

1 **Insights on Lateral Gravity Wave Propagation in the**
2 **Extratropical Stratosphere from 44 Years of ERA5**
3 **Data**

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10 **Key Points:**

- 11 • Climatology of lateral fluxes from ERA5 shows substantial lateral propagation of
12 gravity waves in both hemispheres
- 13 • Climatological contribution of lateral GW fluxes towards zonal mean forcing is the
14 same order of magnitude as that from vertical fluxes
- 15 • Abrupt changes in GW forcing in the upper stratosphere around sudden strato-
16 spheric warmings can last up to 20 days following the event

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Abstract

The study presents (a) a 44-year wintertime climatology of resolved gravity wave (GW) fluxes and associated zonal forcing in the extratropical stratosphere using ERA5, and (b) their composite evolution around gradual (final warming) and abrupt (sudden warming) transitions in the wintertime circulation. The connection between transformed Eulerian mean (TEM) equations and the linear GW pseudomomentum is leveraged to provide a glimpse of the importance of GW lateral propagation toward driving the wintertime stratospheric circulation by analyzing the relative contribution of the vertical vs. meridional flux convergence. The relative contribution from lateral propagation is found to be notable, especially in the Austral winter stratosphere where lateral (vertical) momentum flux convergence provides a peak climatological forcing of up to -0.5 (-3.5) m/s/day around 60°S at 40–45 km altitude. Prominent lateral propagation in the wintertime mid-latitudes also contributes to the formation of a belt of GW activity in both hemispheres.

Plain Language Summary

Internal gravity waves (GWs) exhibit both vertical and horizontal (lateral) propagation in the atmosphere, influenced by the background shear of the flow that supports them. GW model parameterizations, however, represent them in climate models assuming strict vertical propagation. This modeling assumption can have implications for modeled large-scale stratospheric circulation and variability. This study uses ERA5 reanalysis to produce the climatological distribution of resolved GW momentum fluxes and forcing in the stratosphere, and their composite evolution around prominent patterns of extratropical stratospheric variability like sudden stratospheric warmings (SSWs) and springtime final warmings (FWs). The climatology reveals that lateral propagation leads to the formation of a belt of rich GW activity in the upper winter stratosphere, which is otherwise localized over orographic hotspots in the lower stratosphere. The resolved forcing due to lateral GW propagation is found to be roughly the same order of magnitude as resolved forcing due to vertical fluxes, underlining the importance of lateral propagation for future GW parameterizations. Strikingly different GW forcing profiles are evident before vs. after SSWs and FWs, highlighting the strong two-way connection between GWs and the stratospheric mean flow.

1 Introduction

Gravity waves (GWs) dynamically couple the different layers of the atmosphere and are among the key drivers of the meridional overturning circulation in the middle atmosphere (Fritts & Alexander, 2003; Achatz et al., 2023). They provide a zeroth-order contribution towards driving the pole-to-pole mesospheric circulation (Holton, 1982; Fritts & Alexander, 2003; Becker, 2012). In the stratosphere, they influence the quasi-biennial oscillation (QBO) of tropical winds (Giorgetta et al., 2002), and the springtime breakdown of the Antarctic polar vortex (Gupta et al., 2021). GWs can also contribute to rapid breakdowns of the wintertime polar vortex, i.e., sudden stratospheric warmings (Albers & Birner, 2014; Song et al., 2020), eventually influencing tropospheric storm tracks (Kidston et al., 2015; Domeisen & Butler, 2020).

Atmospheric GWs are generated by a myriad of sources (e.g., convection, orography, jets, and fronts) and manifest over spatial scales ranging from $\mathcal{O}(10)$ km to $\mathcal{O}(1000)$ km, and evolve over temporal scales ranging from ~ 5 minutes to over a day (Fritts & Alexander, 2003). The true impact of GWs on the stratospheric circulation, and its evolution under a changing climate, is not fully understood because of limited global observations, inadequately parameterized representation in stratosphere-resolving climate models, and computationally prohibitive costs of running GW-resolving high-resolution models (Kim et al., 2003; Alexander et al., 2010; Geller et al., 2013; Plougonven et al., 2020).

67 Current GW parameterizations assume strict vertical propagation and therefore,
 68 only approximate their vertical momentum transport, i.e., they ignore their lateral (zonal
 69 and meridional) propagation. In this approximation, the net forcing due to dissipating
 70 GWs is typically estimated using the covariances ($\overline{u'\omega'}$, $\overline{v'\omega'}$) and their absolute magni-
 71 tude $\sqrt{\overline{u'\omega'^2} + \overline{v'\omega'^2}}$ (e.g., Wei et al. (2022)). Here u , v , and ω are the zonal, meridional,
 72 and pressure velocities, and the primes denote their deviation from the background flow.
 73 The covariances are approximated from GW-resolving models and observations, and the
 74 estimates are frequently used to tune subgrid-scale GW parameterizations for coarser
 75 climate models.

76 Recent analyses (Kruse et al., 2022; Procházková et al., 2023) quantified the con-
 77 tribution from lateral fluxes (in addition to the usual vertical fluxes) using a suite of mesoscale-
 78 resolving numerical weather prediction models over the Drake Passage. The studies found
 79 notable forcing over a 10-day period from lateral flux terms. The importance of these
 80 terms is further corroborated by Sun et al. (2023) who extracted and compared horizon-
 81 tal GW fluxes using three different techniques. Yet, these lateral fluxes are universally
 82 ignored by model parameterizations of GWs. Representing lateral propagation in param-
 83 eterizations would be expected to ensure a more accurate representation of GWs in cli-
 84 mate models (Sato et al., 2009; Alexander & Grimsdell, 2013; Sato et al., 2012; Plougonven
 85 et al., 2020; Polichtchouk & Scott, 2020; Pahlavan et al., 2023; Gupta et al., 2024).

