similar to those used in previous studies (e.g., Xie et al. 2018; Ma et al. 2019). To avoid discontinuity of ozone change in the vertical direction and be consistent with the TCO anomaly in the reanalysis dataset, the ozone is decreased/increased by 15% in the whole Arctic atmosphere (from surface to the top of the atmosphere) in the experiment R1/R2. However, the ozone change mainly occurs in the stratosphere, i.e., the difference in the Arctic (60°–90° N) ozone above 300 hPa between experiments R1 and R2 is 122.8 DU, while it is only 9.4 DU in the troposphere (300–1000 hPa). Moreover, Xie et al. (2008) reported that the effects of tropospheric ozone anomalies on the atmospheric temperature are rather small based on modelling results (their Fig. 3). Therefore, we can expect that the tropospheric differences between experiments R1 and R2 are mainly due to the stratospheric ozone changes.

2.3 Methods

Following previous studies (e.g., Lee et al. 2012; Chen et al. 2014, 2018), synoptic-scale eddy activity (also called the storm track activity; e.g., Anbauam and Nova 2014) is defined as the root mean square of the 2–8 day band-pass filtered GH using Lanczos filter. The feedback of synoptic eddies to the mean flow can be quantitatively estimated by the feedback
term in the GH tendency equation (e.g., Lau and Holopainen 1984; Lau 1988). Analogue to previous studies (e.g., Cai et al. 2007; Chen et al. 2015, 2020a), we only consider the eddy vorticity flux term since the eddy heat flux term is much smaller (e.g., Lau and Holopainen 1984; Lau and Nath 1991; Chen et al. 2015). The GH tendency due to the eddy vorticity flux forcing is expressed as follows (e.g., Lau 1988; Chen et al. 2020a):

$$F = \frac{g}{f} \nabla^{-2} \left[ -\nabla \cdot \left( V' \zeta' \right) \right]$$

(1)

where $g$ is the acceleration of gravity; $f$ denotes the Coriolis parameter; $V'$ represents synoptic-scale horizontal wind; $\zeta'$ is synoptic-scale vorticity and is calculated according to the synoptic-scale winds.

Following previous studies (e.g., Hoskins and Valdes 1990; Chen et al. 2020b), the atmospheric baroclinicity is represented by Eady growth rate. The Eady growth rate anomaly induced by zonal wind changes is calculated as follows:

$$\sigma_{BI} = 0.31 f |\partial u/\partial z| / N$$

(2)

Table 1 The positive and negative March Arctic (60°–90° N) TCO anomaly events based on 0.8 standard deviation are shown in the first and third columns, respectively

| Positive events | Arctic TCO anomaly | Negative events | Arctic TCO anomaly |
|-----------------|--------------------|-----------------|--------------------|
| 1980            | 34.8 DU            | 1990            | −32.0 DU           |
| 1984            | 21.5 DU            | 1992            | −28.2 DU           |
| 1989            | 37.6 DU            | 1993            | −48.0 DU           |
| 1999            | 31.2 DU            | 1995            | −46.8 DU           |
| 2001            | 26.2 DU            | 1996            | −46.8 DU           |
| 2010            | 35.6 DU            | 1997            | −64.7 DU           |
| 2018            | 29.9 DU            | 2000            | −31.3 DU           |
| 2020            | 24.4 DU            | 2011            | −64.3 DU           |

The March Arctic TCO anomalies (units: DU) of positive and negative TCO events are shown in the second and fourth columns, respectively. The ozone anomaly thresholds for negative and positive Arctic TCO events are −28.2 DU and +21.5 DU, respectively

where $f$ is the Coriolis parameter, $u$ is zonal wind, $N$ is the climatological buoyancy frequency.

