A dissection of the topographic effects from Eurasia and North America on the isentropic meridional mass circulation in Northern Winter

Yueyue Yu1,2 · Rongcai Ren1,2 · Xin Xia3 · Ruxue Liang1 · Jian Rao1

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Abstract
The topographic dynamical effect from Eurasia (EA_Topo) and North America (NA_Topo) on the winter isentropic meridional mass circulation (IMMC) is investigated using the WACCM. The independent effect of EA_Topo and that of NA_Topo, with the former much stronger, are both to strengthen the IMMC that is composed of the lower equatorward cold air branch (CB) and the upper poleward warm air branch in the extratropical troposphere (WB_TR) and stratosphere (WB_ST). Further investigation of the individual contributions from changes in stationary vs. transient and zonal-mean flow vs. waves reveals that, due to the topography-forced mass redistribution, changes in the low-level meridional pressure gradient force a zonal-mean counter-clockwise/clockwise meridional cell in the southern/northern side of topography. This weakens/strengthens the IMMC south/north of 30° N from the troposphere to lower stratosphere, acting as a dominant contributor to the IMMC changes south of 50° N. Meanwhile, the EA/NA_Topo-forced amplification of stationary waves constructively interacts with those determined by land-sea contrast, making the dominant/minor contributions to the strengthening of CB and WB_TR north of 50° N. The related increase in the upward wave propagation further dominates the WB_ST strengthening in the subpolar region. Meanwhile, transient eddy activities are depressed by EA/NA_Topo along with the weakened background westerly, which partly-offset/dominates-over the contribution from stationary flow in midlatitudes and subpolar region. The coexistence of the other topography (NA/EA_Topo) yields destructive mutual interferrence, which can weaken/offset the independent-EA/NA_Topo-forced meridional mass transport mainly via changing the zonal-mean as well as the downstream wave pattern of mass and meridional wind.

1 Introduction
The framework of Isentropic Meridional Mass Circulation (IMMC) was documented by Johnson and his collaborators (Johnson 1989; Cai and Shin 2014 and references therein). The IMMC represents a hemispheric cell which links the tropics to the poles and the troposphere to the stratosphere via the poleward warm air branch (WB) in the upper troposphere (WB_TR) and stratosphere (WB_ST) and the equatorward cold air branch in the lower troposphere (CB). In the IMMC framework, a semi-Lagrangian and quantitative understanding has been gained on the nature of the general atmospheric circulation (Chen 2013; Cai and Shin 2014), on the meridional heat and energy transport (Doos and Nilsson 2011; Pauluis et al. 2011; Yamada and Pauluis 2015; Wu et al. 2019), and on the stratosphere–troposphere coupling associated with oscillations of the polar vortex and jet (or the Northern Annular Modes) (Cai and Ren 2006, 2007; Yu et al. 2014; Yu et al. 2015c, 2018a, b, c; Yu and Ren 2019).
Moreover, the intensity of CB across the polar circle is found to be closely related to winter cold air outbreaks in the mid-latitudes (Iwasaki and Mochizuki 2012; Iwasaki et al. 2014; Shoji et al. 2014). The WB always varies in phase with the CB and is recognized as a useful upper-level precursor of cold air outbreaks (Yu et al. 2015a, b, c; Cai et al. 2016). Therefore, investigating what dominates the intensity of IMMC and the underlying mechanisms can provide insight on many widely known weather systems and climate patterns of interest.

The Northern Hemispheric IMMC is always stronger and wider in winter (Johnson 1989; Cai and Shin 2014), when its driving force—the vertically westward tilted waves (Johnson 1989)—is stronger and more active. Among the wave forcings in the extratropics, planetary-scale waves make the dominant contributions (Yu et al. 2018b). One of the major sources of planetary waves is large-scale topography (Held et al. 2002; Smagorinsky 1953; Nigam et al. 1986, 1988; Kasahara and Washington 1971; Manabe and Terpstra 1974; Lin 1982; Jacqmin and Lindzen 1985; Chen and Trenberth 1988; Sato et al. 2009). In the Northern Hemisphere (NH), the topography over Eurasia (EA_Topo hereafter), mainly composed of the Tibetan Plateau, Iranian Plateau, and Mongolia Plateau, and the Rocky Mountains in the west of North America (NA_Topo hereafter) play a critical role in intensifying stationary planetary waves (Ringler and Cook 1995; Yanai et al. 2006) and enhancing their upward propagation into the stratosphere (Zou et al. 1991a, b; Luo et al. 1985; Ding 1992; Taguchi and Yoden 2002; Yanai et al. 2006; Gerber and Polvani 2009). The dynamical effect of EA_Topo is much stronger than that of NA_Topo in determining the intensity and spatial pattern of planetary waves (e.g., Held 1983; White et al. 2017). Park et al. (2013) pointed out that the stationary waves forced by the Tibetan Plateau can constructively interact with the preexisting stationary waves determined by land-sea contrast (Chang 2009), thus making critical contributions to strengthen the poleward eddy heat transport over East Asian and Eastern Pacific region. Accompanied with enhanced stationary waves by topography, the transient eddy kinetic energy is reduced by topography (Manabe and Terpstra 1974; Yu and Hartmann 1995; Son et al. 2009), and its compensation of the poleward energy transport can be quite robust (Trenberth and Stepaniak 2003).

Most of the abovementioned studies focused on the topographic effects of EA_Topo or NA_Topo with the other topography coexistent. The independent topographic effect and possible mutual interference between them have not been well understood yet. A few linear and nonlinear theoretical model studies (Luo et al. 1985; Held and Ting 1990; Valdes and Hoskins 1991) suggested that either EA_Topo or NA_Topo may influence the topographic forcing of the other on the atmospheric circulation by affecting the background zonal flow across the NH. Ren et al. (2019) and Xia et al. (2019) recently carried out a series of numerical experiments with the comprehensive Whole Atmosphere Community Climate Model (WACCM) forced under different topographic conditions, and investigated the independent (with the other topography nonexistent), the dependent (with the other topography present), and the joint effects, as well as the mutual interference between EA_Topo and NA_Topo in modulating the northern winter westerly jet and stratospheric circulation. They found that the independent effects of EA_Topo and NA_Topo are both to enhance the upward wave propagation and weaken the stratospheric polar vortex and polar jet. Since independent effects tend to be destructively interfered due to the coexistence of the other topography. Especially, the EA_Topo’s interference can even offset the weakening effect of NA_Topo on the stratospheric polar vortex (Ren et al. 2019). They provided a primary explanation for the destructive mutual interference based on the stationary westerly momentum responses in the troposphere in the upstream region of EA_Topo and NA_Topo, and hence the weakened stationary wave responses due to the coexistence of the other topography.

As a follow-up study of Ren et al. (2019), this study aims to first demonstrate the various topographic effects on IMMC, whose stratospheric part is related to the stratospheric polar vortex reported in Ren et al. (2019), and then to provide a physical understanding of the topographic effects by performing a decomposition analysis on the IMMC responses to topographic forcing. The isentropic analysis in the framework of IMMC can provide a clearer picture of the dynamical role of topography in changing the atmospheric circulation. Firstly, based on the surface potential temperature (mostly below 290 K) in the spatial ranges of both EA_Topo and NA_Topo (Fig. S1), the response of the CB (i.e., the equatorward branch under 280 ~ 290 K) can represent the direct topographic effect, as long as there are no additional diabatic processes; Secondly, the topography-forced air mass redistribution quantified by isentropic analysis makes it easier to understand the IMMC changes from changes in meridional gradient of low-level pressure and temperature, and thus modulation on the zonal-mean flow and planetary waves. Therefore, via numerical experiments, isentropic analysis, and decomposition analysis, we try to answer two questions: (i) What are the changes in the three branches of IMMC in winter due to the various topographic forcing, including the forcing of independent and dependent EA_Topo and NA_Topo, the joint forcing, as well as the mutual interference between EA_Topo and NA_Topo? (ii) How does the topography modulate the stationary vs. transient and zonal-mean flow vs. waves and their individual contributions to the total changes in IMMC?

