Orography and the Boreal Winter Stratosphere: the Importance of the Mongolian mountains

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Key Points:

• The Rossby refractive index can provide a positive feedback on stratospheric zonal wind changes via altered wave propagation pathways
• The Mongolian mountains have twice the impact on stratospheric flow than the Tibetan plateau and Himalayas
• Sudden stratospheric warmings dramatically reduce in frequency without the Mongolian mountains

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Abstract

The impact of mountains on stratospheric circulation is explored using the Whole Atmosphere Community Climate Model. The ‘Mongolian mountains’ decrease the boreal winter stratospheric jet strength by ~1/3, and increase the frequency of major sudden stratospheric warmings (SSWs) from 0.08 $\text{year}^{-1}$ to the observed 0.60 $\text{year}^{-1}$. These changes are twice the magnitude of the impacts of the Tibetan plateau and Himalayas. Consistent with the decrease in the zonal jet, there is enhanced Eliassen-Palm flux convergence; this is predominantly from changes in wave propagation pathways through changes to the upper troposphere circulation, not from an increased amplitude of planetary waves reaching the stratosphere. The Mongolian mountains have the greater impact on upper tropospheric circulation due to their meridional location. The Rocky mountains have no significant impact on the stratospheric jet. Changes in wave propagation in response to the Mongolian mountains are similar to those associated with major SSW events in observations.

1 Introduction

Radiative cooling during the long polar night leads to a strong equator-pole temperature gradient in the winter stratosphere. In thermal wind balance with this temperature gradient exists a strong westerly jet, the stratospheric polar vortex. The structure and dynamics of the polar vortex are vital in setting the distributions of trace gases such as ozone. Observational [e.g., Baldwin and Dunkerton, 1999, 2001] and modeling studies [Polvani and Kushner, 2002; Kushner and Polvani, 2004; Gerber et al., 2009; Sheshadri et al., 2015] have established that variability in the polar vortex can influence the troposphere, all the way to the Earth’s surface, on timescales relevant to both weather and climate.

Major sudden stratospheric warming events (SSWs), in which the westerlies in the vortex reverse and the stratospheric polar temperature warms rapidly, occur about every other winter in the Northern Hemisphere [e.g. Matsuno, 1971; Charlton and Polvani, 2007]. Only one such event has occurred in the observational record in the Southern Hemisphere [e.g. Newman and Nash, 2005], where the vortex is colder and stronger than its Arctic counterpart [e.g., Waugh and Polvani, 2010]. These hemispheric differences are thought to be due to inter-hemispheric differences in the amplitude of planetary waves generated in the troposphere. Sources of such planetary-scale waves include topography [Charney and Eliassen, 1949], land-sea heating contrasts [Smagorinsky, 1953], and the
nonlinear interactions of synoptic-scale eddies [Scinocca and Haynes, 1998]. Only the
longest (planetary-scale) waves can propagate into the winter stratosphere [Charney and
Drazin, 1961], where they break, perturbing the flow away from radiative equilibrium.

Previous studies have explored the impact of the location, wavenumber, and am-
plitude of idealized orography on the variability of stratospheric circulation [Gerber and
Polvani, 2009; Sheshadri et al., 2015]. We use an alternative approach in the present study:
starting with a full general circulation model, we examine for the first time the impact of
specific orographic features (e.g., the Tibetan Plateau) on the stratosphere, by flattening
orography in localized regions within the model. These experiments suggest the influence
of orography is primarily via its influence on wave pathways rather than wave amplitudes.
In this paper we first describe the model, the three experiments, and our analysis methods.
We evaluate our control simulation, before showing the effects of individual orographic
regions on zonal mean stratospheric flow, including Eliassen-Palm fluxes and their diver-
gence. Orography-induced anomalies in Eliassen-Palm flux divergence are diagnosed by
studying wave amplitudes and pathways, before the impact of orography on SSWs is pre-
sented. We highlight similarities in EP flux pathways between orography-induced anom-
ies and those associated with SSWs in the model and observations.

