The Midlatitude Lower-Stratospheric Mountain Wave "Valve Layer"

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ABSTRACT

The vertical propagation and attenuation of mountain waves launched by New Zealand terrain during the Deep Propagating Gravity Wave Experiment (DEEPWAVE) field campaign are investigated. New Zealand mountain waves were frequently attenuated in a lower-stratospheric weak wind layer between \( z = 15 \) and \( z = 25 \) km. This layer is termed a "valve layer," as conditions within this layer (primarily minimum wind speed) control mountain wave momentum flux through it, analogous to a valve controlling mass flux through a pipe. This valve layer is a climatological feature in the wintertime midlatitude lower stratosphere above the subtropical jet.

Mountain wave dynamics within this valve layer are studied using realistic Weather Research and Forecasting (WRF) Model simulations that were extensively validated against research aircraft, radiosonde, and satellite observations. Locally, wave attenuation is horizontally and vertically inhomogeneous, evidenced by numerous regions with wave-induced low Richardson numbers and potential vorticity generation. WRF-simulated gravity wave drag (GWD) is peaked in the valve layer, and momentum flux transmitted through this layer is well approximated by a cubic function of minimum ambient wind speed within it, consistent with linear saturation theory. Valve-layer GWD within the well-validated WRF simulations was 3–6 times larger than that parameterized within MERRA. Previous research suggests increasing parameterized orographic GWD (performed in MERRA2) decreases the stratospheric polar vortex strength by altering planetary wave propagation and drag. The results reported here suggest carefully increasing orographic GWD is warranted, which may help to ameliorate the common cold-pole problem in chemistry–climate models.

1. Introduction

Gravity waves (GWs) play an important role in Earth's general circulation by transporting energy and momentum vertically, which are deposited wherever the waves are attenuated (McLandress 1998; Fritts and Alexander 2003). The importance of GWs in the general circulation increases with altitude. Nonattenuating vertically propagating gravity waves increase in amplitude with decreasing density, causing them to eventually induce instabilities (e.g., static and/or Kelvin–Helmholtz instabilities) and attenuate. Additionally, as density decreases, the decelerations (and accelerations) associated with the momentum deposited by attenuating GWs [i.e., gravity wave drag per unit mass (GWD)] become increasingly dominant in the momentum budget. GWD plays a primary role in driving the general circulation in the mesosphere and lower thermosphere (Holton 1983; Garcia and Solomon 1985; Holton and Alexander 2000) as well as the tropical stratosphere (Dunkerton 1997), while possibly playing a more secondary role in the extratropical stratosphere. The focus of this study is the propagation and attenuation of gravity waves within the extratropical stratosphere.

In the stratosphere, the important equator-to-pole Brewer–Dobson circulation is driven by momentum deposited in the extratropics by both planetary-scale Rossby waves and GWs (e.g., Holton et al. 1995). Within chemistry–climate models, planetary waves are resolved while the smaller-scale GWs and their GWD are largely unresolved and parameterized. These parameterizations involve free parameters that are typically tuned to produce a realistic general circulation as opposed to
being constrained around an observational dataset (Alexander et al. 2010; Geller et al. 2013). Hence, the magnitude of GW driving of the Brewer–Dobson circulation (and other circulation features) is uncertain. This is illustrated in the model intercomparison presented by Eyring et al. (2010, their Fig. 4.10a), where GWD is responsible for little to nearly half of tropopause-level tropical upwelling among model members. Surprisingly, despite variable GWD contributions, the circulation strength was found to be relatively constant, implying changes in planetary wave driving compensate variations in GW forcing (e.g., Cohen et al. 2013) and that the mean transport circulation alone may not strongly constrain GWD parameterizations.