86 This study presents a multidecadal climatology of both the vertical and lateral GW
 87 fluxes and provides a glimpse into their contribution to the stratospheric circulation, us-
 88 ing ERA5. The contribution of the lateral fluxes towards forcing the zonal winds in the
 89 stratosphere is evaluated against the forcing provided solely by the vertical flux, $\overline{u'\omega'}$.
 90 This is done by (a) producing a 44-year (1979-2022) DJF and JJA climatology of the
 91 vertical and horizontal transport of GW pseudomomentum during peak winters, and (b)
 92 producing a composite evolution of these terms around sudden stratospheric warmings
 93 (SSWs) in the Northern Hemisphere and the springtime final warmings (FWs) in the South-
 94 ern Hemisphere.

95 2 Background

96 For non-dissipating gravity waves, the vertical flux of zonal pseudomomentum can
 97 be related to the Reynolds fluxes, in pressure coordinates, as (Fritts & Alexander, 2003;
 98 Gill, 1982):

$$F_{zx} = \frac{-1}{g} c_{gz} \frac{E}{\hat{\omega}} k = \frac{-1}{g} \left(1 - \frac{f^2}{\hat{\omega}^2} \right) \overline{u'\omega'} \quad (1)$$

99 where c_{gz} is the vertical group velocity, $\hat{\omega}$ is the intrinsic frequency, k is the zonal wavenum-
 100 ber, E is the kinetic + potential GW energy density, f is the Coriolis parameter, u is
 101 the zonal wind, ω is the vertical velocity in pressure coordinates, and overbar denotes
 102 averaging over single/multiple wave cycles (even a zonal mean).

103 Likewise, the meridional flux of zonal pseudomomentum relates to the Reynolds
 104 fluxes as:

$$F_{yx} = c_{gy} \frac{E}{\hat{\omega}} k = \overline{u'v'} \quad (2)$$

105 where c_{gy} is the meridional group velocity of the GW.

106 Now, the zonal mean zonal wind evolution, in Transformed Eulerian Mean (TEM)
 107 form, is expressed as (Andrews et al., 1987):

$$\bar{u}_t = \left(f - \frac{1}{R \cos \phi} (\bar{u} \cos \phi)_\phi \right) \bar{v}^* - \bar{u}_p \bar{\omega}^* + \underbrace{\frac{1}{R \cos \phi} \vec{\nabla} \cdot \vec{F} + \bar{X}}_{\text{EPFD}} \quad (3)$$

108 where ϕ is latitude, p is pressure, t is time, the overbar denotes zonal mean along con-
 109 stant pressure surfaces, subscripts denote partial derivatives, \bar{u} is the zonal mean zonal
 110 wind, \bar{v}^* and $\bar{\omega}^*$ are respectively the residual meridional and vertical velocities, \bar{X} is the
 111 zonal mean parameterized GW forcing, R is the radius of the earth, and \vec{F} is the Eliassen-
 112 Palm (EP)-flux vector:

$$\vec{F} = \left(F^{(\phi)}, F^{(p)} \right) = R \cos \phi \left(-\overline{u'v'} + \bar{u}_p \frac{\overline{v'\theta'}}{\bar{\theta}_p}, \left(f - \frac{1}{R \cos \phi} (\bar{u} \cos \phi)_\phi \right) \frac{\overline{v'\theta'}}{\bar{\theta}_p} - \overline{u'\omega'} \right) \quad (4)$$

113 where θ is the potential temperature.

114 The r.h.s. covariances in Eqn 4, when computed for large-scale (small-scale) per-
 115 turbations, represent the total meridional and vertical momentum flux due to planetary
 116 waves (gravity waves). The total vertical EP-Flux in Eqn 4 equals the total vertical flux
 117 of zonal pseudomomentum in Eqn 1. Likewise, the total meridional EP-Flux in Eqn 4
 118 equals the total meridional flux of zonal pseudomomentum in Eqn 2. Thus, the EP-Flux
 119 vector, computed for small-scale perturbations, fully estimates the net meridional and
 120 vertical GW momentum flux. The meridional component, which climate model param-
 121 eterizations ignore, quantifies the lateral propagation of momentum by GWs, and as shown
 122 later, can provide notable contributions to mean flow forcing.

123 The divergence of the wave-momentum fluxes, represented by the divergence of the
 124 EP-Flux vector, then, represents the total forcing applied by the dissipating planetary
 125 waves (gravity waves) on the background flow. The EP-Flux divergence (EPFD) can be
 126 expressed as:

$$\frac{1}{R \cos \phi} \vec{\nabla} \cdot \vec{F} = \frac{1}{R \cos \phi} \left(\frac{1}{R \cos \phi} \left(F^{(\phi)} \cos \phi \right)_\phi + F_p^{(p)} \right) \quad (5)$$

127 The total EPFD comprises contributions from four terms:

- 128 i. meridional convergence of momentum: $\frac{-1}{R \cos^2 \phi} (\overline{u'v'} \cos^2 \phi)_\phi$
- 129 ii. meridional heat convergence: $\frac{1}{R \cos^2 \phi} \left(\bar{u}_p \frac{\overline{v'\theta'}}{\bar{\theta}_p} \cos^2 \phi \right)_\phi$
- 130 iii. vertical heat convergence: $\left(\left[f - \frac{(\bar{u} \cos \phi)_\phi}{R \cos \phi} \right] \frac{\overline{v'\theta'}}{\bar{\theta}_p} \right)_p$
- 131 iv. vertical convergence of momentum: $-\overline{u'\omega'}_p$

132 This means both vertical and meridional transport of GW pseudomomentum contribute
 133 to the acceleration/deceleration of the zonal mean zonal wind. In the following sections,
 134 we refer to these four forcing terms as the $\overline{u'v'}_\phi$, $\overline{v'\theta'}_\phi$, $\overline{v'\theta'}_p$, and the $\overline{u'\omega'}_p$ terms respec-
 135 tively.