2 Model, Experiments and Analysis Methods

2.1 WACCM model

To investigate the effects of orography on stratospheric circulation, we use the Whole
Atmosphere Community Climate Model (WACCM) within the CESM1.0.6 framework
[Marsh et al., 2013]. The WACCM has 66 vertical levels extending from the surface to
5.1×10^{-6} hPa, and reproduces the observed climatology and variability of stratospheric
circulation reasonably well [de la Torre et al., 2012; Marsh et al., 2013]. We use a hori-
izontal resolution of 2.5° longitude by 1.9° latitude, and include interactive chemistry with
emissions from the year 2000. Sea surface temperatures are fixed to a monthly mean cli-
matology from the merged Hadley NOAA Optimum Interpolation (NOAA/OI) dataset
[Hurrell et al., 2008]. See Marsh et al. [2013] and references therein for more details on
the WACCM.
2.2 Experiments

We perform a control simulation (CTL) with all present day orography, and three experiments with localized regions of Earth’s orography flattened. Each simulation runs for 40 years after a one year spin-up. Figure 1 shows the regions of orography we focus on: the simulation with orography flattened within the solid white box we denote ‘No Mongolia’, the short-dashed box ‘No Tibet’, and the long-dashed box ‘No Rockies’. The orography flattening is Gaussian weighted at the edges to avoid creating sharp horizontal gradients in orography. In addition to diverting the large-scale flow, orography impacts circulation through sub-grid scale processes, including gravity wave drag and an increase in surface roughness. In WACCM, these sub-grid scale processes are parameterized through two variables, SGH and SGH30; we set these to 30 and 10 m respectively in the flattened regions (approximate values for regions of low orography in Asia; the original values reached up to 1000 and 500 m respectively). See White et al. [2017] for further details of the orography flattening. The impact of each mountain region is found as the difference from the CTL experiment, e.g. ‘Impact of Mongolia’ = CTL minus No Mongolia.

We focus on December - February (DJF), when the coupling between the troposphere and stratosphere is strongest [Kidston et al., 2015] and the majority of SSWs occur [Charlton and Polvani, 2007; Butler et al., 2017].

2.3 Analysis methods

We analyze our experiments through study of the zonal mean variables: zonal wind ($\overline{u}$; overbar denotes zonal mean), Eliassen-Palm (EP) fluxes [Eliassen and Palm, 1961], and Rossby wave refractive index [e.g. Hoskins and Karoly, 1981]. We compare results from the CTL simulation to those from daily averages of 6-hourly ERA-interim re-analysis data at 0.75° resolution [Dee et al., 2011].

EP flux vectors show the propagation of zonal mean wave activity [Eliassen and Palm, 1961; Andrews and McIntyre, 1976; Edmon et al., 1980], while the EP flux divergence gives a measure of the acceleration of zonal mean flow by the waves. We calculate EP fluxes and their divergence on 60 pressure levels from 1000-0.001 hPa using daily data, and then take the climatological DJF mean. For display of the EP flux vectors we follow the scaling of Edmon et al. [1980] for log-pressure co-ordinates, omitting the factor
of $2\pi a^2/g$ (where $a$ is the Earth’s radius), and then divide by the square root of 1000/pressure [Taguchi and Hartmann, 2006].

To study the propagation pathways of wave activity we use a qualitative interpretation of the quasi-geostrophic Rossby refractive index for stationary waves, calculated as an equivalent stationary wavenumber: $K_S^2 = \frac{q_y}{u} - \frac{f_0^2}{4N^2} H^2$, where $q_y$ is the meridional gradient of potential vorticity, $f_0$ the Coriolis parameter, $N$ the stratification and $H$ the scale height [e.g., Hoskins and Karoly, 1981; Karoly and Hoskins, 1982; Andrews et al., 1987; Hu and Tung, 2002]. We calculate $K_S$ on daily data, before taking the climatological DJF mean; imaginary values of $K_S$ are treated as missing in the time averaging. Waves are refracted towards regions of greater $K_S$ [Karoly and Hoskins, 1982; Hoskins and Ambrizzi, 1993], and thus changes in $K_S$ gradients alter wave pathways. In regions where the zonal and meridional wavenumbers $k$ and $l$ are such that $k^2 + l^2 \geq K_S^2$, the vertical wavenumber is imaginary, and thus vertical propagation of waves is inhibited [Li et al., 2007; Hoskins and James, 2014]. Regions of greater $K_S$ therefore allow waves with larger $k$ and $l$ to propagate vertically.