A current common problem in chemistry–climate models is that the Southern Hemisphere pole is too cold in the austral winter stratosphere, resulting in an overly strong stratospheric polar vortex that breaks down too late in spring (Eyring et al. 2010). This cold-pole problem may be due to too little GWD, too little planetary wave drag, or some combination of both acting on the stratospheric polar vortex. McLandress et al. (2012) suggested that this issue is due to missing GWD in these models near 60°S, a latitude band with no major mountain ranges. This missing GWD may be in part due to missing mountain waves launched by mountainous islands near 60°S (e.g., Alexander and Grimsdell 2013), neglected meridional propagation or focusing of GWs into the stratospheric polar vortex jet (e.g., Sato et al. 2009), or underrepresented nonorographic GWs and drag resulting from jet and frontal imbalances near 60°S (e.g., Jewtoukoff et al. 2015) in climate simulations. It is less clear if too little planetary wave drag is part of the cold-pole problem. However, Sigmond and Scinocca (2010) found that increasing orographic GWD significantly reduces the strength of the stratospheric polar vortex primarily by altering planetary Rossby wave propagation and drag. This result suggests that increasing parameterized orographic GWD in chemistry–climate models might reduce the cold-pole problem in free-running climate simulations.

In this paper, the vertical propagation and attenuation of New Zealand mountain waves are studied using deep Weather Research and Forecasting (WRF) Model simulations with realistic topography and initial/boundary conditions. These simulations are extensively validated against research aircraft, radiosonde, and satellite observations collected over New Zealand during the Deep Propagating Gravity Wave Experiment (DEEPWAVE) field campaign during May–July 2014 (Fritts et al. 2016). From these simulations, the vertical fluxes of horizontal momentum and GWD are quantified and compared with DEEPWAVE observations and parameterized quantities within NASA’s Modern-Era Retrospective Analysis for Research and Applications (MERRA) (Rienecker et al. 2011) and MERRA2 (Bosilovich et al. 2010). Throughout the DEEPWAVE period, there was a frequent ambient wind minimum in the lower stratosphere, resulting in significant mountain wave attenuation and GWD. GW dynamics within this layer are the focus of this paper.

The outline of this paper is as follows. The WRF Model configuration and validation are presented in section 2. In section 3, mountain wave propagation and attenuation is investigated in detail over New Zealand for the 2014 austral winter season. Here, vertical fluxes of horizontal momentum and GWD are quantified and controls on vertical mountain wave propagation within the WRF simulations are investigated. Local characteristics of mountain wave attenuation are also presented. In section 4, WRF-simulated momentum fluxes and GWD are quantitatively compared with those parameterized in MERRA and MERRA2. Conclusions are provided in section 5.

2. Model configuration and validation

a. WRF configuration

Two types of WRF simulations are presented in this study: a single “long run” and five “event runs.” The long run is a 6-km-resolution simulation (outer domain in Fig. 1a) initialized once at 0000 UTC 24 May 2014 and run continuously through 0000 UTC 1 August 2014. The long run was effectively only forced by boundary conditions provided by 0.125° × 0.125° resolution operational ECMWF analyses every 6 h. The event runs had a 6-km-resolution outer domain and a 2-km-resolution inner nest centered on the New Zealand region (Fig. 1a) and were completed for five mountain wave events sampled during the DEEPWAVE field campaign. Start and end times, simulation names, and research flights included within the simulations are summarized in Table 1. All New Zealand transects on the 26 (12) research flights flown by the NSF/NCAR Gulfstream V (German DLR Falcon) research aircraft [NGV (DLRF)] during DEEPWAVE were contained within the time-space domain of the long run. Seven (four) NGV (DLRF) research flights were included within the event-run domains. The long-run solution was output every 3 h and event-run solutions were output hourly.