136 3 Computing the Resolved GW Forcing in ERA5

137 The GW fluxes and forcing were computed using the hourly reanalysis, ERA5 (Hersbach
 138 et al., 2020), from the European Centre for Medium-Range Weather Forecasting (ECMWF)
 139 on pressure levels over 1979-2022. The data is publicly available at a $0.25^\circ \times 0.25^\circ$ hor-
 140 izontal resolution and 37 vertical (pressure) levels from 1 hPa to 1000 hPa. The small-
 141 scale perturbations of the fields were computed by removing the first 21 total wavenum-
 142 bers from the full fields (u , v , ω , and T), and then tapering the wavenumbers 21 to 42
 143 (scales 500-1000 km in the midlatitudes) using a Gaussian tapering in spectral space with
 144 a half-width of ~ 5.5 . This means the spectral coefficients were almost completely damped
 145 for wavenumber 22, damped by a factor of ~ 2 for wavenumber 35, almost fully retained
 146 for wavenumber 40, and fully retained for wavenumbers 42 and above. The gradual taper-
 147 ing leads to a smoother separation between the large- and small-scales. The filtered
 148 variables were then multiplied to compute the covariances.

149 Accounting for grid-scale hyperdiffusion and other numerical effects, ERA5 still re-
 150 solves GWs with wavelengths 200 km and longer. Stratospheric and mesospheric sponges

151 are applied at pressures less than 10 hPa and 1hPa respectively, to numerically “absorb”
 152 vertically propagating GWs.

153 3.1 Defining Sudden Stratospheric Warmings (SSWs)

154 An SSW is broadly defined as an extreme, abrupt deceleration of the wintertime
 155 stratospheric polar vortex within a short period of 5-7 days. Major SSWs are SSW events
 156 where the deceleration of the vortex is so strong that it leads to a total, albeit short-term,
 157 westerly-to-easterly reversal of the polar night jet (Butler et al., 2017; Baldwin et al.,
 158 2021). To create composites around SSWs, we identify a major SSW as the date when
 159 the abrupt wind reversal first occurs at 60°N and 10 hPa. Over the 1979-2023 DJF pe-
 160 riod, 30 such SSW events have been identified in the Northern Hemisphere (Table S1).

161 3.2 Defining Final Warmings (FWs)

162 FWs in the Austral stratosphere are defined as the springtime westerly-to-easterly
 163 transition of the zonal mean zonal wind. In this study, we identify the FW date as the
 164 first day following Austral winters when the zonal mean zonal wind at 60°S and 10 hPa
 165 turns easterly. All FW composites are produced around this date (Table S1).

166 4 44-year Climatology of GW Forcing in the Extratropical Stratosphere

167 The climatology of zonal mean GW forcing is shown in Figure 1. In both hemispheres,
 168 the $\overline{u'\omega'_p}$ term provides the strongest contribution, providing an average resolved forc-
 169 ing of up to -2 m/s/day in the Northern Hemisphere (DJF) and up to -4 m/s/day in
 170 the Southern Hemisphere (JJA) (Figure 1a). Most GW dissipation occurs above 10 hPa,
 171 and spans the midlatitudes in both hemispheres. Downward protrusions in the JJA forc-
 172 ing pattern, between 3-10 hPa, at 45°S and 75°S respectively show contributions from
 173 GWs excited over the Andes and Antarctic peninsula. The direction of flux propagation,
 174 shown by the small-scale EP-Flux vectors (Figure 1a), shows upward and poleward prop-
 175 agation of GWs, and focusing of momentum towards the polar night jet.

176 The $\overline{u'v'_\phi}$ term provides the second strongest zonal mean forcing (Figure 1b) in the
 177 upper stratosphere. The forcing is strongest in the Southern Hemisphere midlatitudes,
 178 with a net zonal acceleration between 40-50°S, and a net zonal deceleration of up to $-$
 179 0.5 m/s/day poleward of 50°S. Between 10-30 hPa, the JJA forcing from the $\overline{u'\omega'_p}$ and
 180 $\overline{u'v'_\phi}$ terms are, in fact, equally strong. The Northern Hemispheric forcing is weaker, on
 181 average, due to a weaker vortex perturbed with frequent warming events. For strong vor-
 182 tex days, the DJF deceleration is at least double the climatological average, and there-
 183 fore, similar in strength to the JJA forcing (Figure S1). The vertical momentum con-
 184 vergence provides a bulk of the forcing, but the notable contribution from lateral flux
 185 convergence highlights the prominence of lateral propagation of GWs in the upper strato-
 186 sphere.

187 The vertical heat flux convergence provides strong forcing between 30°-50° latitudes,
 188 but an order of magnitude weaker forcing in the upper stratosphere (Figure 1c vs. 1a).
 189 This indicates strong contributions from resolved inertio-gravity waves likely due to geostrophic
 190 adjustment around the midlatitude jet core (Plougonven & Zhang, 2014).