We use the definition of SSWs of Charlton and Polvani [2007]: a major SSW occurs when $\bar{u}$ at 60°N and 10 hPa becomes easterly during November-March, after an interval of 20 or more consecutive days with westerly winds. We consider only major SSWs in this paper. Final warmings, identified as when $\bar{u}$ does not return to westerly for at least 10 consecutive days before 30 April, are excluded. SSWs can be categorized as splits, during which the polar vortex splits into two distinct pieces, and displacements, in which the vortex is displaced away from the pole [Charlton and Polvani, 2007]. We categorize each SSW using a subjective analysis of daily geopotential height fields at 10 hPa from 5 days before to 10 days after the date when $\bar{u}$ at 60°N and 10 hPa first becomes easterly: if on any day the geopotential height shows two distinct vortices of similar magnitude then the event is considered a split, otherwise it is classified as a displacement [de la Torre et al., 2012; Charlton and Polvani, 2007].

3 Results

3.1 Control Experiment (CTL)

Values of $\bar{u}$ and EP flux divergence from the CTL simulation are shown in Figure 2a, with the corresponding $K_S$ and EP flux vectors in Figure 2e. The mean circulation is
almost identical to that described by Richter et al. [2010] and Marsh et al. [2013] - zonal mean winds and temperatures in boreal winter agree relatively well with observations. There is a cold pole bias, leading to a winter vortex that initiates too early and persists for too long (see supplemental Figure S1) which may affect the stratospheric response to orography towards the beginning and end of winter. EP flux divergence (Figure 2a) and vectors (Figure 2e) agree well with those from the NCEP/NCAR re-analysis [Li et al., 2011], as does the distribution of $K_S$ in Figure 2e [Li et al., 2007]. For a detailed discussion of model biases see Richter et al. [2010] and Marsh et al. [2013].

We evaluate stratospheric variability in the CTL simulation through SSW frequency and histograms of DJF daily $\bar{u}$ at 60°N, 10 hPa. Our CTL simulation has 0.60 (±0.09) SSWs per year, consistent with previous results [de la Torre et al., 2012], and values from re-analysis data [Charlton and Polvani, 2007]; however, the ratio of splits to displacements is 1:4, in contrast to the observed ratio of approximately 1:1.2 [Charlton and Polvani, 2007]. This bias is consistent with an over-estimation of wavenumber 1 and under-estimation of wavenumber 2, as found by de la Torre et al. [2012], and shown in supplemental Figure S2. We also calculate the frequency distribution of daily $\bar{u}$ values at 60°N, 10 hPa to further evaluate variability: the CTL simulation reproduces the ERA-interim distribution relatively well (see supplemental Figure S3).

3.2 Flattened Orography Experiments

Flattening orography results in substantial differences in stratospheric flow from October - May, with the greatest differences in November - December (see supplemental Figure S1). The impact of orography on DJF zonal mean circulation is shown in Figures 2b-d and f-h. Contrary to expectations based on mountain heights, but consistent with tropospheric responses [White et al., 2017], the Mongolian mountains have the largest impact on stratospheric flow. The effect of the Mongolian (Tibetan) mountains is to decrease $\bar{u}$ by up to 21 (11) m s$^{-1}$, while the Rockies produce no significant zonal wind changes (Figures 2b-d). The stratospheric jet zonal wind changes are in thermal wind balance to within 15% (not shown).

The mountain-induced changes in EP flux divergence (Figures 2b-d) are generally consistent with the changes in $\bar{u}$: the Mongolian and Tibetan mountains both produce anomalous convergence (i.e. negative divergence) near the region of largest zonal wind de-
crease. Analysis of the individual terms in the EP divergence equation shows that changes
in horizontal divergence dominate the response to Mongolia, while both vertical and hori-
zontal divergence changes are important for the Tibetan response (not shown). In contrast
to the Asian orography, the Rocky mountains slightly increase EP flux divergence in the
stratospheric jet region.

Orography is a significant source of gravity waves, which can transport momentum into the stratosphere; however, such waves are not resolved in this climate model. Orography-induced changes in the parameterized impact of gravity waves on zonal winds in the stratosphere (obtained as a direct output of the WACCM model) are generally in the opposite direction to the zonal wind forcing from the resolved EP fluxes, consistent with the ‘compensation mechanism’ discussed by Cohen and Gerber [2013]. Orography-induced changes in gravity wave forcing are centered above 10 hPa poleward of 50°N (see supplemental Figure S4) and thus unresolved waves are not a dominant forcing of the changes in the stratospheric jet we study here, although they have a strong influence higher up in the atmosphere.