Fig. 1 Differences in horizontal oceanic velocity (arrows; cm/s) in the North Pacific in April between negative and positive Arctic TCO anomaly events from 1980 to 2018 at the depth of a 5 m, b 15 m, c 25 m, d 35 m, e 45 m and f 55 m based on GODAS datasets. Colors denote magnitudes (cm/s) of the arrows. The dotted regions are statistically significant at the 95% confidence level according to Student’s $t$ test.
3 Observed variations in surface ocean currents in the North Pacific associated with March Arctic TCO anomalies

Figure 1 shows composite differences in horizontal oceanic velocity (0–55 m) in April between negative and positive March Arctic TCO events. Note that the linear trends of all data have been removed before composite analysis in this study. The events and Arctic TCO anomalies are listed in Table 1, and the results are not sensitive to the selections of Arctic TCO events (e.g., 1.0 standard deviation) (not shown). Figure 1 indicates that as the March Arctic TCO decreases, significant northwestward ocean current anomalies occur in the northern North Pacific (45°–55° N, 160° E–170° W) while southward ocean current anomalies appear in the central North Pacific (25°–35° N, 140° E–160° W) in April. Ocean current anomalies associated with ozone occur mainly in the surface (5 m, Fig. 1a), and anomalies at depths of 15–25 m are relatively weak and localized (Fig. 1b–c). Current anomalies are negligible below 35 m (Fig. 1d–f). Based on Fig. 1, hereafter, we focus on the ocean current anomaly averaged between 0 and 25 m, and the term “near-surface” refers to 0–25 m.

Figure 2 shows differences in the strength of ocean currents in the near-surface layer (0–25 m) between negative and positive Arctic TCO events. Figure 2a displays climatological distributions of ocean currents. Figure 2b, c show the strength of near-surface ocean currents during positive and negative Arctic TCO events, respectively. The anomalous northwestward current in the northern North Pacific (Fig. 1) associated with decreased ozone is out of phase with

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**Fig. 2**  
(a) Climatological ocean currents  
(b) Ocean currents (Positive TCO events)  
(c) Ocean currents (Negative TCO events)  
(d) Difference in current speed

**Fig. 2**  
(a) Climatological (1980–2018) distributions of horizontal oceanic velocity (arrows; cm/s) and its speed (colors; cm/s) in near-surface layer (0–25 m) in April based on the original GODAS dataset.  
(b, c) Distributions of horizontal oceanic velocity in April and its speed (0–25 m) during (b) positive and (c) negative Arctic TCO events.  
(d) Differences in current speed (cm/s) between negative and positive Arctic TCO events in near-surface layer. The dotted regions in (d) are statistically significant at the 95% confidence level according to Student’s t test.
the climatological southeastward current (Fig. 2a), leading to a weakening of the ocean current (color) in the northern North Pacific (40°–50° N, 150° E–150° W) during negative events (Fig. 2c) compared to those during positive events (Fig. 2b). However, the ozone-related anomalous southward current in the central North Pacific (Fig. 1) is in phase with
the climatological current (Fig. 2a), favoring an enhancement of ocean current in the central North Pacific (25°–35° N, 150° E–140° W) during negative events (Fig. 2c) compared to those during positive events (Fig. 2b). Figure 2d also indicates that the strength of near-surface ocean current is weakened/strengthened in the northern/central North Pacific due to the ozone-related current anomalies. Therefore, Figs. 1–2 suggest a close lead-lag relationship between the March Arctic TCO and the North Pacific surface ocean currents in April, implying that the Arctic TCO changes may affect the surface ocean currents.

To understand the relationship between ozone and ocean currents, Fig. 3 shows differences in GH and wind between negative and positive March Arctic TCO events. The decrease of Arctic TCO induces a decrease in shortwave heating (e.g., Ramaswamy et al. 1996; Hu et al. 2011; Ma et al. 2019) and thereby favors cooling and strengthening of the Arctic SPV (Figs. 3a–b). The intensified stratospheric circulation induces the tropospheric positive AO (+AO)-like anomaly (Fig. 3c, d) via stratosphere-troposphere dynamical coupling (e.g., Hoskins et al. 1985; Haynes et al. 1991; Ambaum and Hoskins 2002; Song and Robinson 2004). Figure 4 further displays ozone-related circulation anomalies over the North Pacific in April. There are negative GH anomalies (anomalous cyclonic circulation) over the central North Pacific and positive GH anomalies (anomalous anticyclonic circulation) over the northern North Pacific, resembling a negative North Pacific Oscillation (−NPO) anomaly.

Figure 5 shows the details of tropospheric circulation anomalies associated with ozone. The +AO-like anomaly reaches largest in late March (Fig. 5c). This +AO anomaly is accompanied by positive GH anomalies over high-latitude Asia (50°–70° N; Fig. 5c) and easterly anomalies over midlatitude Asia (35°–45° N, 60°–110° E; Fig. 6c) in late March, which persist and develop into early April (Fig. 6d).