191 The horizontal maps of the DJF and JJA mean $u'\omega'$ and $u'v'$ are illustrated in Fig-
 192 ure 2. In the lower stratosphere, $u'\omega'$ is mostly localized near orographic hotspots includ-
 193 ing the Rocky Mountains, Himalayas, Scandinavian Mountains, and European Alps (Fig-
 194 ure 2a,b; green). In the middle stratosphere, the flux increasingly spreads horizontally
 195 beyond the mountain ranges (Figure 2a,b; blue) to the extent that in the upper strato-
 196 sphere, the fluxes form almost a global belt of GW activity spanning at least half the
 197 latitudinal circle (Figure 2a,b; color). The belt spans from $\sim 60^\circ$ W to $\sim 180^\circ$ E.

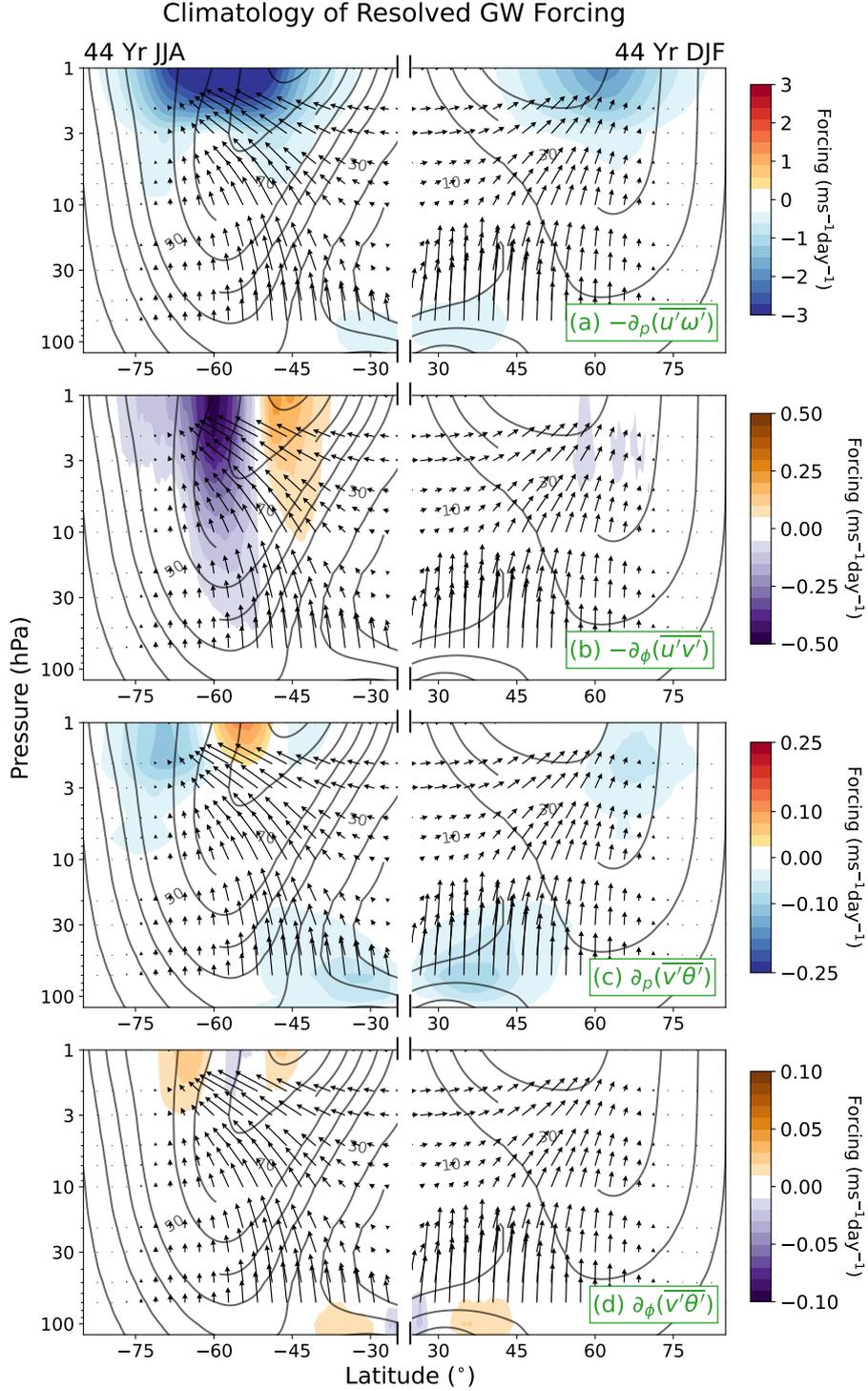


Figure 1. 44-year (1979–2022) JJA and DJF climatology of the four forcing terms (m/s/day) in Eqn 5 forming the total resolved small-scale forcing. The black curves and black arrows show the zonal mean zonal wind (m/s) and the small-scale EP-Flux respectively.

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The hotspots in both hemispheres identified in ERA5 match well with those identified from AIRS temperature data (Hindley et al., 2020). The hotspots contributing most

200 to the belt in the Northern Hemisphere include Newfoundland and Long-Range moun-
 201 tains in Canada, southeastern Greenland, the British Isles, Scandinavian mountain ridge,
 202 the Italian Alps, the Ural mountains in Eurasia, Altay-Sayan and the Greater Khingan
 203 mountains in Central and East Asia (Figure 2a). Interestingly, the strong fluxes over the
 204 Rocky Mountains, the Himalayas, and the Japanese islands do not contribute to the up-
 205 per stratospheric belt as they dissipate in the lower stratosphere.