3.3 Impacts on Wave Activity

Orography-induced anomalies in EP flux divergence can be caused either by a change in wave activity amplitude, or by changes in wave propagation pathways. If changes in wave amplitude were the dominant cause of the orography-induced anomalies in EP flux divergence, then the pattern of anomalous EP fluxes would be nearly identical to that in the control just with reduced amplitude. Figures 2f and g show that, near the band of latitudes spanned by the mountains (~30-55°N), the presence of the Mongolian or Tibetan mountains produces a strong increase in wave activity propagating upwards from the surface into the upper troposphere; however the orography-induced EP flux anomalies have generally different propagation pathways from the troposphere to the upper stratosphere than the climatology (c.f. Figures 2f and g with Figure 2d). Mountains thus also induce changes in wave propagation paths. Additionally we find that orography-induced anomalies in wave amplitudes for Z at 10 hPa (for wavenumbers k = 1, 2) are inconsistent with the changes in EP flux divergence across the three experiments (see supplemental Figure S2), further suggesting that the orography-induced changes in EP flux divergence cannot be attributed solely to the orography acting as an additional source of Rossby waves.
Since changes in wave amplitude alone cannot explain the changes in EP flux divergence, we examine orography-induced anomalies in $K_s$ (Figures 2f-h), which can alter wave propagation pathways and thus EP flux divergence. Between $\sim$55-80°N and $\sim$200-10 hPa the Mongolian mountains induce a change in $K_s$ with a positive poleward gradient. This would act to deflect wave activity poleward relative to the control case and thus cause an increase in EP flux convergence poleward of $\sim$50°N, consistent with the orography-induced anomalies in EP flux vectors and convergence in this region (Figure 2b). The mountains also generally increase $K_s$ in the stratosphere, which will allow more wave activity to propagate vertically into the stratosphere, where it can converge. Orography-induced changes in $K_s$ are primarily due to changes in the meridional gradient of potential vorticity ($\overline{q}_y$) and the zonal wind ($\overline{u}$), and not changes in buoyancy ($f_0^2/4N^2H^2$) (not shown).

The change in $K_s$ due to the Tibetan mountains has a similar spatial structure to the changes due to the Mongolian mountains, but are of smaller magnitude (Figure 2g), consistent with the smaller changes in EP flux divergence and $\overline{u}$. Compared to Mongolia or Tibet, the Rockies have a much smaller impact on $K_s$ (Figure 2h), consistent with the small impact of the Rockies on the stratospheric zonal wind.

These results indicate that changes in refractive index are of central importance for the impact of orography on the wintertime stratospheric circulation. The changes in $K_s$ stem from changes in the circulation of the upper troposphere associated with the various orographic features. As shown by White et al. [2017], the Mongolian mountains have a greater impact on the upper tropospheric wintertime circulation than the Tibetan plateau.

3.4 Impacts on Major Sudden Stratospheric Warmings and Stratospheric Variability

By changing the climatological mean flow, and propagation of wave activity, the presence of mountains affects the frequency of major SSWs. The first column of Table 1 shows the SSW frequency in the different experiments, using a standard definition of SSWs (see section 2.3). Without the Mongolian mountains, the frequency of SSWs drops from 0.6 SSWs per year to 0.08. Removing the Tibetan mountains also reduces the frequency of SSWs, albeit weakly compared to the Mongolian mountains, while removing the Rockies has no statistically significant impact. The presence of mountains causes no
significant change in the date of the seasonal vortex breakdown at the end of polar winter, as defined by Black and McDaniel [2007] (at 10 mb); the delay in the switch from westerlies to easterlies when the orography is not present, as seen in supplemental Figure 3, is not statistically significant with 40 years of data. Reductions in both displacement and split SSWs occur when the Mongolian or Tibetan mountains are removed; however, given the strong bias towards displacement events in this version of the model (and subsequently the small number of split events), the effect of mountains on the ratio of splits to displacements cannot be robustly determined.