This paper is organized as follows. Section 2 describes the data, model, and numerical experiments, and the calculation
scheme for the IMMC-related variables. In Sect. 3, we make deliberate parallel comparisons of the simulated fields of isentropic meridional mass fluxes to demonstrate the independent effects of EA_Topo and NA_Topo and their mutual interference on each branch of the IMMC in the NH. Then a decomposition analysis is performed to diagnose the contributions to the IMMC responses to topographic forcing from stationary vs. transient and zonal-mean flow vs. waves. Section 4 discusses the physical processes related to stationary zonal-mean flow and waves, mainly via which the topography takes effect. Conclusions and discussions about the topographic effect on the transient flow are provided in Sect. 5.

2 Data and methods

2.1 Model and experiment design

Version 4 of the WACCMM (WACC4M) is used in this study. WACC4M is a “high top” model, with the model top reaching ~ 150 km (66 levels), and the horizontal resolution is 1.9° × 2.5° (latitude × longitude) (Marsh et al. 2013). We grid ~ 150 km, and the horizontal resolution is WACCM4 is a “high top” model, with the model top reaching ~ 150 km (66 levels), and the horizontal resolution is .

2.2 Calculation scheme of the IMMC-related variables

The same methods as in Yu et al. (2015a, b) (also see Pauluis et al. (2008) and Cai and Shin (2014)) are adopted to obtain the IMMC-related variables. The variable fields used for calculation include daily surface air temperature (SAT), surface pressure (Ps), surface meridional wind (v), and three-dimensional (3D) air temperature (T), meridional wind (v), zonal wind (u), and geopotential height (h) fields. The 3D and surface potential temperature (θ and θn) fields are derived from the daily fields of T, SAT, and Ps. The datasets used in this study include not only the daily data obtained from the model experiments, but also those derived from the six-hourly ERA-Interim data from January 1979 to December 2016 (ECMWF 2012; Simmons et al. 2007; Dee et al. 2011) on 1.5° latitude × 1.5° longitude grids and at 37 pressure levels from 1000 to 1 hPa.

2.2.1 Isentropic layer mass

Fifteen potential temperature surfaces θn (n: 1–15) are preselected: 260, 270, 280, 290, 300, 315, 330, 350, 370, 400, 450, 550, 650, 850, and 1200 K. All variables are defined in the 14 layers between the θn and θn+1 surfaces, plus two additional layers: one is the lowest layer, which accounts for all air mass between the ground and 260 K, in case the surface potential temperature at the grid is below 260 K; and the other is the top layer, which accounts for all air mass above 1200 K. We use the bottom isentropic surface of each layer, i.e., θn (n = 1, 14), and 250 K and 1200 K respectively for the lowest and top layers in referencing the variables defined in these isentropic layers.

The daily potential temperature and wind fields are interpolated onto 200 equally spaced sigma (σ) levels from 1 to 0. The air mass between two adjacent sigma surfaces per unit area is: \( m_\sigma = \frac{\Delta \sigma}{g} P \), where \( g \) is the gravitational constant and \( \Delta \sigma = 1/200 \). We then derive the mass between two isentropic levels at each grid point on day \( t \) as

\[ \text{mass between isentropic levels} \]
where $\lambda$ is the longitude and $\phi$ is the latitude; $Y(x, x_1, x_2) = 1$ if $x_1 \leq x < x_2$, and otherwise $Y(x, x_1, x_2) = 0$; and $A(\Phi)$ is the area of a grid box centered at the grid point. Via summing up the $M$ along a given latitudinal band $\phi$, we obtain the zonally integrated mass, and denote it using brackets, i.e., $[M]$. We further calculate the vertical sum of $[M]$ above each isentropic surface, which is proportional to the pressure at isentropic levels.

The winter mean $[M]$ pattern derived from the model CTL experiment largely resembles that in ERA-Interim (Fig. 1a), indicating the capability of WACCM4 in reproducing the characteristics of the meridional distribution of cold and warm air mass.

### 2.2.2 Isentropic meridional mass fluxes

We derive the meridional mass transport (MF) in the layer between two isentropic levels, $\Theta_n$ and $\Theta_{n+1}$, at each grid point on day $t$ in units of kg s$^{-1}$ as

$$M(\lambda, \phi, \Theta_n, t) = A(\phi) \cdot \int_0^1 m_n(\lambda, \phi, \sigma, t) \cdot Y(\theta(\lambda, \phi, \sigma, t), \Theta_n, \Theta_{n+1}) d\sigma$$

(1)

where $\lambda$ is the longitude and $\phi$ is the latitude; $Y(x, x_1, x_2) = 1$ if $x_1 \leq x < x_2$, and otherwise $Y(x, x_1, x_2) = 0$; and $A(\Phi)$ is the area of a grid box centered at the grid point. Via summing up the $M$ along a given latitudinal band $\phi$, we obtain the zonally integrated mass, and denote it using brackets, i.e., $[M]$. We further calculate the vertical sum of $[M]$ above each isentropic surface, which is proportional to the pressure at isentropic levels.

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$$MF(\lambda, \phi, \Theta_n, t) = A(\phi) \cdot \int_0^1 m_n(\lambda, \phi, \sigma, t) \cdot v(\lambda, \phi, \sigma, t) \cdot Y(\theta(\lambda, \phi, \sigma, t), \Theta_n, \Theta_{n+1}) d\sigma$$

(2)

We further calculate and denote the zonally integrated MF as $[MF]$. Seen from the winter (DJF) climatology of $[MF]$ derived from ERA-Interim data (shadings in Fig. 1b), the winter mean $[MF]$ is positive/negative in the upper isentropic layers but negative/positive in the lower layers in the NH/Southern Hemisphere, indicating an isentropic meridional overturning mass circulation (IMMC) consisting of a poleward WB and an equatorward CB. The WB is composed of the portion in the middle and upper troposphere (denoted as “WB_TR”) and that in the stratosphere (denoted as “WB_ST”). The winter mean $[MF]$ pattern derived from the model CTL experiment (contours in Fig. 1b) largely resembles that derived from ERA-Interim Reanalysis data, except that the separating level between CB and WB where the strongest baroclinic instability exists in the model looks relatively colder than in the observation. Nevertheless, this systematic bias does not affect the results representing the dynamical roles of large-scale topography on the IMMC in this study.

### 2.2.3 Separation of the warm and cold branches of IMMC

The boundary level that separates the CB and WB of opposite directions at latitude $\phi$ and at day $t$ is identified by searching for the isentropic level $\Theta_{n*}(\phi, t)$ such that the vertical sum of $[MF]$ for all $n < n*$ or $\Theta_n < \Theta_{n*}$ reaches its maximum negative value. As shown by the green curve in Fig. 1b, the isentropic surface of climatological mean $\Theta_{n*}(\phi, t)$ derived from CTL experiment effectively separates the CB and WB and decreases with increasing latitudes. The climatological mean $\Theta_{n*}(\phi, t)$ derived from the other three experiments are highly consistent.

Based on the tropopause temperature and pressure fields of NCEP/NCAR Reanalysis 1 in DJFs from 1979 to 2011, we derive the winter mean potential temperature near tropopause and then find the closest predefined isentropic level $\Theta_b$ (black curve in Fig. 1b) to indicate the approximate boundary level between WB_TR and WB_ST, denoted as $\Theta_{n*}(\phi, t)$.