All simulations presented were configured identically to those presented in Kruse and Smith (2015) and so are not discussed in full detail here. The same parameterization selections [see appendix A of Kruse and Smith (2015)] were used for all simulations. No convection
A parameterization was used and a 10-km upper sponge layer was specified below the domain top at 1 hPa or ~45 km. The ACM2 turbulence parameterization (Pleim 2007) was selected for all simulations. Model terrain within the 6 (2)-km-resolution domains was derived from the 2-arc-min (30-arc-s) digital elevation model (DEM) distributed with WRF. This terrain height was bilinearly interpolated onto the WRF grid and then smoothed with a single pass of the “smoother–desmoother” filter (Grell et al. 1994) to highly damp resolution-scale (2–4 km) terrain features.

b. Simulation validation

Here, the WRF simulations are compared against observations collected on board the NGV and DLRF research aircraft, radiosondes launched from Hokitika, Haast, and Lauder (see Fig. 1b for locations), and NASA’s Aqua satellite by the Atmospheric Infrared Sounder (AIRS). Comparisons of mean quantities representing the ambient environment, mean vertical flux of zonal momentum, and temperature variance near the top of the simulation domain are presented.

To compare the WRF simulations to the aircraft and radiosonde observations, WRF variables were linearly interpolated in space and time to every observation (sampling frequency of 1 Hz) of a particular flight. The actual and model flights were then postprocessed identically, allowing quantitative comparison. Aircraft quantities were averaged along flight legs, which are defined as a straight and level transect over the Southern Alps of New Zealand (e.g., red line in Fig. 1b). Data along a single cross-mountain flight leg flown through the actual, long-run, and event-run atmospheres are shown in Fig. 2. Qualitatively, the perturbation amplitudes are somewhat underrepresented in the 6-km long run but better represented in the 2-km event run. This

<table>
<thead>
<tr>
<th>Simulation name</th>
<th>Resolution (km)</th>
<th>Start time</th>
<th>End time</th>
<th>Hours simulated</th>
<th>Research flights</th>
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<td>6</td>
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<td>0000 UTC 1 Aug</td>
<td>1653</td>
<td>RF01–26, FF01–12</td>
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<tr>
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<td>2</td>
<td>0000 UTC 17 Jun</td>
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<td>2</td>
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was the case for most legs. The agreement between the two simulations and aircraft observations in Fig. 2 was particularly good for this flight leg. Observed and simulated mean zonal wind \((U = \bar{u})\), meridional wind \((V = \pi)\), and temperature \(T\) are compared in Fig. 3, where the overline indicates a spatial average over the flight leg [i.e., \(\overline{\cdots} = \frac{1}{L} \int_{x_i}^{x_f} \cdots \ dx\), where \(L = x_f - x_i\) with \(x_f > x_i\)]. These leg-mean quantities simulated within the long run are plotted versus observations for all 135 legs flown over New Zealand by both the NGV and DLRF during DEEPWAVE in the top row of Fig. 3. The agreement is very good, with regression slopes near unity and coefficients of determination \(R^2\) better than 0.96 for all three quantities. In the bottom row of Fig. 3, the 2-km-resolution event runs are compared against the 74 NGV and DLRF legs flown within these runs, resulting in similar accuracy compared with the 6-km-long run. Mean error (ME) or bias, percentage ME, mean absolute error (MAE), and percentage MAE validation statistics are provided in Table 2. WRF exhibits little bias for the horizontal wind components \((|ME| < 1.4 \text{ m s}^{-1})\) and temperature \((|ME| \leq 0.5 \text{ K})\) and scatter \((\text{MAE} \leq 1.9 \text{ m s}^{-1}\) and \(\text{MAE} \leq 0.9 \text{ K}\), respectively) for both the long run and event runs. These validation statistics are similar for both the long run and event runs, suggesting the higher resolution and more recent initialization in the event runs do not improve or degrade model solutions with respect to mean or ambient quantities.