This decrease in SSW frequency could be caused by a decrease in the variability of $\bar{u}$ when mountains are removed, or simply to the increased climatological mean flow (with no change in variability) as $\bar{u}$ must decrease by a larger value to reach the 0 m s$^{-1}$ threshold. To study the relative importance of these two aspects, we create a new SSW definition to remove the influence of the change in climatological flow: instead of a fixed $\bar{u}$ threshold of 0 m s$^{-1}$, we define an SSW as a specified deviation in $\bar{u}$ ($\Delta \bar{u}$) from the DJF climatological value for each simulation. We take $\Delta \bar{u}$ to be the difference between the DJF climatological mean value in the CTL simulation (28.1 m s$^{-1}$) and 0 m s$^{-1}$. By this new definition, the frequency of SSWs only drops from 0.6 per year in the CTL simulation to 0.3 per year when the Mongolia mountains are removed, while the standard definition renders 0.08 per year without the Mongolian mountains (compare columns in table 1). Thus, approximately half of the decrease in SSW frequency when the mountains are removed is due to a simple increase in the mean $\bar{u}$, with the other half from a decrease in the variability of $\bar{u}$.

Lastly, we examine how the anomalies associated with the presence of the Mongolian mountains compare with anomalies associated with SSWs in the CTL simulation and ERA-interim re-analysis data. We show the difference in $K_S$ and EP fluxes in winters with at least one SSW relative to winters with no SSWs in WACCM and observations (Figures 3a and c). We also calculate a composite of $K_S$ and EP fluxes during all days in SSW maturation phases, defined as the 10 days prior to the minimum in $\bar{u}$ at 60°N, 10 hPa (Figures 3b and d); this is the period during which $\bar{u}$ is decreasing most rapidly. Three conclusions are drawn from Figure 3. First, the anomalies in $K_S$ and EP fluxes associated with SSWs in WACCM are relatively similar to those in observations (c.f. Figures 3a with c, and 3b with d). Second, as expected, the transient anomalies in $K_S$ and EP fluxes due to SSW events have a very similar spatial structure to the seasonal average values associ-
ated with winters in which SSW occur relative to winters with no SSWs in both WACCM and observations (c.f. Figure 3a with b, and 3c with d; note changes in scales). Finally, there is a high degree of similarity between the response to Mongolia and these SSW-associated anomalies (c.f. Figure 3 with Figure 2f), with poleward EP fluxes in the upper troposphere/lower stratosphere, an increase in vertical EP flux centered around 60°N, and general increases in $K_S$ in the stratosphere with a local maximum near the pole between 200-10 hPa. Corresponding changes in $\overline{u}$ and EP flux divergence are shown in supplemental Figure S5. Our results are consistent with the conclusion that SSWs are associated anomalous poleward propagation of wave activity, and with $K_S$ conditions that are more conducive to this propagation pathway.

### 4 Discussion and Conclusions

This study uses the WACCM model to examine the relative importance of the three major NH orographic features in shaping Northern Hemisphere (NH) stratospheric circulation. The Mongolian mountains, located north of the higher altitude and more expansive Tibetan plateau and Himalaya, have the greatest impact on NH wintertime stratospheric flow, reducing the strength of the stratospheric jet by a third. The impact of the Tibetan plateau and Himalaya is similar, although with half the magnitude. The Rocky mountains have no significant impact on stratospheric flow. The frequency of SSWs decreases when the Mongolian or Tibetan mountains are removed, partially because the threshold of 0 m s$^{-1}$ is more difficult to reach from a faster mean flow, and partially because the variability of $\overline{u}$ decreases without the mountains.

The orography-induced changes in zonal mean zonal wind are consistent with anomalous EP flux divergence, indicating that resolved wave-mean-flow interactions are key for the stratospheric response to orography. The presence of mountains has a strong impact on wave propagation paths, which affects the EP flux divergence. Orography-induced changes in $K_S$, the Rossby refractive index, exhibit meridional gradients that should increase poleward propagation of waves, increasing convergence. We conclude that orographically induced changes in background (upper tropospheric) flow, not changes in the amplitude of stratospheric transient or stationary wave activity [e.g., Plumb, 1981], are the dominant cause of mountain-induced changes in EP flux divergence.
The decrease in $\bar{u}$ from the increased EP flux convergence will generally act to reinforce the local changes in refractive index. This highlights a positive feedback mechanism in troposphere-stratosphere coupling: the $K_S$ anomalies associated with initial decreases in $\bar{u}$ alter wave propagation pathways, leading to greater EP flux convergence in the stratosphere, further slowing winds. Mongolia-induced changes in EP fluxes and refractive index are found to be similar in pattern to anomalies associated with SSWs in the WACCM model and ERA-interim re-analysis. EP fluxes associated with SSWs also show anomalous poleward propagation in the upper troposphere from 22 days prior to an SSW, with anomalous vertical fluxes centered around 70°N [Limpasuvan et al., 2004], similar to the response we find to the Mongolian mountains. Our results therefore suggest a role for positive $K_S$ feedbacks in SSW events.