206 Similarly, in the Southern Hemisphere, the most notable contributions to the GW
 207 activity belt in the upper stratosphere are found over the Andes, the Antarctic penin-
 208 sula, and the Southern Ocean with some contributions from New Zealand (Figure 2b;
 209 color). In the lower stratosphere, most of the GW activity is localized over these two moun-
 210 tain ranges (Figure 2b; green). As the GWs propagate vertically (and laterally), the GW
 211 activity in the middle stratosphere steadily spreads wider to regions downstream of the
 212 Andes, including most parts of the Southern Ocean (Figure 2b; blue curve).

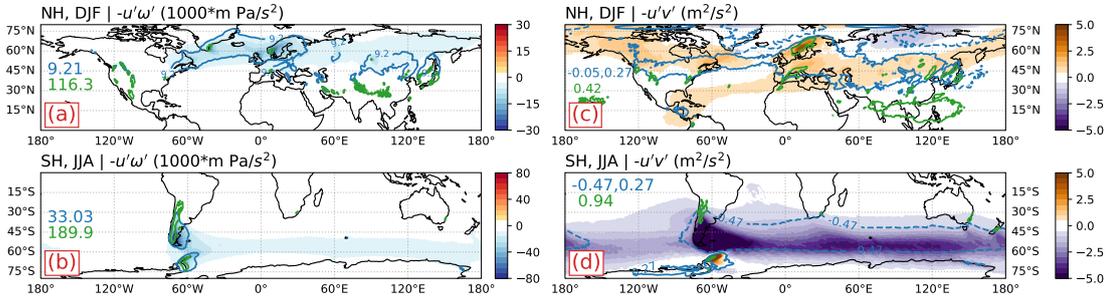


Figure 2. The map of the 44-year averaged (a)-(b) vertical flux ($u'\omega'$) and (c)-(d) lateral flux ($u'v'$), at 2 hPa altitude. Superimposed green and blue curves show the 10th-percentile envelope of the respective flux in the lower stratosphere (100 hPa) and the middle stratosphere (20 hPa) respectively. The values for the solid (positive) and dashed (negative) blue and green curves are specified in the respective figures, with units as specified in the subplot titles.

213 Lateral propagation is evident in both hemispheres, more so around prominent moun-
 214 tain ranges. The horizontal flux, $\overline{u'v'}$, in the Northern Hemisphere maximizes over the
 215 Canadian Rockies, Appalachian Mountains, the Scandinavian mountains, and the Eu-
 216 ropean Alps, and indicates a predominantly poleward transport of zonal momentum (Fig-
 217 ure 2c). Strong meridional convergence over these spots contributes the most to the zonal
 218 mean forcing provided by lateral fluxes.

219 In the Southern Hemisphere (Figure 2d), strongly negative $\overline{u'v'}$ indicates strong
 220 poleward propagation of momentum by GWs. Negative (poleward) fluxes over and down-
 221 stream of the Andes and positive (equatorward) fluxes around the Antarctic peninsula
 222 indicate momentum convergence over the Drake Passage. Though the fluxes maximize
 223 around these topographies, a streak of lateral fluxes spans the whole latitudinal circle.
 224 Using the 7-km GEOS Nature run, Holt et al. (2017) identified midlatitude-to-subtropical
 225 convection near 100 hPa as a primary source of GWs in the Southern Hemisphere. These
 226 strong but interspersed sources could form key contributions to the upper stratospheric
 227 streak.

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5 Composite Evolution of Vertical and Lateral GW Fluxes

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5.1 Evolution around SSWs

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To analyze the GW forcing evolution during abrupt dynamical changes in the stratosphere, we assess the composite evolution of the resolved forcing around 30 major SSWs over 1979-2023 (Figure 3). On average, the vortex decelerates by 35-40 m/s over 20 days leading to the SSWs, the deceleration being much stronger for the 7 days prior to wind reversal (Figure 3a).