Since the ozone-related +AO anomaly occurs mainly in
late March and the \(-NPO\) anomaly appears mainly in early-middle April (Fig. 5), we focus on late March, early April and middle April hereafter. Note that background westerlies exist over the eastern Asia-North Pacific regions at middle latitudes (Fig. 7a–c). The background westerly favors eastward momentum advection and thereby results in easterly accelerations over the eastern Asia-western North Pacific regions (30°–45° N, 120°–170° E; box regions in Fig. 7d) in late March, implying that the background westerly is one factor that favors the eastward extension of the ozone-related easterly anomalies from the eastern Asian to the western North Pacific. Figure 8 further shows the time evolutions of the ozone-related zonal wind anomalies at middle latitudes. Note that the easterly anomalies over eastern Asia in late March are located over approximate 35°–45° N (Fig. 6c), and the North Pacific easterly anomalies are located over approximate 40°–50° N in early-middle April (Fig. 6d, e). The ozone-related zonal wind anomalies averaged over 35°–45° N and 40°–50° N are both shown in Fig. 8. The ozone-related easterly anomalies over eastern Asia gradually extend eastward in late March (red thick arrows), and they extend into the North Pacific and induce easterly anomalies there in early April (Fig. 8). By this path, the effects of ozone reach the North Pacific. Additionally, the tropospheric (300–1000 hPa) background westerly averaged over the region of 40°–50° N, 120°–180° E is about 12.3 m/s in late March (Fig. 7a), meaning that the eastward traveling speed of the background westerly is about 13.5 degree per day over 40°–50° N and 120°–180° E.

In addition, note that the easterly anomalies over midlatitude Asia in late March persist and develop into early April (Fig. 6c, d). Thus, some processes may induce the development of the Asian easterly anomalies. Previous studies found that synoptic eddies have positive feedback on zonal wind changes (e.g., Gong et al. 2011; Chen et al. 2015). Hence, it is necessary to analyze the role of synoptic eddies in the development of the Asian easterly anomalies. Figures 9a–c show ozone-related anomalies in near surface temperatures.
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Previous studies (e.g., Broccoli et al. 2001; Cohen and Barlow 2005) reported that +AO anomaly (Fig. 5c) favors positive surface temperature anomalies over high-latitude Eurasia. Indeed, there are positive near surface temperature anomalies over high-latitude Asia (solid contours; Fig. 9a), favoring anomalous positive gradients of near surface temperatures over midlatitude Asia (colors; Fig. 9a) and implying the decrease of lower tropospheric baroclinicity over midlatitude Asia. The decrease of lower tropospheric baroclinicity tends to induce the decrease of synoptic eddies over midlatitude Asia (Fig. 9d) (e.g., Eady 1949; Chen et al. 2020b). Note that climatological westerlies exist over the eastern Asia (Fig. 7). Thus, the ozone-related easterly anomalies over the midlatitude Asia (Fig. 6c) mean a weakening of local westerlies, which induces a weakening of vertical shear of zonal wind and thereby also favors the decrease of synoptic eddies (Fig. 9d) (e.g., Chen et al. 2014). The decreased synoptic eddy further induces negative GH tendencies south of it (20°–50° N; Fig. 9g) via vorticity advection, consistent with previous studies (e.g., Gong et al. 2011; Zheng et al. 2021). Due to the negative GH tendencies induced by synoptic eddies, the negative GH anomalies over the subtropical Asia (25°–45° N) persist and develop into early April (Figs. 9j, k and 5c, d), which favor the persistence and development of the easterly anomalies over midlatitude Asia from late March to early April (Fig. 6c, d) by geostrophic wind relations. Therefore, the feedback of synoptic eddies is a factor that favors the persistence and development of the easterly anomalies over the midlatitude Asia from late March to early April, which may also contribute to the North Pacific easterly anomalies (Figs. 6d and 8).