With the two boundary isentropic levels obtained, the intensity of CB at each latitude can be measured by the vertical sum of $[MF]$ below $\Theta_{n*}(\phi, t)$, the intensity of WB_TR can be measured by the vertical sum of $[MF]$ from $\Theta_{n*}(\phi, t)$ to $\Theta_1$.
to $\Theta_{n+1}(\phi, t)$, and the intensity of WB_ST is measured by the vertical sum of $[\text{MF}]$ from $\Theta_n \star (\phi, t)$ to the top of the atmosphere.

### 2.3 Spatial and temporal decomposition of the isentropic layer mass and its transport

A Lorenz circulation resolution method (Lorenz 1967) is utilized, in which any variable at sigma levels, $X(\lambda, \phi, \sigma, t)$, can be decomposed into zonal-mean and wave components. The isentropic mass and meridional mass flux driven by zonal-mean flow can be derived as

$$M_{\text{zonal}}(\phi, \Theta_n, t) = A(\phi) \cdot \int_0^1 \{m_\sigma\}(\phi, \sigma, t) \cdot Y(\theta)(\phi, \sigma, t, \Theta_n, \Theta_{n+1}) d\sigma$$

The MF driven by the total stationary flow (MF$^S$) and its zonal-mean part (MF$^S_{\text{zonal}}$) can be respectively derived as

$$\text{MF}^S(\lambda, \phi, \Theta_n) = \frac{\bar{\text{V}}(\lambda, \phi, \Theta_n, t) \cdot \bar{\text{M}}(\lambda, \phi, \Theta_n, t)}{R \Delta \phi}$$

$$\text{MF}^S_{\text{zonal}}(\phi, \Theta_n) = \frac{\bar{\text{V}}_{\text{zonal}}(\phi, \Theta_n, t) \cdot \bar{\text{M}}_{\text{zonal}}(\phi, \Theta_n, t)}{R \Delta \phi}$$

where $(\lambda, \phi, \Theta_n)$ indicates the stationary part, $\bar{\text{V}}$ denotes the 21 day running mean, and $(\lambda, \phi, \Theta_n)$ denotes the long-term climatological mean. The MF driven by stationary waves (MF$^S_{\text{wave}}(\lambda, \phi, \Theta_n)$) can be removed by obtaining the MF$^S_{\text{zonal}}(\phi, \Theta_n)$ from MF$^S(\lambda, \phi, \Theta_n)$. The 21 day running mean is chosen following Park et al. (2013) to include quasi-stationary waves such as Pacific-North America (PNA) pattern as stationary wave part, thus the transient eddies include mostly storm tracks.

The MF driven by transient flow (MF$^T(\phi, \Theta_n)$) is equal to either the residual of the MF after removing the MF$^S$ or $\frac{\bar{\text{V}}(\lambda, \phi, \Theta_n) \cdot \bar{\text{M}}(\lambda, \phi, \Theta_n, t)}{R \Delta \phi}$, where $\Delta$ is the difference of daily fields from the 21 day running mean fields. Similarly, we can derive the MF due to the transient zonal-mean flow (MF$^T_{\text{zonal}}(\phi, \Theta_n)$) from the MF$^S_{\text{zonal}}$ and MF$^S_{\text{wave}}$. Finally, the MF driven by transient eddies (MF$^T_{\text{wave}}(\lambda, \phi, \Theta_n)$) can be obtained by removing the MF$^T_{\text{zonal}}$ from MF$^T$.

### 3 Topography-forced changes in the NH IMMC

#### 3.1 Zonally integrated meridional mass fluxes

We begin by comparing the features of IMMC among experiments. $\Delta$ is used to denote the differences between them to present the change of a specific variable due to the topographic forcing of EA_Topo and NA_Topo. Displayed in Fig. 2 are the latitudinal patterns of the topography-forced \(\Delta[\text{MF}]\) in various isentropic layers.

It can be seen from Fig. 2a that, the independent effect of EA_Topo is to significantly strengthen the tropospheric
portion of IMMC in the midlatitudes and subpolar region, as manifested by the positive/negative Δ[MF] in the layers where the WB_TR/CB lies, but to weaken it significantly in the low latitudes. Meanwhile, the independent effect of EA_Topo is to strengthen the stratospheric portion of IMMC (WB_ST) across the entire NH, as seen from the positive Δ[MF] covering the NH latitudes. The maximum strengthening of poleward [MF] is around 45° N in the lower stratosphere but shifts poleward with increasing height. The independent EA_Topo forced strengthening of IMMC can be seen more clearly from the vertical integral of Δ[MF] in isentropic layers respectively within the CB, WB_TR, and WB_ST (Figs. 3a, 4a, 5a).

The independent NA_Topo forced changes of CB and WB_TR show remarkable difference with those forced by independent EA_Topo around 45–70° N, where the NA_Topo turns to weaken the CB and WB_TR (Figs. 2c, 3c, and 4c). Though the hemispheric scale WB_ST is still strengthened by the independent forcing of NA_Topo, the magnitudes of positive Δ[MF] are only ~1/4 of those forced by the independent EA_Topo (Figs. 2c and 5c).

The topographic effect of mutual interference between EA_Topo and NA_Topo (Figs. 2f, 3f, 4f, and 5f) generally counteracts with their independent effects, characterized by a weakening of both the WB_TR and CB in 20–50° N and the WB_ST in the extratropics. In the troposphere, the magnitudes of Δ[MF] forced by mutual interference are comparable to those forced by independent EA_Topo, and even larger than those forced by independent NA_Topo. As a result, the EA_Topo-forced strengthening of CB and WB_TR becomes less significant and limited within a narrower latitude band when the NA_Topo coexists (see dependent effect of EA_Topo in Figs. 2f, 3b, and 4b), while the less remarkable strengthening effect of independent NA_Topo near 35–50° N is even offset to a significant weakening effect of CB and WB_TR in the entire extratropics when the EA_Topo coexists (Figs. 2d, 3d, and 4d). In the stratosphere, the magnitudes of Δ[MF] due to mutual interference are about 1/3 of those forced by independent EA_Topo but again much larger than those forced by independent NA_Topo. The independent-EA_Topo forced strengthening of WB_ST is thus slightly suppressed when the NA_Topo coexists (Fig. 5b), while that forced by independent NA_Topo is totally offset to a weakening of the WB_ST when the EA_Topo coexists (Fig. 5d).

In addition, because of the mutual interference in changing IMMC in both the troposphere and stratosphere, the Δ[MF] related to the joint effect of EA_Topo and NA_Topo (Figs. 2e, 3e, 4e, and 5e) is not a linear summation of their independent effects. The higher resemblance of the isentropic-latitudinal pattern of the joint effect to that of the EA_Topo’s independent suggests the dominant role of EA_Topo in changing the IMMC.

Considering that the dependent as well as the joint effects of EA_Topo and NA_Topo can be inferred from their independent effects and mutual interference, and changes of WB_TR are highly coupled with changes of CB [as already demonstrated in Yu et al. (2015a) and confirmed by the overall out-of-phase changes of [MF] in the two branches shown in Figs. 2–4], our investigation in the following sections will mainly focus on the topographic effects of independent EA_Topo and independent NA_Topo and their mutual interference in changing the CB and WB_ST in the NH.

### 3.2 Spatial–temporal decomposition of the IMMC responses to topography

To figure out individual contributions to the topography-forced IMMC changes, respectively from stationary and transient flow, both of which are composed of zonal-mean flow and waves, we conduct a Lorenz spatial–temporal resolution analysis on the total Δ[MF] within CB (Fig. 3) and WB_ST (Fig. 5).