A similar comparison of ambient quantities was performed between the WRF simulations and 144, 50, and 95 radiosondes launched from Hokitika, Haast, and Lauder (see Fig. 1b for locations), respectively. Data from observed and simulated radiosonde flights were smoothed with a 4-km vertical moving average and then compared every kilometer. ME and MAE for \(u\) and \(T\) are shown at every altitude in Fig. 4. For \(u\) (and \(v\), not shown), ME was within \(\pm 1 \text{ m s}^{-1}\) at all altitudes except below 5 km. The MAE was relatively high in the troposphere \((\sim 2.75 \text{ m s}^{-1})\) and low in the stratosphere \((\sim 1.25 \text{ m s}^{-1})\). Temperatures within WRF were generally biased cold, with the cold bias increasing from about \(-0.2 \text{ K}\) at 5 km to \(-0.9 \text{ K}\) near 27 km. MAE increased slightly with height, but was generally at or below 1 K. This sounding comparison was repeated for the 2-km event simulations (not shown), producing similar results. This remarkable validation of mean quantities highlights the accuracy of the ECMWF data assimilation system, likely largely attributable to assimilation of satellite observations in the Southern Hemisphere (e.g., Bouttier and Kelly 2001).

Below 5 km, the zonal winds within WRF were notably stronger than observed, while the meridional winds were slightly weaker than observed (not shown). This may be due to reduced flow blocking within WRF relative to reality associated with reduced model terrain heights. This issue was modestly improved in the 2-km-resolution event runs (not shown).

Long-run and event-run leg-averaged zonal momentum fluxes \((\text{MF}_x = \bar{p} \bar{u} \bar{w}')\) are compared against those observed by the NGV in Fig. 5. Along flight legs, perturbations, denoted by primed quantities, are defined relative to the leg average following Smith et al. (2016). In Fig. 5a, leg-average \(\text{MF}_x\) in the observed, long-run, and event-run atmospheres are plotted versus research flight number. There is significant intraevent variability of leg-averaged \(\text{MF}_x\) in both the observations and simulations, which took place within the 6–9-h duration of each research flight. Long-run- and event-run-simulated \(\text{MF}_x\) are plotted versus NGV observations in Figs. 5b and 5c, respectively. There is significant scatter between both types of simulations and the observations, suggesting leg-averaged quadratic wave quantities like \(\text{MF}_x\) may not be predictable on a leg-by-leg basis despite the accurately represented ambient environment (Figs. 3 and 4). The 2-km-resolution event runs produced larger fluxes than the long run and to some extent the observations. \(\text{MF}_x\) validation statistics are summarized Table 2. Overall, long-run \(\text{MF}_x\) is about 5% weaker than observed, while the event-run \(\text{MF}_x\) is about 50% stronger than observed (note that a positive ME
implies a weaker flux since $\text{MF}_x$ is negative). However, $\text{MF}_x$ along a single flight leg might be off by $\sim 60\%$ and $\sim 85\%$ in the long-run and event-run simulations (see Percentage MEA for $\text{MF}_x$ in Table 2), respectively, again emphasizing that momentum fluxes measured along an aircraft transect may be unpredictable. This may be due to inherent unpredictable variability associated with non-linear generation mechanisms or sampling/averaging over short time and space scales within a complex 3D wave field. Further work is needed understand this unpredictable variability.

Two-dimensional linear theory suggests that horizontally averaged mountain wave momentum flux is proportional to the cross barrier wind speed and the squared terrain height (e.g., McFarlane 1987). The fact that the model topography in both the long run (max = 2133 m) and event runs (max = 2819 m) is less than the actual terrain height (max = 3724 m) may act to reduce simulated $\text{MF}_x$. However, the stronger-than-observed cross-barrier flow may act to increase $\text{MF}_x$. These two effects may compensate in the long run so that momentum fluxes are only weakly ($\sim 5\%$) underrepresented. The winds in the

![Diagram of leg-average wind and temperature distributions](image)

**FIG. 3.** Leg-average (a),(d) zonal wind, (b),(e) meridional wind, and (c),(f) temperature from (top) the 6-km-resolution long run and (bottom) the 2-km-resolution event runs plotted vs aircraft observations. Legs flown by the NGV and DLRF are shown in black and red, respectively.