When the ozone-related easterly anomaly reaches the midlatitude North Pacific (Fig. 6d, e), it favors positive/negative GH anomalies north/south of it directly via inducing negative/positive vorticity anomaly (−∂u/∂y) by zonal wind shear, like a −NPO anomaly. Meanwhile, the ozone-related easterly anomaly over midlatitude North Pacific also influences the −NPO anomaly via synoptic eddies: The easterly anomaly over midlatitude North Pacific (dashed contours; Fig. 10b, c) can weaken local atmospheric baroclinicity (colors; Fig. 10b, c) via weakening vertical shear of zonal wind (Eq. (2)). The weakening of atmospheric baroclinicity induces the decrease of synoptic eddy over the midlatitude North Pacific (Fig. 10e, f). The decreased synoptic eddy further results in positive GH tendencies north of it and negative GH tendencies south of it (Fig. 10h–i) via vorticity advection, resembling a −NPO and consistent with previous studies (e.g., Chen et al. 2014; Zheng et al. 2021). The −NPO-like GH tendencies in early-middle April (Figs. 10h–i) favor −NPO-like GH anomalies associated with ozone (Fig. 5d, e, Fig. 4), which further reinforce the easterly

![Fig. 6](image)

Same as Fig. 5 but for differences in zonal wind (m/s)
Fig. 7 The distributions of climatological (1980–2018) tropospheric (300–1000 hPa) horizontal wind (arrows; m/s) and its speed (colors; m/s) in (a) late March (March 21 to March 31), (b) early April (April 1 to April 10) and (c) middle April (April 11 to April 20). (d)–(f) Differences in the divergence ($1 \times 10^{-5}$ m s$^{-2}$) of zonal momentum advection induced by climatological westerlies (i.e., background westerlies) between negative and positive Arctic TCO events in (d) late March, (e) early April and (f) middle April. The divergence of zonal momentum advection induced by the background westerly is expressed as $\partial u' \bar{u} / \partial x$, where $u'$ and $\bar{u}$ denote anomalous and climatological zonal wind, respectively. Note that the divergence of zonal momentum advection in (d–f) has been multiplied by $-1$, and thus negative values denote easterly accelerations and vice versa. The data are from ERA-Interim. Dots (hatching) denote statistically significant anomalies at the 90% (95%) confidence level according to Student’s $t$ test.

Fig. 8 Time-longitude cross sections of differences in the tropospheric (300–1000 hPa) zonal wind (m/s) averaged over (a) 35°–45° N and (b) 40°–50° N between negative and positive Arctic TCO events based on ERA-Interim datasets. The ordinate and abscissa represent date (March 1 to April 30) and longitude (30° E–120° W), respectively. Heavy black lines represent the zero contours. Dots (hatching) denote statistically significant anomalies at the 90% (95%) confidence level according to Student’s $t$ test.
Surface ocean current variations in the North Pacific related to Arctic stratospheric ozone anomalies over the midlatitude North Pacific. Note that the synoptic eddies over the North Pacific are stronger than that over Asia (not shown), and the strength of synoptic eddy feedback on the mean flow perturbations is proportional to the strength of synoptic eddies (e.g., Jin 2010; Chen et al. 2015). Therefore, the ozone-related easterly anomalies over the North Pacific are larger in magnitude than that over the midlatitude Asia (Figs. 6, 8).

The atmospheric circulation anomalies can affect the near-surface ocean currents through direct friction of wind (e.g., Marshall and Plumb 1989; Huang 2010; Benoit and Beckers 2011). Moreover, Robert (2001) reported that the ocean surface currents adjust nearly instantaneously to changes in local wind forcing. Therefore, the ozone-related NPO anomaly (Fig. 4) may further influence the North Pacific surface ocean currents. Figure 11 displays divergence anomalies (color) in ocean currents (0–55 m) overlapped by horizontal oceanic velocity anomalies (arrows) associated with the ozone. There are a divergence anomaly in the central North Pacific (35°–45° N, 160° E–150° W) and a convergence anomaly in the northern North Pacific (50°–60° N, 160° E–170° W) in the near-surface layer (Fig. 11a–c), corresponding well to the negative GH anomaly over the central North Pacific and positive GH anomaly over the northern North Pacific.
The anomalous surface ocean currents flow from the divergence region to the convergence region (Fig. 11). The corresponding mechanism is easy to understand: The lower tropospheric cyclonic/anticyclonic circulation anomalies favour oceanic cyclonic/anticyclonic circulation anomalies in surface layer through direct friction of wind (Fig. 4d), which then deflect to the right due to the Coriolis force (Marshall and Plumb 1989) and thereby lead to a divergence/convergence of the surface current. This process explains why the anomalous surface ocean currents associated with ozone flow from the divergence region to the convergence region (Fig. 11). Figure 11 also indicates that the ozone-related anomalies
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in surface ocean currents and divergences occur mainly at the surface (5 m). Benoit and Beckers (2011) also reported that the direct effect of the wind on the ocean currents is limited to a thin layer (10 m or so) due to slightly viscous and strong rotation of seawater, consistent with our results.