We start with investigating the independent effects of EA_Topo and NA_Topo. It is seen from Figs. 3a and 5a that the stationary component (Δ[MFS]) of the CB response to both the independent forcing of EA_Topo and NA_Topo exhibits negative/positive values in the mid–high/low latitudes. The Δ[MFS] within WB_ST exhibits positive values across the entire NH but larger magnitudes in mid–high latitudes. The Δ[MFS] forced by independent NA_Topo has smaller magnitudes compared with that forced by independent EA_Topo. This indicates the positive contributions of independent topography-forced stationary flow changes to the significant strengthening of IMMC in the extratropical troposphere and stratosphere, with the EA_Topo’s effect stronger than NA_Topo’s. A closer look at the stationary zonal-mean (Δ[MFSzonal]) and stationary wave (Δ[MFSwave]) components yields that, in the region south of 50° N, the independent-EA_Topo-forced Δ[MFS] in both the troposphere and stratosphere is dominated by changes in zonal-mean flow; while north of 50°N, the strengthening of Δ[MFS] is mainly contributed from changes in the stationary waves. Seen from Figs. 3c and 5c, the topographic effect on both stationary zonal-mean and wave components are weaker associated with the independent forcing of NA_Topo.
than that of EA_Topo, but the weakening of the topographic effect is more severe for the wave component ($\Delta[MFS_{\text{wave}}]$) than the zonal-mean component ($\Delta[MFS_{\text{zonal}}]$). As a result, the independent-NA_Topo-forced $\Delta[MFS]$ is dominated by $\Delta[MFS_{\text{zonal}}]$ in the entire NH, whereas the $\Delta[MFS_{\text{wave}}]$ makes minor positive contributions in the region north of 45°N. The Independent-topography-forced transient component of IMMC changes ($\Delta[MFT]$), however, is always opposite to the stationary component ($\Delta[MFS]$). This supports the strong compensation between stationary waves and transient eddies in their contributions to the meridional heat fluxes reported by Trenberth and Stepaniak (2003). We further find that, the $\Delta[MFT]$ is almost fully contributed by its transient eddy component ($\Delta[MFT_{\text{wave}}]$). The relative importance of $\Delta[MFS]$ and $\Delta[MFT]$ in determining the $\Delta[MF]$ depends on various factors including topography forcing, the IMMC branches, and the latitudes. To be specific, the Independent-EA_Topo-forced $\Delta[MFS]$ is always strong enough to overwhelm the $\Delta[MFT]$ in the extratropics, dominating the changes of both CB and WB_ST (Figs. 3a and 5a). However, the independent-NA_Topo-forced changes in transient eddy component plays a dominant role in changing the CB north of 50° N (Fig. 3c).

The $\Delta[MF]$ forced by mutual interference is dominated by its stationary component ($\Delta[MFS]$), which is nearly out-of-phase with the independent-EA_Topo-forced $\Delta[MFS]$ but

![Fig. 3](image-url)
A dissection of the topographic effects from Eurasia and North America on the isentropic...

with smaller magnitudes (Figs. 3f and 5f). The weakening of CB in 20–50° N dominated by $\Delta[MF^S]$ is mainly contributed by the $\Delta[MF^S_{zonal}]$ component, while the positive $\Delta[MF^S_{wave}]$ in 20–40° N also contributes. As to the WB_ST, $\Delta[MF^S]$ is dominantly contributed by the $\Delta[MF^S_{zonal}]$ south of 50° N, but by the $\Delta[MF^S_{wave}]$ in high latitudes. This also indicates the growing/decreasing importance of the contributions from the mutual-interference-forced changes in stationary waves/zonal-mean flow from the troposphere to stratosphere. Meanwhile, the transient eddies related to mutual interference lead to positive $\Delta[MF^T_{wave}]$ in the CB layer around 40° N and 65° N but negative $\Delta[MF^T_{wave}]$ in the WB_ST layer north of 50° N. This implies an enhancement of the destructive effect of mutual interference on the IMMC changes by independent-topography-forced transient eddies in these latitude bands.

We summarize in Table 1 the main features of the independent-topography-forced IMMC changes in CB and WB_ST, and that related to the mutual interference between EA_Topo and NA_Topo, as well as the relative contributions from the stationary vs. transient, and from the zonal-mean flow vs. wave components.
4 On the stationary component of the topography-forced IMMC changes

4.1 Troposphere

4.1.1 Stationary zonal-mean flow

As abovementioned, one of the major approaches for the topography to affect the tropospheric portion of IMMC is by modifying the stationary zonal-mean flow ($\Delta [MFS_{zonal}]$). We can see a high consistency in the $\Delta [MFS_{zonal}]$ (Figs. 3, 4, 5) with the zonal-mean meridional wind velocity and mass stream function at pressure levels (Fig. 6), namely the strengthening/weakening of the poleward WB_TR and equatorward CB in the mid–high/low latitudes corresponds to a topography-forced clockwise/counterclockwise zonal-mean meridional cell in the troposphere below 150 hPa. Next we try to figure out how the topography forces such a zonal-mean meridional cell.

Displayed in Figs. 7a and c are the independent-topography-forced changes in the zonal integrated isentropic air mass and its vertical sum above, which is proportional to the isentropic pressure. It is seen that due to the existence of the EA_Topo and NA_Topo, there exists significant loss of air mass in the latitude ranges of topography but gain of air mass to the northern and southern sides in the lower isentropic layers. Meanwhile, the positive air mass changes in the stratosphere above 315 K north of 40° N.

Fig. 5 As in Fig. 3, but for the WB_ST
Table 1 Summary of the main changes in CB and WB_ST of IMMC forced by independent EA and NA topography and their mutual interference. *indicates the component that makes dominant contributions

| Topographic forcing | Topography-forced changes in IMMC | Topography-forced changes in decomposed components and their contributions to IMMC changes |
|---------------------|----------------------------------|------------------------------------------------------------------------------------------|
|                     | Branch Main facts                | Component Main facts                                                                 |
| Independent EA      | CB Weakening south of 30° N       | Stationary zonal-mean flow* Weakening south of 30° N and strengthening in 30–70° N;     |
|                     |                                 | Dominates the changes south of 50° N                                                   |
|                     | Strengthening in 30–70° N        | Stationary wave* Strengthening in 15–80° N;                                            |
|                     |                                 | Dominates the changes north of 50° N                                                   |
|                     | WB_ST Strengthening in the NH (mainly in the mid and high latitudes) | Transient eddy Weakening in 15–80° N                                                   |
| Independent NA      | CB Weakening in 10–35° N          | Stationary zonal-mean flow* Weakening/strengthening in 10–35° N/35–70° N;               |
|                     |                                 | Dominates the region south of 50° N                                                    |
|                     | Slight strengthening in 35–50° N  | Stationary wave Slight strengthening in 15–70° N                                       |
|                     | Weakening in 50–70° N            | Transient eddy* Weakening in 15–70° N                                                   |
|                     |                                 | Dominates the changes north of 50° N                                                   |
|                     | WB_ST Strengthening in NH (mainly in the mid and high latitudes) | Stationary zonal-mean flow* Strengthening south of 70° N;                               |
|                     |                                 | Dominates the region south of 65° N                                                    |
|                     |                                 | Stationary wave Slight strengthening north of 50° N;                                    |
|                     |                                 | Dominates the changes north of 50° N                                                    |
|                     |                                 | Transient eddy Weakening in 40–75° N                                                    |
| Mutual interference | CB Strengthening/weakening in 5–20°N/20–50°N | Stationary zonal-mean flow* Strengthening/weakening in 5–20° N/20–50° N;               |
|                     |                                 | Dominates the region south of 50° N                                                    |
|                     | Slight strengthening/weakening in 50–65° N/65–75° N | Stationary wave* Weakening/strengthening in 20–40° N/50–65° N;                       |
|                     |                                 | Dominates the region 50–65° N and make important contributions in 20–40° N             |
|                     |                                 | Transient eddy* Slight weakening around 35–50° N and 50–75° N;                          |
|                     |                                 | Dominates the region north of 65–75° N                                                  |
|                     | WB_ST Weakening North of 30° N    | Stationary zonal-mean flow* Weakening/strengthening in 30–60°N/north of 60° N;        |
|                     |                                 | Dominates the region south of 50° N                                                    |
|                     |                                 | Stationary wave* Weakening north of 50° N but slight strengthening in 30–50° N;         |
|                     |                                 | Dominates the changes north of 50° N                                                    |
|                     |                                 | Transient eddy Slight weakening/strengthening in north of 50° N/30–50° N;              |
|                     |                                 | Also positively contributes to the changes north of 50° N                                |