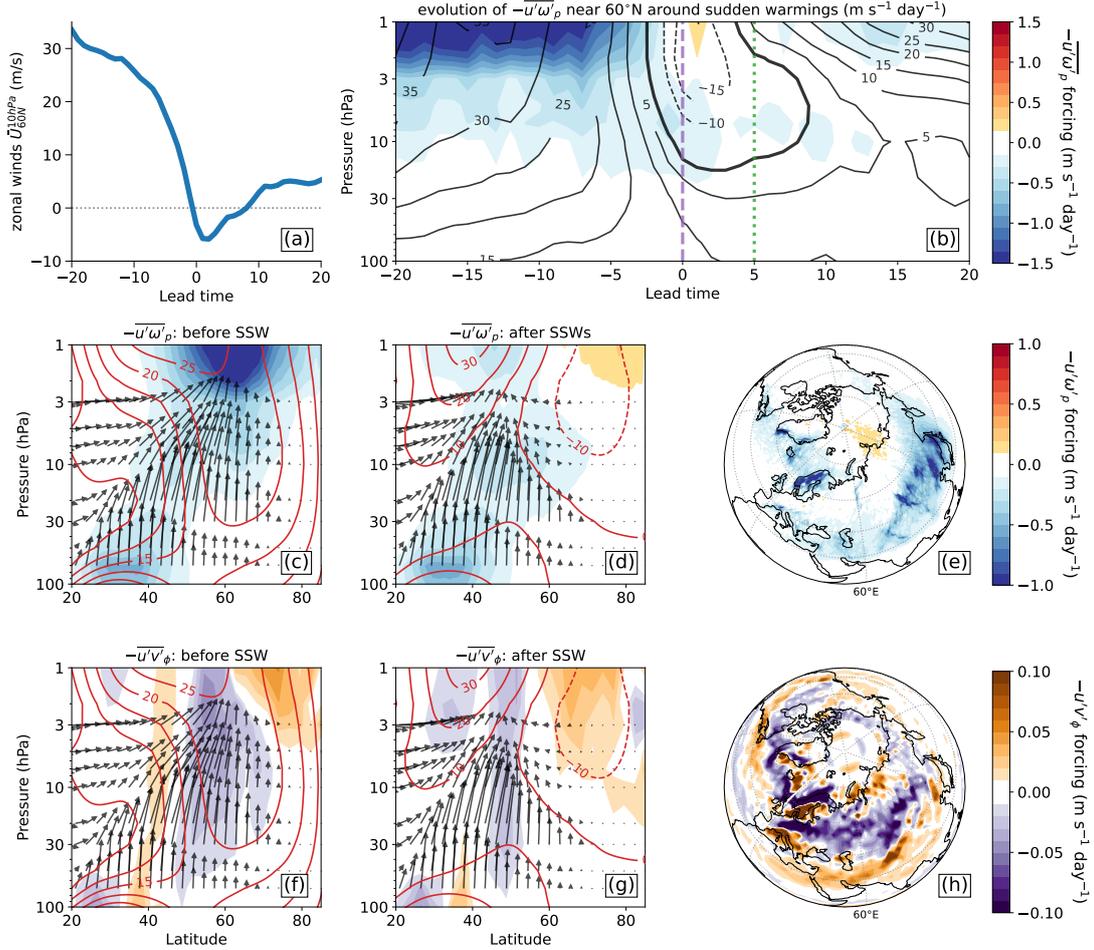


Figure 3. (a) Composite evolution of zonal mean zonal wind (m/s) at $60^{\circ}N$ and 10 hPa around 30 major SSWs over 1979-2023 and (b) composite evolution of resolved GW forcing (m/s/day) due to vertical flux convergence, i.e. the $\overline{u'\omega'_p}$ term, at $60^{\circ}N$. (c) The latitude-pressure profile of the $\overline{u'\omega'_p}$ term before SSWs averaged over lead times -20 to 0 (to the left of violet bar in (b)), (d) the $\overline{u'\omega'_p}$ term shortly after SSWs averaged over lead times 0 to 5 (enclosed by violet and green bars in (b)). (e) The map of $-\overline{u'\omega'_p}$ in the upper stratosphere (10 hPa) before SSWs, i.e., averaged over lead times -20 to 0. (f-h) Same as figures (c-e) respectively, but for the $\overline{u'v'_\phi}$ term. Black curves in (b) show the zonal mean zonal wind at $60^{\circ}N$. Red curves and black arrows in (c)-(d) and (f)-(h) respectively show the zonal mean zonal wind and small-scale EP-Flux.

235 A gradual reduction in $-\overline{u'\omega'_p}$ in the upper stratosphere is noticed 7-20 days be-
 236 fore the event, followed by a dramatic reduction in the upper atmosphere GW forcing
 237 within 7 days of the warming (Figure 3b). Upon wind reversal, the deceleration happens
 238 at slightly lower altitudes (10-20 hPa). The reversal is also accompanied by a reversal
 239 in the GW forcing in the upper stratosphere, i.e., GW dissipation provides a net accel-
 240 eration to the zonal mean flow, likely due to predominantly westward propagating GWs
 241 experiencing critical levels lower within the stratosphere.

242 The zonal wind below 10 hPa remains weak even 2-3 weeks after the SSW. Despite
 243 the zonal wind recovering to its original strength above 10 hPa, the GW forcing in the
 244 upper stratosphere remains weak relative to pre-warming strength. In the lower strato-
 245 sphere, the GW forcing remains largely unaffected before and after the SSWs at all lat-
 246 itudes, contrasting the dramatic decrease in forcing in the upper stratosphere (Figure
 247 3c vs. 3d). The decrease is accompanied by (a) an equatorward shift in the westward
 248 drag dissipation with wave focusing towards the new jet maximum at 30-35°N, and (b)
 249 GWs providing a net acceleration poleward of 60°N (Figure 3c vs. 3d).

250 The composite map of $-\overline{u'\omega'_p}$ before SSWs exhibits a wave-1 structure likely asso-
 251 ciated with wind anomalies around SSWs (Figure 3e); computing anomalies from the DJF
 252 climatologies reveal strengthening of the westward GW dissipation over the Central and
 253 East Asian mountains (Supplementary Figure S2). This seems consistent with the find-
 254 ings from the topography-removal experiments of White et al. (2018) that found these
 255 mountain ranges to strongly influence the Northern Hemisphere SSW frequency.

256 Changes in the $\overline{u'\omega'_p}$ term are accompanied by changes in the $\overline{u'v'_\phi}$ term. Before
 257 SSWs, lateral flux dissipation provides net deceleration in the jet-center region (Figure
 258 3f; purple). Following SSWs, the equatorward shift in the jet leads to an equatorward
 259 shift in the lateral flux dissipation. Moreover, the dissipation provides a net acceleration
 260 in the region with polar easterly winds (Figure 3g). The map of $u'v'_\phi$ (Figure 3h) shows
 261 that most of the contribution to the midlatitude convergence (deceleration) noted in the
 262 zonal mean (in Figure 3f) occurs over Northern Atlantic, mainland Europe, and North-
 263 ern Asia. Likewise, the divergence (acceleration) between 35-45°N occurs mostly over
 264 continental Asia, Middle East, and Southern Europe.