Now a question is whether the ozone-related anomalies mentioned above are distinguished from that of the Arctic stratospheric final warmings (SFW). Based on the onset dates of the SFW during 1981–2016 from Bulter et al. (2019), only 8 of 36 SFW events happened in March while the rest (28 events) of the SFW events happened in April and May. Moreover, the current anomalies composited by Arctic TCO (Fig. 1) are similar to those composited based on Arctic TCO events with excluding years in which the SFW happened in March (not shown). These results imply that the ozone-related anomalies in the North Pacific are distinguished from that of SFW.

4 Simulated surface ocean current anomalies in the North Pacific related to ozone

Previous section suggests a close relationship between the Arctic TCO and the North Pacific surface ocean currents. In this section, we use numerical simulations to verify the results obtained from the reanalysis data. The model we used is CESM1.2.2, and the details of the model and experimental designs are provided in Sect. 2.2.

Figure 12 shows differences in GH and horizontal wind between experiments R1 (ozone decreased by 15%) and R2 (ozone increased by 15%). The decrease of Arctic stratospheric ozone favours negative GH anomalies over the central North Pacific (15°–35° N, 160° E–150° W) and positive GH anomalies over the northern North Pacific (40°–60° N, 160° E–150° W) (Fig. 12a, c), similar to a –NPO anomaly. Correspondingly, cyclonic/anticyclonic circulation anomalies occur over the central/northern North Pacific in the lower troposphere due to ozone (Fig. 12d). Although the simulated position of the

Fig. 11 Differences in horizontal oceanic velocity (arrows; cm/s) and its divergence (colors; $1 \times 10^{-8}$ s$^{-1}$) at the depth of a 5 m, b 15 m, c 25 m, d 35 m, e 45 m and f 55 m in April between negative and positive Arctic TCO anomaly events based on GODAS dataset. The dotted regions are statistically significant at the 95% confidence level according to Student’s $t$ test
NPO anomaly induced by ozone (Fig. 12) is slightly south (about 10°) of that composited by ozone in the reanalysis data, both their patterns are similar to –NPO. Figure 12 indicates that the decreased Arctic stratospheric ozone favors –NPO-like anomalies over the North Pacific in April.

Figure 13 displays differences in ocean currents (arrows) and their divergences (colors) at the depths of 0–55 m between experiments R1 and R2. There are anomalous northwestward currents associated with ozone in the northern North Pacific (30°–50° N, 170° E–150° W; Fig. 13a–c), which further lead to anomalous divergence in the near-surface current in the central North Pacific (20°–35° N, 170° E–140° W) and a convergence in the northern North Pacific (40°–55° N, 170° E–150° W). The position of the anomalous divergence and convergence of ocean currents (Fig. 13) corresponds to the position of the –NPO-like anomaly (Fig. 12c). Note that because the negative center of the NPO anomaly over the central North Pacific (15°–35° N, 160° E–150° W) is relatively weak (Fig. 12c), the anomalous southward current in the central North Pacific is also not prominent (Fig. 13). However, the ozone-related anomalies in the near-surface ocean currents and their divergences correspond well to the –NPO anomaly, supporting the mechanism by which the Arctic stratospheric ozone affects the surface (5 m) ocean currents in the North Pacific by the NPO anomaly.