| Table 2: WRF validation statistics, validating WRF against aircraft observations, for wind speed ($\text{m s}^{-1}$), temperature (K), and $\text{MF}_x$ (mPa). |
|---------------------------------|---------|---------|---------|---------|---------|---------|---------|---------|
| No. of legs | 135 | 74 | 135 | 74 | 135 | 74 | 97 | 58 |
| Regression slope | 0.957 | 0.947 | 0.981 | 0.969 | 0.988 | 0.998 | 0.667 | 1.236 |
| Regression $R^2$ | 0.956 | 0.967 | 0.974 | 0.975 | 0.973 | 0.214 | 0.372 |
| ME | −0.795 | −1.365 | 0.601 | 0.757 | −0.435 | −0.469 | 3.838 | −46.584 |
| Percentage ME | −3.81 | −5.10 | −8.86 | −12.52 | −0.20 | −0.21 | −5.56 | 50.85 |
| MAE | 1.812 | 1.855 | 1.725 | 1.623 | 0.773 | 0.816 | 40.554 | 78.735 |
| Percentage MAE | 8.69 | 6.94 | 14.28 | 13.53 | 0.35 | 0.37 | 58.76 | 85.69 |
2-km-resolution event runs are slightly weaker than the long-run winds but still 5%–10% stronger than reality. The increased terrain height may explain the stronger momentum fluxes relative to the long run, while the stronger-than-observed cross barrier winds may explain the over predicted momentum fluxes relative to the observations.

Brightness temperature variances from AIRS channels 81 and 82 are compared to temperature variances within the WRF simulations. Details on the AIRS GW extraction and averaging procedures can be found in Eckermann and Wu (2012) and examples of AIRS-imaged GW events from specific DEEPWAVE research flights are presented and discussed in Fritts et al. (2016). To quantitatively compare WRF and AIRS, the average vertical weighting function for these two AIRS channels (Fig. 6) was applied to long-run-simulated 3D temperature perturbation fields at a given time to produce 2D fields of forward-modeled AIRS brightness temperature perturbations using the following relation:

$$T_{WRF}^\prime(x, y) = \frac{\int_{z_{dc}}^{z_{top}} l(z) T_{WRF}^\prime(x, y, z) \, dz}{\int_{z_{dc}}^{z_{top}} l(z) \, dz},$$  \hspace{1cm} (1)$$

where $z_{dc}$ and $z_{top}$ are the heights of the lower- and upper-domain boundaries, $l(z)$ is the vertical weighting function in Fig. 6, and $T_{WRF}^\prime$ perturbations were computed following the methodology briefly discussed in section 3a. This 6-km-resolution field of long-run $T_{WRF}^\prime$ was then averaged to 18-km resolution, similar to the horizontal resolution of AIRS (nadir footprint diameter of ~13.5 km, 41 km $\times$ 22.4 km footprint at swath edges, along-track resolution of 18 km (Aumann and Miller 1995)]. Finally, a $3 \times 3$ moving-average smoother was applied to the WRF $T_{WRF}^\prime$ field, similar to a three-point along-track smoother applied to the AIRS observations to reduce noise (Eckermann and Wu 2012).

Figure 7 shows time series of simulated and observed brightness temperature ($T_{WRF}^\prime$ and $T_{AIRS}^\prime$, respectively) variances over the boxed region in Fig. 1b. There is some quantitative agreement between the two time series, particularly for the deep events on 24 May and 20 June. However, the WRF variances are 7 times smaller than observed owing to the effect of instrument noise on the observations. Noise floors of these channels act to keep $T_{AIRS}^\prime$ variances above 0.015 K$^2$, likely obscuring many real but weak GW events. The qualitative agreement between the series for particular events (e.g., 4–5 and 14 June and 11–12 and 21 July) provides evidence that the long run was able to reproduce deeply propagating mountain wave events. Additionally, a number of shallow mountain wave events took place between 23 June and 5 July. Both time series have relatively low variances throughout this period despite the tropospheric forcing, suggesting that both the simulated and actual mountain waves were attenuated well below the peak of the AIRS weighting function (~27 km; see Fig. 6). These results suggest the long run was able to reproduce some aspects of both deep and shallow mountain wave events.