collaboratively enhance the total mass gain in the northern side of topography (the mass changes in the stratosphere are mainly resulted from WB_ST changes driven by stationary waves, which will be discussed in Sect. 4.2). This causes a significant decrease of surface and low-level pressure in the latitude ranges of topography, but a significant increase of pressure to the northern and southern sides. The resultant low-level meridional pressure gradient favors southward/ northward zonal-mean meridional wind in the northern/southern side of mountain below 700 hPa (Fig. 6a and c). The convergence effect (around 30° N for EA_Topo and 35° N for NA_Topo) is compensated by the upward motion
Fig. 6 Topographic effects on the winter climatology of the zonal-mean meridional wind (units: m s$^{-1}$, contours) and mass stream function from the top (units: $10^{10}$ kg m s$^{-1}$, shadings) at pressure levels in the NH. Dotted areas and thickened contours mark the 95% confidence level.
and an opposite pattern of zonal-mean meridional wind in the mid-to-upper troposphere (200–300 hPa). Thus, a counterclockwise/clockwise zonal-mean meridional cell in the low/mid-to-high latitudes is formed due to the existence of independent EA_Topo or NA_Topo, contributing to weaken/strengthen the CB and WB_TR in the low/mid-to-high latitudes, as have been presented by $\Delta[M_{\text{wave}}^S]$ in Figs. 3a, c 4a, c.

The different effects of NA_Topo and EA_Topo, namely the much weaker and slightly northward shifted low-level meridional gradient of pressure (cp. Fig. 7a and c) and thus the forced zonal-mean meridional cell (cp. Fig. 6a and c), can be attributed to the smaller volume as well as the northward location of NA_Topo than EA_Topo. It should be noted that, the much less mass gain in the extratropical stratosphere forced by independent NA_Topo also contributes to the weaker meridional gradient of low-level pressure.

As shown in Fig. 7f, the mutual interference acts to enhance/weaken the low-level air mass export from the topography to its northern/southern side. This helps to increase/decrease the surface pressure in the northern sides (45°–60° N) and southern sides (20°–45° N). However, slightly above the surface, the column air mass or pressure changes are dominated by an increase/decrease in the air mass above 315 K in the region south/north of 40° N forced by mutual interference, resulting in maximum positive/negative values of the vertical sum of [M] or pressure at around 20°/60° N. Therefore, the mutual interference leads to a poleward/equatorward pressure gradient and thus the zonal-mean meridional wind in 20–60°N/0–20° N at low levels (contours in Figs. 6f and 7f). Together with the opposite-signed changes in the zonal-mean meridional wind in the upper layer, a counterclockwise meridional cell in the midlatitudes (20–60° N) and a clockwise meridional cell in the low latitudes (0–20° N) are formed due to the mutual interference, which is almost opposite to the meridional cell forced by independent EA_Topo and NA_Topo. The magnitudes of the mutual-interference-forced zonal-mean meridional cell are smaller than those of the EA_Topo-forced meridional cell, but comparable with or even larger than the NA_Topo-forced meridional cell. Therefore, the EA_Topo-forced clockwise meridional cell in the midlatitudes is slightly weakened when the NA_Topo coexists (Fig. 6b); the NA-forced clockwise meridional cell in the midlatitudes is offset and almost reversed while the counterclockwise meridional cell in the subtropics is strengthened and extends northward (Fig. 6d). This explains the different responses of the stationary zonal-mean component of CB and WB_TR to the topographic forcing of EA_Topo and NA_Topo between without and with the other topography coexistent (cp. Figs. 3b, 4b with Figs. 3a and 4a and cp. Figures 3d and 4d with Figs. 3c and 4c).

### 4.1.2 Stationary waves

Stationary waves have crucial contributions to the strengthening of tropospheric branches of IMMC in the midlatitudes and subpolar region. To understand how the existence of topography and the mutual interference change the stationary waves as well as their resultant meridional mass transport in the troposphere, we first focus on the CB layer, where the air mass and its transport are directly affected by the uplifting, obstructing, and deflections of both mountains.

The stationary wave pattern of the total air mass within the CB is equivalent to the long-term average of the vertical integral of $M_{\text{wave}}^S(\lambda, \phi, \Theta_n, t)$ below the boundary level between cold and warm branches of the IMMC $(\Theta_n(\phi, t))$. The stationary wave component of the mass-weighted meridional wind in the CB ($V_{\text{CB}}^S(\lambda, \phi)$) can be obtained based on Eqs. (7–9) by substituting the isentropic layer mass and meridional mass fluxes with the vertical integral of mass and meridional mass fluxes within the CB layer. The stationary part of mass-weighted mean zonal wind velocity within the CB ($U_{\text{CB}}^S(\lambda, \phi)$) can be derived in the similar fashion to the $V_{\text{CB}}^S$ by substituting the $v$ with $u$. Then, the MR driven by independent-topography-forced stationary waves within the CB can be approximately linearized as,

$$
\Delta \left( \sum \frac{M_{\text{wave}}^S(\lambda, \phi, \Theta_n, t)}{\partial t} \right) \approx \Delta M_{\text{CB}}^S \cdot \Delta V_{\text{CB}}^S + M_{\text{CB}}^S \cdot \Delta V_{\text{CB}}^S + \Delta M_{\text{CB}}^S \cdot V_{\text{CB}}^S \tag{12}
$$

where the first term at the righthand side of Eq. (12) denotes the spatial coherence between the $\Delta V_{\text{CB}}^S$ (Fig. 8c and g) and $\Delta M_{\text{CB}}^S$ (Fig. 9c and g) due to the independent topographic effect, the second term indicates the spatial coherence between the independent-topography-forced $\Delta V_{\text{CB}}^S$ and the preexisting $M_{\text{CB}}^S$ associated with land-sea contrast as shown in noEA&noNA experiment (Fig. 9a), and the third term indicates the spatial coherence between the preexisting $V_{\text{CB}}^S$ (Fig. 8a) and the independent-topography-forced $\Delta M_{\text{CB}}^S$.