Now a question is whether the ozone-related anomalies in the North Pacific in April come from the persistence of anomalies in preceding months (Jan, Feb and Mar). Thus, time evolutions of these anomalies in reanalysis data and numerical simulations are shown in Figs. 14 and 15.
respectively. Note that there is a tight relationship between month-to-month (Jan-Feb-Mar-Apr) variations in GH (colors) and surface ocean current divergence (colors) in both reanalysis data and modeling results (Figs. 14, 15), consistent with previous studies that the surface ocean currents adjust nearly instantaneously to changes in local wind forcing (e.g., Robert 2001). Figure 14 indicates that the ozone-related GH anomalies and corresponding surface current anomalies are weak and insignificant in preceding months (Jan, Feb and Mar). However, these anomalies are statistically significant in April (Figs. 14d, h), suggesting that the ozone-related tropospheric circulation and surface ocean current anomalies in April are not from the persistence of the anomalies in preceding months but due to the stratospheric anomalies. The modeling results also indicate that the ozone-related anomalies in the tropospheric GH and surface ocean current are not significant in Jan-Mar, while they are statistically significant in April (Fig. 15).

5 Mechanism by which the Arctic TCO influences the North Pacific SSTs

Preceding sections indicate that the Arctic TCO has a close relationship with the North Pacific surface ocean currents, which may further influence oceanic heat transport and SSTs. It is necessary to further investigate the effects of the ozone-related current anomalies on heat transport and SSTs in the North Pacific. Following Hall and Bryden (1982), the meridional heat transport in near-surface ocean is calculated by Eq. (3):

\[
H = \rho c_p \int_{Z_D}^{Z_W} \int_{X_E}^{X_W} \theta d\phi dz \tag{3}
\]

where \( \rho \) (1 \( \times 10^3 \) kg m\(^{-3}\)) is the density of the ocean; \( c_p \) (4 \( \times 10^3 \) J kg\(^{-1}\) K\(^{-1}\)) is the specific heat capacity of the ocean; \( \theta \) is the potential temperature of the ocean; \( Z_D \) is 25 m; \( X_E \) (140° W) and \( X_W \) (140° E) are the eastern and western boundaries of the North Pacific regions (140° E–140° W), respectively, in which there are significant ocean current anomalies associated with ozone (Fig. 1). Figure 16 shows differences in the meridional heat transport due to ocean currents between negative and positive Arctic TCO events.
The ozone-related ocean current anomalies induce anomalous northward heat transports in the northern North Pacific (45°–60° N) and southward heat transports in the central North Pacific (20°–35° N) (Fig. 16), which may further influence the North Pacific SSTs.

The contributions of oceanic currents, as well as surface heat flux, to ozone-related SSTAs are estimated based on Eq. (4) (Foltz et al. 2013; Fang and Yang 2016):

$$\frac{\partial T'}{\partial t} = -(u' \frac{\partial T'}{\partial x} + v' \frac{\partial T'}{\partial y}) - w' \frac{\partial T'}{\partial z} + \frac{Q'}{\rho c_p h} + \epsilon$$

where $T'$ is the climatological SST; primes denote the ozone-related oceanic temperature, oceanic three-dimensional velocity and surface heat flux anomalies; $h$ is 100 m, consistent with Fang and Yang (2016); and the residual $\epsilon$ is a combination of errors in the first three terms and other much smaller terms. The SST tendency includes that contributed by the horizontal oceanic temperature advection ($-(u' \frac{\partial T'}{\partial x} + v' \frac{\partial T'}{\partial y})$), oceanic vertical temperature advection ($-w' \frac{\partial T'}{\partial z}$), and surface heat flux ($\frac{Q'}{\rho c_p h}$). The surface heat flux includes surface sensible heat flux, latent heat flux, solar radiative flux and longwave radiative flux.