Different mountain ranges have substantially different impacts on the stratospheric circulation, with orographic height clearly not the most important factor. White et al. [2017] show that the Mongolian mountains have a stronger impact on the DJF tropospheric jet than the Tibetan plateau, due to the latitudinal distribution of impinging wind, and potential vorticity gradient, both affecting the near-field response strength; and the horizontal distribution of $K_S$ affecting propagation pathways. The peak of the Rocky mountains lies at a latitude between the Mongolian and Tibetan mountains (Figure 1); however the Rocky mountains have no significant impact on stratospheric $\bar{u}$. Whether this is due to the shape of the orography, or from zonal variations in flow or wave propagation, is yet to be determined. It is worth noting that Mongolia has a particularly large impact on the upper tropospheric flow in the North Pacific [White et al., 2017], a region thought to be particularly important for SSWs [e.g., Garfinkel et al., 2012]. The region over Eastern Asia and the North Pacific has been previously highlighted regarding longitudinal asymmetries in the stratosphere [Kozubek et al., 2015; Šácha et al., 2015].

There is a large hemispheric asymmetry in stratospheric flow, with a slower and more variable jet in the Northern Hemisphere, and SSWs an extremely rare occurrence in the Southern Hemisphere (SH) [Thompson et al., 2005]. These differences are typically attributed to differences in orography and/or zonal gradients in diabatic heating (i.e. land-sea distribution). Our results give further insight into the relative importance of these asymmetries: comparison of the mean speed and variability of the SH stratosphere wintertime (JJA) jet with those of the NH wintertime jet in a simulation without Mongolia, Tibet OR the Rockies suggests that the mean speed of the jet is impacted predominantly
by the orography, while the variability about this mean is affected both by the orography
and the zonal gradients in diabatic heating (see supplemental Figure S6).

Along with the changes in stratospheric circulation, these model experiments also
show that springtime ozone over the NH pole (70-90°N) increases by up to 80 DU (20%)
with the presence of the Mongolian mountains (see supplemental Figure S7). This mountain-
effect is approximately half the magnitude of the SH ozone reduction since the 1950s
[Solomon, 1999]. This mountain-induced increase in ozone is likely a combination of
chemistry and dynamics: the mountain-induced increase in stratospheric polar tempera-
tures results in reduced ozone loss from ozone-depleting substances such as CFCs [Solomon,
1999], while an increase in the Brewer-Dobson circulation suggests an increase in troposphere-
stratosphere transport of ozone (not shown). Further work investigating the role of moun-
tains on stratospheric ozone may be of interest to paleoclimate studies.

Our results show that the boreal winter stratospheric circulation is significantly shaped
by the presence of the Mongolian mountains, due to their impact on the upper tropo-
spheric mean flow and subsequent impacts on wave propagation pathways. Thus the strato-
spheric circulation, and its variability, may be sensitive to changes in the flow impinging
on Mongolia under past climates with differing orography, such as Last Glacial Maximum
conditions, or in a future, warmer, climate.
Table 1.

<table>
<thead>
<tr>
<th>Experiment</th>
<th>SSW frequency</th>
<th>SSW_{Δu} frequency</th>
</tr>
</thead>
<tbody>
<tr>
<td>CTL</td>
<td>0.60 (0.09)</td>
<td>0.60 (0.09)</td>
</tr>
<tr>
<td>No Mongolia</td>
<td>0.08 (0.04)</td>
<td>0.30 (0.08)</td>
</tr>
<tr>
<td>No Tibet</td>
<td>0.28 (0.07)</td>
<td>0.40 (0.09)</td>
</tr>
<tr>
<td>No Rockies</td>
<td>0.70 (0.10)</td>
<td>0.63 (0.09)</td>
</tr>
</tbody>
</table>

Summary of SSW frequency in the CTL simulation and the three ‘no mountain’ experiments. First column: frequency of SSWs based on the standard definition. Second column: SSW_{Δu} frequency: SSWs defined as a decrease of Δu = 28.1 m s^{-1} from the DJF climatological mean flow for each simulation; see text for further details. Values in parentheses give one standard error based on Charlton and Polvani [2007].