265 5.2 Evolution around Antarctic Final Warmings

266 We extend the analysis of Gupta et al. (2021) to assess lateral flux evolution around
 267 FWs.

268 Approaching the FW, strong westerlies in the extratropical winter stratosphere grad-
 269 ually weaken with an average deceleration of -1.2 m/s/day (Figure 4a). Composite evo-
 270 lution of the $\overline{u'\omega'_p}$ term around 60°S during this period shows a forcing of up to -3.5 m/s/day
 271 in the upper stratosphere. The GW deceleration in the upper stratosphere rapidly weak-
 272 ens 30-35 days before the FWs (Figure 4b color). The weakening is accompanied by a
 273 steady downward migration of the zero wind line and GW dissipation to lower altitudes.
 274 During this period, GWs from over a broad range of latitudes propagate upward and pole-
 275 ward and, on average, provide a peak resolved forcing of -1 m/s/day centered around
 276 60-65°S (Figure 4c). Following the FW, the reversal in the mean winds leads to the fil-
 277 tering of all stationary and westward GWs in the lower-to-middle stratosphere. The east-
 278 ward GWs propagating into the upper stratosphere and mesosphere provide a weak ac-
 279 celeration of the easterly winds (Figure 4d, red). A majority of the contribution to the
 280 zonal forcing by the $\overline{u'\omega'_p}$ term is due to waves excited over the Andes and the penin-
 281 sula (Figure 4f). The fluxes from these waves, along with non-orographic waves from storm
 282 tracks (Hendricks et al., 2014; Holt et al., 2017) converge over the Southern Ocean around
 283 60°S, providing a spiral belt of GW forcing. Near the Andes, the belt is centered around
 284 55°S but around the Ross Sea (120°E) the belt center shifts to $\sim 65^\circ\text{S}$.

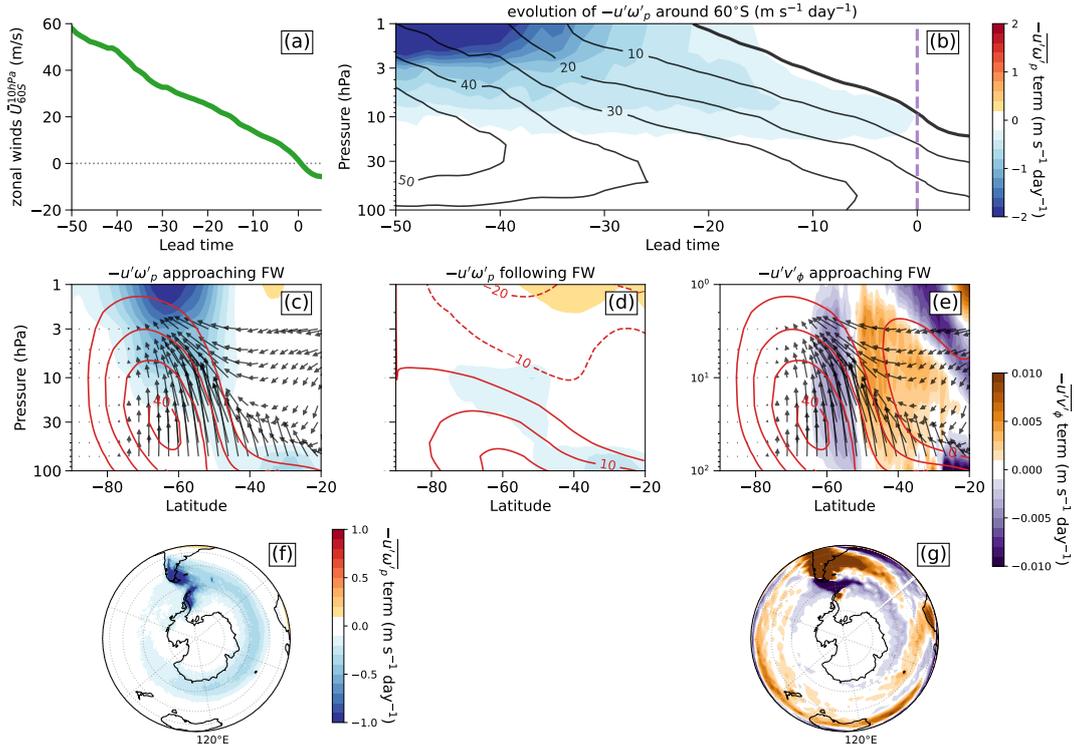


Figure 4. (a) Composite evolution of zonal mean zonal wind (m/s) at 60°S and 10 hPa around 44 FWs over 1979-2023, and (b) composite evolution of resolved GW forcing (m/s/day) from the $\overline{u'w'_p}$ term, at 60°S. (c) The latitude-pressure profile of the resolved forcing before FWs averaged over lead times -50 to 0 (left of the violet bar in (b)). (d) resolved GW forcing shortly after FWs averaged over lead times 0 to 5 (right of the violet bar in (b)). (e) The latitude-pressure profile of the forcing from lateral flux convergence, i.e. the $\overline{-u'v'_\phi}$ term averaged over lead times -50 to 0. (f)-(g) Map of the GW forcing from the $\overline{-u'w'_p}$ and $\overline{-u'v'_\phi}$ terms respectively averaged over lead times -50 to 0. Black curves in (b) show the zonal mean zonal wind at 60°S. Red curves and black arrows in (c)-(e) respectively show the zonal mean zonal wind and small-scale EP-Flux. (c), (d) and (f) share the same colorscale, so do (e) and (g).