Figure 17 shows ozone-related SSTAs and SST tendency anomalies. There are negative SSTAs in the middle-western North Pacific and positive SSTAs in the northern and eastern North Pacific associated with decreased ozone (Fig. 17a), similar to a positive VM (+VM) anomaly. This ozone-related VM anomaly is consistent with previous studies that a decrease in Arctic stratospheric ozone in March favors a +VM anomaly in April (Xie et al. 2016; Xie et al. 2017a; Ma et al. 2019; Wang et al. 2020). Figure 17b–d show ozone-related SST tendencies contributed by surface heat flux, vertical oceanic temperature advection and horizontal oceanic temperature advection, respectively. The horizontal oceanic temperature advection anomaly contributes to the VM-like SSTAs in the middle-western North Pacific (Fig. 17d), indicating an important role of ozone-related horizontal ocean current anomaly in the North Pacific SST variations. The surface heat flux anomaly may also have a contribution to the VM-like SSTAs (Fig. 17b), and the contribution of oceanic vertical temperature advection anomaly is relatively small (Fig. 17c). Furthermore, the ocean current anomalies are divided into a meridional component (V) and a zonal component (U), and their contributions to the SST tendencies are shown in Figs. 17e and f, respectively. The contribution of meridional ocean current anomaly related to ozone is large (Fig. 17e), while the contribution of zonal current anomaly to the SSTAs is negligible (Fig. 17f). Figure 18 further shows the simulated SSTAs induced by ozone and contributions of ocean currents and surface heat flux to these SSTAs. There are positive SSTAs in the northern (35°–50° N, 150° E–130° W) and eastern (10°–30° N, 140°–120° W) North Pacific and negative SSTAs in the middle-western (20°–30° N, 120° E–140° W) North Pacific (Fig. 18a), also similar to a +VM anomaly and corresponding to the ozone-related –NPO anomaly (Fig. 12). Moreover, both surface heat flux anomalies and horizontal ocean current anomalies associated with ozone contribute to this VM anomaly while the contribution of oceanic vertical velocity anomaly is relatively small (Fig. 18). Therefore, Figs. 17–18 indicate that the ozone-related ocean current anomaly, especially its meridional component, plays an important role in the formation of the VM-like SSTAs associated with ozone.

Note that in the reanalysis data, there are two-way interactions between the Arctic SPV and stratospheric ozone (e.g., Calvo et al. 2015; Liu et al. 2020; Manney et al. 2020). Thus, the composite anomalies in ocean currents and SST in the reanalysis data are actually the combined impacts of the SPV and ozone, and thus they are relatively large (Figs. 14, 17). On the other hand, the simulated anomalies are mainly due to ozone since only the ozone is changed in the numerical experiments. Hence, it is reasonable that the simulated anomalies induced by ozone are smaller in magnitudes than that composited by ozone in the reanalysis data. However, the numerical experiments suggest that the effects of Arctic ozone on the tropospheric circulation and near-surface ocean in the North Pacific are statistically significant (Figs. 12, 13), indicating that the role of Arctic ozone in climate variations should be considered in the climate model.

6 Conclusions and discussions

Using reanalysis datasets and a coupled general circulation model (CESM1.2.2), the relationship between the Arctic TCO changes and near-surface (0–25 m) ocean current anomalies in the North Pacific is explored. It is found that as March Arctic TCO decreases, an anomalous northward/southward ocean current occurs in the northern/central North Pacific in April (Fig. 1). Due to the ozone-related ocean current anomalies, the near-surface ocean currents are weakened/strengthened in the northern/central North Pacific during negative Arctic TCO events, and vice versa (Fig. 2d).

Our analysis indicates that the decrease of March Arctic TCO favors an enhancement of Arctic stratospheric
Surface ocean current variations in the North Pacific related to Arctic stratospheric ozone

circulation (Fig. 3b), which extends into the troposphere (Fig. 3c) and induces tropospheric +AO anomaly (Figs. 3d, 5c) with an easterly anomaly over midlatitude eastern Asia in late March (Fig. 6c). The easterly anomaly over eastern Asia in late March further extends eastward and induces an easterly anomaly over the midlatitude North Pacific (Figs. 8 and 6d–e), which favors –NPO-like anomaly via zonal wind shear anomaly and constructive interactions between synoptic eddies and the mean flow in early-middle April (Fig. 10). The –NPO-like anomaly induced by the synoptic eddies (Figs. 10h–i) further reinforces the North Pacific easterly anomalies at middle latitudes. The wind anomaly associated with this –NPO forces anomalous northwestward/southward surface ocean currents in the northern/central North Pacific through direct frictions of wind and the Coriolis force, flowing from the divergence region to the convergence region (Fig. 11). Modeling results also support that the Arctic stratospheric ozone affects the North Pacific surface ocean currents (Fig. 13) through NPO anomalies over the North Pacific (Fig. 12). Moreover, the ozone-related horizontal current anomaly, especially its meridional component, contributes to VM-like SSTAs in the North Pacific in April (Figs. 17d–e and 18d–e).