Next let us briefly discuss the Eq. (12) term by term, which are displayed in Fig. 10a–c, to understand the independent effect of EA_Topo on the stationary wave component of the CB that is shown in Fig. 10d:

i) The first term ($\Delta M_{\text{CB}}^S \cdot \Delta V_{\text{CB}}^S$) tends to strengthen the equatorward CB in the extratropics except a narrow latitude band around 40° N. Seen from Fig. 8c, the mid-latitude westerly within the CB layer is weakened most severely in the upstream region of EA_Topo (90° W–90° E, 45°–70° N). Such weakening is accompanied by the strengthening of southerly/northerly in the latitudes north/south of 40° N on the

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windward side of EA_Topo and the strengthening of westerly to both the north and south of EA_Topo. This presents the mid-latitude westerly flow being blocked, deflected, and split into two branches due to the existence of EA_Topo. On the lee side, there is significantly strengthened northerly from the polar region to the southeastern China and southerly from lower latitudes. The downstream effects can be seen from the alternating positive and negative values of \( \Delta V_{CB}^{S} \) farther downstream, covering the entire mid-latitude circles. As to the independent effect of EA_Topo on the air mass distribution, we see from Fig. 9c that around 30° N, where the peak of EA_Topo mainly locates, there is a significant reduction of \( M_{CB}^{S} \) in the mountain range but an increase in the other longitude bands. Besides, the EA_Topo-forced \( \Delta M_{CB}^{S} \) tends to have an opposite sign to the \( \Delta V_{CB}^{S} \), with a slight westward shift relative to the \( \Delta V_{CB}^{S} \). This is because stronger northerlies always bring more cold air mass while stronger southerlies bring more/less warm/cold air mass in the midlatitudes, where the westerly flow is weakened due to the existence of EA_Topo. Therefore, the independent-EA_Topo-forced \( \Delta M_{CB}^{S} \cdot \Delta V_{CB}^{S} \) tends to be mostly dominated by negative values over the extratropical Eurasia (Fig. 10a), implying a strengthening of the equatorward mass transport within the CB. However, in the region around 40° N and 90°–120° E, the low-level air mass loss is severe because of the uplifting of the mountain and the \( \Delta V_{CB}^{S} \) is significantly equatorward there. The other region with significantly positive \( \Delta M_{CB}^{S} \cdot \Delta V_{CB}^{S} \) is the western north Pacific region. The positive \( \Delta M_{CB}^{S} \cdot \Delta V_{CB}^{S} \) in these two regions helps weaken the CB around 40° N.

The second term (\( M_{CB}^{S} \cdot \Delta V_{CB}^{S} \)) is negative over the midlatitudes and subpolar region of East Asia and Europe respectively upstream and downstream of EA_Topo, playing a dominant role in strengthening the CB in the extratropics (Fig. 10b). This is because there is a significant out-of-phase patterns of the independent-EA_Topo-forced \( \Delta V_{CB}^{S} \) (Fig. 8c) and the preexisting \( M_{CB}^{S} \) that has larger magnitudes than \( \Delta M_{CB}^{S} \) (Fig. 9a) over the Eurasian continent.

The third term (\( \Delta M_{CB}^{S} \cdot V_{CB}^{S} \)) also helps to strengthen the CB in the midlatitudes, indicated by an out-of-phase pattern of independent-EA_Topo-forced \( \Delta M_{CB}^{S} \) (Fig. 9c) and preexisting \( V_{CB}^{S} \) (Fig. 8a) on the east coast of Eurasia and some regions in the western hemisphere but with smaller magnitudes than \( M_{CB}^{S} \cdot \Delta V_{CB}^{S} \). However, as the term \( \Delta M_{CB}^{S} \cdot \Delta V_{CB}^{S} \), the preexisting equatorward wind brings severely decreased cold air mass over the mountain ranges (90°–120° E, 30°–50° N), which offsets the negative values of \( \Delta M_{CB}^{S} \cdot V_{CB}^{S} \) over the rest longitudes around 40° N.

Therefore, the \( M_{CB}^{S} \cdot \Delta V_{CB}^{S} \) which represents the constructively interaction of the independent-EA_Topo-forced stationary wave pattern of meridional wind with the stationary wave pattern of cold air mass determined by land-sea thermal contrast, makes the largest contributions to strengthen the stationary wave component of CB in the extratropics, while the other two terms involving the independent-EA_Topo-forced cold air mass changes (\( \Delta M_{CB}^{S} \cdot \Delta V_{CB}^{S} \) and \( \Delta M_{CB}^{S} \cdot V_{CB}^{S} \)) can explain the minimum strengthening of CB around 40° N.

Similar story can be told for the independent effect of NA_Topo (Figs. 8g, 9g, and 10e–h). But the magnitudes of independent-NA_Topo-forced \( \Delta V_{CB}^{S} \) and \( \Delta M_{CB}^{S} \) are much smaller than those forced by EA_Topo. In addition, the independent-NA_Topo-forced \( \Delta V_{CB}^{S} \) is more local than the EA_Topo, manifested by the alternating positive and negative values of \( \Delta V_{CB}^{S} \) covering only 1/3 of the latitude band (furthest impact to 60° E). Therefore, the strengthening of CB by independent-NA_Topo-forced stationary waves is less significant compared to the independent effect of EA_Topo (Fig. 3c).

Next let us look at the mutual interference between the EA_Topo and NA_Topo. Recall that the stationary wave changes due the mutual-interference are to weaken the CB in 20°–50° N but slightly strengthen it in 50°–65° N. Now the \( \Delta V_{CB}^{S} \) and \( \Delta M_{CB}^{S} \) in Eq. (12) represent the changes forced by the dependent topography (Figs. 8f, h, 9f, and h), while the preexisting fields. \( M_{CB}^{S} \) and \( V_{CB}^{S} \), are associated with land-sea contrast with the coexistence of the other topography (Figs. 8d and 9d for EA_Topo’s dependent effect and Figs. 8b and 9b for NA_Topo’s dependent effect). Each term in Eq. (12) is displayed in Fig. 11.

Firstly, we will address how the EA_Topo severely interferes with the dynamical effect of NA_Topo on the intensity of CB driven by stationary waves. As we all know, the EA_Topo is characterized by its highest altitudes over the world but relatively narrower meridional scale. With the existence of EA_Topo, the massive air in the lower isentropic layers replaced by the mountain has to spread out to other longitudes (Fig. 9c), leading to a significant increase of the preexisting \( M_{CB}^{S} \) around 30°–45° N in the upstream region of NA_Topo (cp. Fig. 9b and a). Such increased amount of air mass approaching the windward side of NA_Topo favors a stronger deflection effect of NA_Topo, as indicated by a
significantly strengthened southerly/northerly north/south of 45° N near the western boundary of NA_Topo that is accompanied with a strengthened northerly/southerly north/south of 30° N on the lee side of NA_Topo (cp. Fig. 8h and g). We see a farther extended stationary wave pattern of both ΔV_CB^S wave and ΔM_CB^S wave forced by NA_Topo when EA_Topo coexists, compared to that forced by independent NA_Topo. Such strengthened meridional wind exhibits an overall out-of-phase relationship with the preexisting M_CB^S wave in the downstream region of NA_Topo, thus it acts as a dominant contributor to the strengthening of CB in 40–65° N (Fig. 11f). The mutual-interference-forced weakening of CB in 20–40° N, however, is mainly contributed by the positive V_CB^S wave • ΔM_CB^S wave over the midlatitude regions of East Pacific and North America. It is because the more air mass approaching the windward side of NA_Topo because of the coexistence of EA_Topo allows a much significant deduction of air mass over the spatial range of NA_Topo (cp. Figure 9h with 9 g or their difference shown in Fig. 9i). Accompanied with this additional mass loss
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Fig. 10 The total and linearized stationary-wave component of MF changes within the CB layer (units: \(10^9 \text{ kg s}^{-1}\)) forced by the independent EA and NA. a \(\Delta M_{CB}^{\text{wave}} \cdot \Delta V_{CB}^{\text{wave}}\): the independent-EA-forced wave pattern of meridional stationary wind multiplied by the independent-EA-forced cold air mass (left panel) and its zonal integral (right panel). b \(M_{CB}^{\text{wave}} \cdot \Delta V_{CB}^{\text{wave}}\): the independent-EA-forced wave pattern of meridional stationary wind multiplied by the preexisting wave pattern of cold air mass associated with the land-sea contrast (left panel) and its zonal integral (right panel). c \(\Delta M_{CB}^{\text{wave}} \cdot V_{CB}^{\text{wave}}\): the independent-EA-forced wave pattern of cold air mass multiplied by the preexisting wave pattern of meridional stationary wind associated with the land-sea contrast (left panel) and its zonal integral (right panel). d The sum of the three terms in panels (a–c) (shadings in the left panel) and its zonal integral (bars in the right panel), overlaid by the vertical integral of \(M_{CB}^{\text{wave}}\) within the CB layer (contours in the left panel) and its zonal integral (green curves in the right panel). e–h are the same as (a–d), but for the independent effect of NA. Green contours indicate the location of EA/NA_Topo