Figure 1. Map of Northern Hemisphere orography, with flattened regions outlined with white boxes. 'Tibet' in short dashed line; 'Mongolia' in solid line, and 'Rockies' in long dashed line.
Figure 2.  
a. DJF zonal mean zonal wind (black contours; dashed indicating negative) and EP flux divergence (colored shading) for the CTL simulation; 
e. EP flux vectors (arrows, m$^2$ s$^{-2}$) and $K_S$ for the CTL simulation. 
Subsequent rows are as in the top row but for changes due to the Mongolian mountains (second row), Tibetan mountains (third row) and the Rocky mountains (bottom row). The EP flux is divided by the square root of 1000/pressure to aid visualization; the scale arrow is for pressure = 1000 hPa. Due to scaling of arrows, EP flux vectors should not be used to estimate divergence.
Figure 3. $K_S$ and EP flux vectors (m$^2$ s$^{-2}$) associated with SSWs in CTL (left column) and ERA-interim (right column). a, c: SSW-winter anomalies - composite of all DJF daily values for seasons with at least one SSW minus that for seasons with no SSWs. b, d: composite anomalies during the growth phase of SSWs (relative to DJF composite for seasons with no SSWs). All values calculated on daily data before time averaging. Note changes in scales between top and bottom rows.
Acknowledgments

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Supporting Information for ”Orographic impacts on boreal winter stratospheric circulation: Mongolian mountains matter most”

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1. Figures S1 to S7

Introduction Additional Figures to provide more information on the results of the WACCM simulations with mountain regions flattened.

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Figure S1. Seasonal cycle of daily 10 mb, 60°N, $\bar{\pi}$ in ERA-interim (black), the CTL simulation (blue), the No Mongolia simulation (red long dash), the No Tibet simulation (orange dot-dash), and the No Rockies simulation (green short dash).
**Figure S2.** DJF wave amplitudes for $Z$ at 10 hPa, for wavenumber 1 (a, b) and 2 (c, d) from ERA-interim (black line) and WACCM experiments (colored lines), calculated using Fast Fourier Transforms. Panels a, c show daily transients and b, d show stationary waves. Shading shows +/- one standard error calculated from seasonal averages. Stationary components $k=1$ and 2 are calculated from the climatological seasonal mean $Z$. The amplitude of the transient components $k=1$ and 2 are found by subtracting the climatological seasonal mean $Z$ from daily data, calculating the Fourier components of these daily values, and then taking the time mean of these daily amplitudes. ERA-interim re-analysis data are at $2^\circ$ spatial resolution, and are daily means of 6-hourly values.
Figure S3. Histograms of DJF daily 10 mb, 60°N, $\pi$ in ERA-interim (black) and the CTL simulation (blue).
Figure S4. Mountain-induced changes in zonal wind forcing from the gravity wave drag parameterization for a. Impact of Mongolia, b. Impact of Tibet, c. Impact of Rockies.
Figure S5. $\pi$ (black contours; dashed indicates negative) and EP flux divergence (colored shading, m/s/day) associated with SSWs in CTL and ERA-interim. a (c) SSW-winter anomalies (composite of all DJF daily values for seasons with at least one SSW minus that for seasons with no SSWs) in CTL (ERA-interim). b (d) composite anomalies during the growth phase of SSWs (relative to DJF composite for seasons with no SSWs) in CTL (ERA-interim). All values calculated on daily data before time averaging. Note changes in scales between top and bottom rows.
**Figure S6.**  a. Histograms of wintertime daily 10 mb \( \tau \) in the CTL simulation for the: NH (60°N; DJF; blue solid line) and SH (60°S; JJA; blue dashed line), and the NH (60°N) for a simulation with no Mongolia, Tibet OR Rockies (orange solid line). The simulation without the orography is much closer to the SH CTL distribution than to the NH distribution, suggesting that the orography produces the majority of the inter-hemisphere difference in the mean wintertime stratospheric jet speed. b. As for a, but with the seasonal mean value removed to compare variability about the mean. The distribution for the simulation without Mongolia, Tibet or the Rockies lies in between the NH and SH CTL distributions, suggesting that the orography produces only about half of the inter-hemisphere difference in variability.
Figure S7. Climatological total column ozone for 70-90°N in the CTL and No Mongolia simulations. Ozone concentration is produced as a direct output of the model; we convert to DU, integrate over height, and then take a latitude-weighted mean from 70-90°N.