The time scale and magnitude of the responses in surface ocean currents to the Arctic stratospheric ozone are consistent with previous studies. Regarding the time scale, Robert (2001) reported that the adjustment of the upper ocean proceeds much faster than that in the interior of ocean, and especially, the surface ocean currents adjust almost instantaneously to local wind forcing changes. Previous studies also found that monthly/seasonal mean atmospheric circulation anomaly affects near-surface ocean currents and SST in the same month/season (e.g., Di Lorenzo 2003; Chhak et al. 2009; Fang and Yang 2016). Thus, the time scale in these previous studies is consistent with that the April NPO anomaly associated with ozone affects surface ocean currents in April in this study. Regarding the magnitude, previous studies found that the Antarctic stratospheric ozone affects the Southern Ocean overturning circulation in the deep ocean and temperature in the upper ocean (e.g., Sigmond and Fyfe 2010; Solomon et al. 2015; WMO 2018). Several studies also suggested that the Arctic stratospheric ozone in March favors VM anomalies in April (e.g., Xie et al. 2016; Xie et al. 2017a; Ma et al. 2019). Our numerical experiments also imply that the Arctic stratospheric ozone inserts surface ocean current anomalies in the North Pacific (Fig. 13), consistent with previous studies that the stratospheric ozone variations are able to affect ocean currents and SST (e.g., Sigmond and Fyfe 2010; Bitz and Polvani 2012; WMO 2018).

This study considered the impacts of extreme ozone events and found that the climate effects of Arctic stratospheric ozone are significant (Figs. 12, 13). However, Harari et al. (2019) considered all ozone events and found that the effects of the stratospheric ozone on the North Pacific are not prominent. This may be related to different magnitudes of ozone change used in this study and in Harari et al. (2019). Previous studies also suggested that extreme ozone events induce significant tropospheric anomalies (e.g., Smith and

![Image](54x208 to 286x391)

**Fig. 15** a–d Differences in lower tropospheric (600–1000 hPa) geopotential height (colors; m) and horizontal wind (vectors; m/s) in a January, b February, c March and d April between experiments R1 (ozone decreased by 15%) and R2 (ozone increased by 15%). e–h Same as a–d but for surface (5 m) ocean current (vectors; cm/s) and its divergence (colors; $1 \times 10^{-8}$ s$^{-1}$). Dots (hatching) denote statistically significant anomalies at the 90% (95%) confidence level according to Student’s t test.

![Image](528x752)

**Fig. 16** Differences in meridional heat transport (units: PW, 1 PW = $1 \times 10^{15}$ W) in the North Pacific (averaged over 140° E–140° W) in near-surface layer (0–25 m) between negative and positive Arctic TCO events based on GODAS dataset. The bold lines denote statistically significant anomalies at the 95% confidence level.
Hence, future works are needed to address to what extent a certain ozone change induces the tropospheric anomalies. Additionally, the simulated NPO anomaly induced by ozone perturbations is located slightly south (~ 10°) than the NPO anomaly composited by ozone in reanalysis data, inducing the simulated oceanic anomalies to also locate south than that in the reanalysis data. This difference in the position of the ozone-related NPO anomaly may come from that the ozone forcing used in the model is zonally symmetric while it is zonally asymmetric in the reanalysis data, and may also be due to the internal variability of the climate system. In addition, the reanalysis includes changes in both the SPV and ozone while the model experiments only include the ozone change, which may also contribute to the position differences. However, the numerical experiments still support the mechanism proposed by this study, i.e., the Arctic ozone influences the North Pacific surface ocean currents by NPO anomalies, and the ozone-related current anomalies further contribute to VM-like SSTAs, which have not been reported by previous studies. In addition, it is interesting to explore the effects of Arctic stratospheric variability (including stratospheric ozone) on the North Pacific deep ocean in future studies.
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Declarations

Conflict of interest  The authors have no competing interests.

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Fig. 18  a Differences in sea surface temperatures (K) in April between experiments R1 (ozone decreased by 15%) and R2 (ozone increased by 15%). The heavy black line represents the zero contour to reflect regions with negative SSTAs. Dots (hatching) denote statistically significant anomalies at the 90% (95%) confidence level according to Student’s t test. b Differences in SST tendencies (1 × 10^{-7} \text{K s}^{-1}) contributed by surface heat flux between experiments R1 and R2. c–f Same as b but for those contributed by c vertical oceanic velocity, d horizontal oceanic velocity, e meridional component (V) of horizontal oceanic velocity and f zonal component (U) of horizontal oceanic velocity.
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