(negative \(\Delta M_{CB}^{\text{wave}}\) ) are the strong preexisting northerlies (negative values of \(V_{CB}^{\text{wave}}\) shown in Fig. 8b). This results in large positive values of \(V_{CB}^{\text{wave}} \cdot \Delta M_{CB}^{\text{wave}}\), making dominant contributions to the weakening of the equatorward CB in 20–40° N (Fig. 11g).

The interference of NA_Topo with the dynamical effect of EA_Topo in changing the stationary waves in the CB layer, however, is quite weak. Compared to EA_Topo, NA_Topo has much lower altitudes and a much larger meridional scale. The air mass redistribution by NA_Topo covers almost all the NH latitudes, but with much smaller magnitudes compared to that by EA_Topo, particularly those around 30–45° N (cp. Figure 9g and Fig. 9c). Correspondingly, the EA_Topo-forced changes in the wind fields exhibit trivial difference between with and without the coexistence of NA_Topo (cp. Figures 8c and f). In regard of this, we skip the detailed discussions on the NA_Topo’s interference with EA_Topo in modulating CB via changing stationary waves, though terms in Eq. (12) presenting the EA_Topo’s dependent effect when NA_Topo coexists, are still displayed in Figs. 11a–d.

4.2 Stratosphere

Unlike the tropospheric portion of IMMC that is directly modified by the topography via mechanical processes, the stratospheric portion is mainly modulated via the topography-forced quasi-stationary planetary waves that can propagate upward. Following Zhang et al. (2013), we first calculated the daily amplitude of quasi-stationary waves at
pressure levels as the root mean square of the zonal deviations of 21-day running mean fields of geopotential height, and then obtained the long-term mean wave amplitude on isentropic surfaces after linear interpolation based on the potential temperature at pressure levels.

It can be seen from Fig. 12a that the independent effect of EA_Topo is to increase the amplitudes of stationary waves in the midlatitudes in lower isentropic layers. This can be explained by the pattern coherence of the preexisting $\Delta M_{\text{CB}}$ wave (Fig. 9a) with the independent-EA_Topo-forced $\Delta M_{\text{CB}}$ wave (Fig. 9c) over the mid-latitude Eurasian continent and Western Pacific region. Such amplified signals of large-scale stationary planetary waves enhance upward wave propagation into the upper stratosphere following the poleward wave guide (Huang and Gambo 1981, 1982; Chen et al. 2003, 2005). This is consistent with the independent-EA_Topo-forced strengthened upward EP fluxes reported in Ren et al. (2019). The stationary waves with larger amplitudes strengthen the poleward mass transport ($\Delta[M_{\text{wave}}]$), making dominant contributions to strengthen the WB_ST north of $50^\circ$ N (Fig. 5a). Note that the independent-EA_Topo-forced maximum increase in the wave amplitude shifts poleward with height, which is highly consistent with the high latitude waveguide (Huang and Gambo 1981, 1982). This explains the similar poleward shifting feature found in the latitude bands with positive values of $\Delta[M]$ at stratospheric levels (Fig. 2a). Such changes in the meridional mass transport dominantly lead to positive/negative changes of stratospheric air mass in the northern/southern side of the maximum strengthening of $\Delta[M]$ (Fig. 7a), which plays an important role in determining the low-level pressure as mentioned in the previous Sect. 4.1.1.

In the region south of $50^\circ$ N, however, the topography-forced zonal-mean stationary meridional wind plays a dominant role in changing the WB_ST (Fig. 5a). The latitude band of $30^\circ$–$50^\circ$ N with positive values of vertical integral of $\Delta[M_{\text{zonal}}]$ in the WB_ST layer (Fig. 5a) is consistent with the topography-forced zonal-mean northward wind in the lower stratosphere (Fig. 6a). We can tell that such northward zonal-mean wind is affected by the topography-forced zonal-mean meridional cell in the troposphere, since the independent-EA_Topo-forced zonal-mean northward wind in $30^\circ$–$50^\circ$ N extends upward to 50 hPa. It is also noted that the zonal-mean flow in the upper stratosphere (above 50 hPa) is poleward in the midlatitudes but equatorward in the high latitudes (Fig. 6a). This is because the maximum

![Fig. 11](image-url)
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Fig. 12  As in Fig. 2, but for the differences in stationary wave amplitudes (units: m)
intensification of poleward mass transport by the EA_Topo-forced stationary waves locates near the polar circle, which implies local divergence/convergence to its southern/northern side. This requires poleward/equatorward mass transport from higher/lower latitudes to compensate the local mass and accompanied pressure changes. Nevertheless, such wave-driven changes in the zonal-mean meridional cell in the upper stratosphere does not dominate the changes in the contributions from forced zonal-mean flow to the entire WB_ST because of the rareness of air mass there.

The same story can be said for the topographic effects of NA_Topo on the WB_ST. The independent-NA-forced ΔM_CB^S pattern (Fig. 9c) is in phase with the preexisting M_CB^S wave (Fig. 9a) over the midlatitude region of North America, which leads to larger wave amplitudes in the lower isentropic layers and stronger upward propagation into the stratosphere (Fig. 12c). The positive changes in wave amplitude forced by independent NA_Topo are much smaller than that forced by independent EA_Topo, which corresponds to much weaker positive changes in Δ[MF^S] within the WB_ST in the midlatitudes (Fig. 5c) as well as the related mass changes in the stratospheric layers (Fig. 7c). The less gain of stratospheric mass in the higher latitudes due to independent NA_Topo also contributes to the smaller equatorward pressure gradient at low levels thus consequently zonal-mean meridional cell forced by NA_Topo than EA_Topo, as we have discussed above. Possibly because of the weaker intensity of the NA_Topo-forced zonal-mean meridional cell, the altitude that the NA_Topo-forced zonal-mean northward wind in midlatitudes can extend to is much lower than the EA_Topo-forced one (cp. Figures 6c and a). This explains why the NA_Topo-forced strengthening of WB_ST dominated by zonal-mean flow in low and middle latitudes is also much weaker than the EA_Topo’s effect (cp. Δ[MF^S] zonal in Figs. 5c and a).

The mutual interference weakens the in-phase coherence of the preexisting M_CB^S wave pattern with the topography-forced ΔM_CB^S wave mainly in the mid-latitude regions of central Asia and the eastern North America (Fig. 9f), resulting in a less amplification of the stationary waves (Fig. 12f) as well as the related Δ[MF^S] in the midlatitudes and subpolar region. The Δ[MF^S] causes a Δ[M] pattern almost opposite to that forced by independent EA_Topo, which dominates the reversed meridional gradient of low-level pressure (Fig. 7f). As a result, the mutual-interference-forced zonal-mean meridional cell is almost the opposite to that forced by independent EA_Topo (Fig. 6f). And in addition, the equatorward zonal-mean meridional wind can also extend to the lower stratosphere, acting to weaken the stationary zonal-mean component of WB_ST. Therefore, the effect of mutual interference counteracts with the independent effect of EA_Topo and NA_Topo via modifying both the stationary zonal-mean flow in the midlatitudes as well as stationary waves in the subpolar region (Fig. 5f). The effect of mutual interference is weaker than the effect of independent EA_Topo on the stratospheric stationary flow, but strong enough to overwhelm the effect of independent NA_Topo, explaining the dependent effects of EA_Topo and NA_Topo in changing the stationary component of WB_ST when the other topography coexists (Figs. 5b and d).