285 Contrary to the JJA mean, the zonal mean forcing from the $\overline{-u'v'_\phi}$ term around FWs
 286 is one-to-two orders of magnitude weaker than that from the $\overline{u'w'_p}$ term (Figure 4e vs
 287 4c). This is because strong deceleration from this term is localized around the Andes and
 288 in the zonal mean is balanced by the acceleration provided by lateral fluxes over other
 289 sources around Southern Africa and Oceania (Figure 4g). Nevertheless, strong local de-
 290 celeration from this term can be important for an accurate representation of mesoscale
 291 variability around the Drake Passage and over the Southern Ocean.

292 **6 Conclusions and Discussion**

293 We produce a 44-year DJF and JJA climatology of resolved zonal GW forcing in
 294 the extratropical stratosphere using ERA5 and assess its composite evolution around Bo-
 295 ream SSWs and Austral FWs. We analyze both the vertical and the meridional flux of
 296 GW pseudomomentum to quantify the impact of lateral propagation towards the zonal
 297 flow forcing. Model parameterizations of GWs typically ignore lateral effects and only
 298 focus on vertical propagation when approximating subgrid-scale fluxes. Relative forc-

ing contribution from these terms demonstrates that lateral propagation effects are prominent in the midlatitudes, especially near orography, and could be important for the middle-to-upper stratospheric circulation.

The analysis complements other efforts to (i) produce GW climatology in the stratosphere using observations (Geller et al., 2013; Ern et al., 2018; Hindley et al., 2020; Wei et al., 2022, for instance), (ii) analyze GW contributions towards stratospheric circulation and variability (Polichtchouk et al., 2018; Sato & Hirano, 2019; Eichinger et al., 2020; Gupta et al., 2021; Cullens & Thuraijah, 2021; Pahlavan et al., 2021), and (iii) assess the complete GW forcing (Kruse et al., 2022; Procházková et al., 2023; Sun et al., 2023).

Following the Transformed Eulerian Mean framework, we estimate the mean GW pseudomomentum flux by estimating the vertical flux of zonal momentum ($u'\omega'$), meridional flux of zonal momentum ($u'v'$), and meridional heat flux ($v'\theta'$). The vertical convergence of $u'\omega'$ dominates the total GW forcing in both the DJF and JJA stratospheric midlatitudes, with the peak resolved JJA forcing at 40-45 km height (-4 m/s/day) being more than double in magnitude than the respective climatological DJF forcing (-1.5 m/s/day). Still, meridional convergence of lateral fluxes forms a considerable fraction of the forcing around orography and over the Southern Ocean, providing a zonal mean resolved forcing of -0.5 m/s/day at those altitudes.

The lateral effects in the Southern Hemisphere are stronger during peak winter than during springtime. Lateral propagation of GWs, together with local GW sources, leads to the formation of belts of GW activity in both hemispheres' upper stratosphere. Momentum flux hotspots appear over orography but appear to spread over a much broader region in the upper stratosphere due to lateral propagation.

The composite evolution of GW forcing around major SSWs and FWs demonstrates the sensitivity of GW forcing to changes in the stratospheric mean state, suggesting possible changes in stratospheric GW forcing in a changing climate. Abrupt changes to the polar vortex are associated with abrupt changes in the upper stratospheric GW forcing due to changes in GW propagation conditions. Even after the vortex recovers to pre-SSW strength in the upper stratosphere, persisting wind anomalies in the middle stratosphere prevent tropospheric GWs from propagating into the upper stratosphere.

The analysis only provides a glimpse into the true GW climatology, as ERA5 and even high-resolution models underestimate the resolved GW forcing in the stratosphere on account of prescribed dissipation or limited grid resolution (Holt et al., 2016; Wicker et al., 2023; Gupta et al., 2024). These unresolved GWs, with wavelength $\in (10, 100)$ km, can account for a major chunk of extratropical GW forcing (Polichtchouk et al., 2022, 2023). The stratospheric sponge in ERA5 between 1-10 hPa could also attenuate the resolved GWs. Further, computing vertical convergence on 37 pressure levels, as opposed to 137 model levels, likely underestimates the forcing. Lastly, Gaussian tapering of complex coefficients dampens the contributions from spatial scales 500-1000 km to some degree (Figure S3), resulting in weaker GW forcing profiles than those in previous studies which employ a fixed-wavenumber cutoff (Geller et al., 2013; Wicker et al., 2023; Gupta et al., 2021).

The findings affirm the importance of lateral propagation, suggesting its importance for GW parameterization development. Neglecting lateral propagation is believed to be a prime reason for "missing drag" around 60°S , causing temperature biases and delayed Antarctic vortex breakdown in climate models (Sato et al., 2012). In fact, in addition to GWs (Plougonven et al., 2020; Eichinger et al., 2023; Voelker et al., 2023), model representation of a multitude of mesoscale processes including tropical (slantwise) convection (Chen et al., 2018), planetary boundary layers (Xie et al., 2012), radiative transfer (Jakub & Mayer, 2017), and convective boundary layer (Sorbjan, 2009), could stand to benefit from a nonlocal (three-dimensional) parameterized representation.

Data Availability Statement

ECMWF's ERA5 data can be freely accessed from <https://www.ecmwf.int/en/forecasts/datasets/reanalysis-datasets/era5>

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