5 Summary and discussions

5.1 Summary

This study investigates the dynamical effects of EA_Topo and NA_Topo in changing the Isentropic Meridional Mass Circulation in both the troposphere and stratosphere in the NH winter, which is physically linked to global mass and energy transport, the polar vortex variability, cold air outbreaks, etc. Via performing a series of numerical experiments with the stratosphere-resolving WACCM, we examine not only the independent dynamical effects of EA_Topo and NA_Topo but also their mutual interference, which further yields their dependent effects when the other topography coexists stably and their joint effect. A Lorenz circulation resolution method is then applied to diagnose the individual contributions from changes in stationary vs. transient and zonal-mean flow vs. waves forced by various topographic forcing to the IMMC changes. Finally, the physical processes related to stationary flow mainly via which the topography takes effect are investigated.

A summary in Table 1 shows that the independent effects of both EA_Topo and NA_Topo are to strengthen the equatorward cold and poleward warm branch of the IMMC in the troposphere (CB and WB_TR) in the midlatitudes but to weaken them in the low latitudes. The independent-NA-forced changes are much weaker than the independent-EA_Topo-forced changes, in terms of both the magnitude and spatial span. Such changes of CB and WB_TR south of 50° N are dominated by contributions from the topography-forced stationary zonal-mean meridional cell, while those north of 50° N are dominated by contributions from the topography-forced waves. The wave contributions are mainly from stationary waves for the EA_Topo’s effect but from transient eddies for the NA_Topo’s effect. For the stratospheric warm branch of IMMC (WB_ST) across the entire NH, the strengthening effects of EA_Topo and NA_Topo are mainly contributed from the stationary zonal-mean flow in the low and middle latitudes (south of 50° N/65° N) but from the stationary waves in the high latitudes. When the other topography coexists, the mutual interference between EA_Topo and NA_Topo is always destructive, which acts to weaken the independent-EA_Topo-forced effect but completely offset the independent-NA_Topo-forced effect.
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We further investigated how the existence of EA_Topo and NA_Topo modifies the stationary zonal-mean flow and stationary waves from the troposphere to stratosphere (as summarized in Fig. 13). Firstly, the uplifting of EA_Topo dominantly causes a decrease in the low-level pressure over the spatial ranges of topography but an increase in the pressure both to the north and to the south. This topography-forced meridional gradient of zonal-mean pressure favors a clockwise/counterclockwise cell to the north/south of the topography, which yields a strengthening/weakening of the CB and WB_TR in the middle/low latitudes. The independent-EA_Topo-forced zonal-mean poleward flow in the upper troposphere can extend to the lower stratosphere, dominantly contributing to the strengthening of the WB_ST in the same latitude bands. Secondly, the blocking and deflecting effect of EA_Topo generates strong stationary waves in the troposphere, via redistributing the surrounding air mass. The topography-forced wave patterns of meridional flow and air mass changes constructively interact with the preexisting wave patterns determined by land-sea contrast, contributing to the strengthening of the equatorward CB and poleward WB_TR in the extratropics. The topography-forced amplification of stationary planetary waves can propagate upward into the stratosphere, leading to a strengthening of the poleward WB_ST in the subpolar region. And the stratospheric mass exchange between lower and higher latitudes dominated by WB_ST changes in the subpolar region can enhance the topographic effect on the meridional gradient of low-level pressure and thus the zonal-mean meridional cell. All the abovementioned processes in changing the IMMC can be said for NA_Topo, except that the NA_Topo-forced changes are much weaker.

The destructive interference of one topography on the other is, on the one hand, mainly via redistributing the air mass and changing the meridional flow in the downstream region of the topography that is also the upstream region of the other topography, and consequently weakening the large-scale stationary waves as well as their driven meridional mass fluxes in the CB and WB_TR in the midlatitudes and subpolar region. On the other hand, the weakened large-scale stationary waves forced by the mutual interference can propagate upward to the stratosphere, weakening the WB_ST in the subpolar region. The weakened WB_ST in the subpolar region further decreases/increases the upper-level air mass in the mid-high/low latitudes, which dominates the meridional gradient of low-level pressure. Thus, the mutual interference forces a counterclockwise zonal-mean meridional cell in the midlatitudes from the troposphere to lower stratosphere, explaining the weakening of IMMC in the midlatitudes contributed by zonal-mean meridional flow due to mutual interference.

In the framework of isentropic meridional mass circulation, the results of this study help provide a comprehensive and clearer picture of the dynamical role of topography in changing the atmospheric circulation. A step further from Ren et al. (2019), we found that the topographic effects are achieved not only via modifying waves as many previous studies demonstrated (e.g., Yanai et al. 2006; Ding 1992; Park et al. 2013), but also via changing the zonal mean flow. The topographic effects on both the tropospheric circulation and the stratospheric circulation as well as their possible interaction, and the respective roles of stationary flow and transient eddies are also revealed. Investigation on the air...
mass and winds makes the underlying processes linking the surface topography and atmospheric circulation more physically direct to understand.

5.2 Discussions on the dynamical role of EA_Top and NA_Top in changing the transient eddy activities

It is still not answered yet, how the EA_Top and NA_Top change the transient eddies that also play an important role in changing the IMMC in the extratropics, particularly for the independent and dependent effects of NA_Top. A previous study by Park et al. (2013) shows that the early growth rate (EGR), which is widely used to estimate transient eddy activity from the mean (Lindzen and Farrell 1980), is dominated by the EA_Top-forced zonal wind shear. Since the independent effects of both EA_Top and NA_Top are to slow down the westerlies in the midlatitudes in both the troposphere and stratosphere (Ren et al. 2019), the EGR as well as the transient eddy activities are expected to be decreased by both EA_Top and NA_Top. Our primary analysis has confirmed this conjecture for the independent effects of EA_Top and NA_Top (see Figs. S3a and S3c), and the larger magnitudes of the decrease in the independent-EA_Top-forced EGR than that of the independent-NA_Top-forced EGR are also consistent with the more significant independent-EA_Top-forced weakening of the IMMC attributed to transient eddies (cp. Figures 3a, c 4a, c 5a, c).

However, when considering the mutual interference, a clear inconsistency between the EGR and IMMC changes related to transient eddies (cp. Fig. S3f with Figs. 3f, 4f, and Sf) can be found in the subpolar stratosphere. Inconsistency between EGR and meridional heat flux is also found by Park et al. (2013). They stated that this could be relevant to the nonlinear interaction between stationary waves and transients including the upstream eddy seeding effect on downstream transients (Penny et al. 2010; Son et al. 2009) and barotropic wind shear on transients (Harnik and Chang 2004). Further investigation is needed to figure out how the EA_Top and NA_Top changes the transient eddies and the resultant IMMC.

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Data availability The ERA-Interim datasets used in this work are available from the ECMWF (http://www.ecmwf.int). The NCEP/NCAR Reanalysis 1 datasets are available from the NOAA (https://www.esrl.noaa.gov/psd/datasets/data.ncep.reanalysis.tropopause.html).

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