In summary, both the 6- and 2-km-resolution long run and event runs were able to quantitatively reproduce ambient horizontal winds and temperatures over the whole DEEPWAVE field campaign between $z = 2$ and 30 km. This agreement suggests there is little interior drift within the continuously running long run. Despite accurately reproducing slowly varying ambient quantities, there is significant intraevent (<9 h) variability in MF,
both observed and simulated, suggesting MF$_x$ averaged along a single aircraft leg may be unpredictable. Higher resolution is important in reproducing the magnitude of this intraevent variability. Despite the lack of predictability, the MF$_x$ bias was small relative to the mean (~5% weaker) in the long run but was considerable in the event runs (~50% stronger). Qualitatively, significant observed deep and shallow wave events within satellite temperature observations were also deep and shallow within the long run. We proceed to study wave attenuation in the long run in the next section having established that the ambient atmosphere, wave momentum fluxes, and wave attenuation levels are well represented.

3. Mountain wave attenuation over New Zealand

a. The valve layer

Mountain waves play an important role in the general circulation of the atmosphere through the momentum they carry and deposit as they attenuate. This momentum deposition represents a forcing in the horizontal momentum equations for the ambient (e.g., horizontally averaged) flow (McLandress 1998). Here, wave attenuation is quantified as GWD per unit mass,

$$GWD = \frac{\Delta MF}{\rho \Delta z},$$

which is the acceleration a volume of air of a given density would undergo given deposited wave momentum. Here, MF = MF$_x$i + MF$_y$j = p($\mathbf{u}\cdot\mathbf{w}$i + $\mathbf{v}\cdot\mathbf{w}$j) is the vertical flux of horizontal momentum vector and the overline represents an area average $[\langle \cdot \rangle = 1/A \int \langle \cdot \rangle \, dA]$. In the absence of GW attenuation, the momentum flux vector is conserved with height (i.e., $\partial$MF/$\partial z$, GWD = 0) for steady 3D vertically propagating GWs entirely contained within the averaging region (Vosper and Mobbs 1998).

Zonal GWD is computed within WRF using the method presented by Kruse and Smith (2015), where
events gradually and relatively smoothly deposit their momentum as they propagate (Fig. 8e), while waves in shallow events are strongly attenuated within a layer between 15 and 25 km, producing a local maximum in zonal GWD in the time-averaged profile (Fig. 8f). This layer is termed a mountain wave “valve layer,” as it restricts mountain wave momentum fluxed into the upper stratosphere, analogous to a valve restricting mass flux of some fluid through a pipe. This valve layer is associated with wind speed minima that occur above tropopause-level wind maxima (Fig. 8a), which cause mountain waves to steepen, become nonlinear (Fig. 8g), and attenuate. Nonlinearity is quantified pointwise by the nonlinearity ratio, NLR(x, y) = (u^2 + v^2)/(U^2 + V^2), where U, V = U(x, y), V(x, y) are the smoothly varying background wind components such that u, v = U(x, y), V(x, y) + u’(x, y), v’(x, y) as opposed to horizontally averaged values used elsewhere in this paper. Note that some regions of NLR in Fig. 8g are associated with periods of very weak winds resulting in high nonlinearity rather than mountain wave attenuation.

This mountain wave valve layer is distinct from a critical level. A critical level is defined as the level where the ambient wind speed through a wave (i.e., wind speed perpendicular to the phase fronts U^⊥) is equal to the horizontal phase speed c_p of the gravity wave. For a north–south-oriented stationary mountain wave (U^⊥ = U, c_p = 0), a critical level is a level of zero ambient zonal wind. Mountain waves of all amplitudes and scales will be almost completely attenuated at a zonal wind reversal (i.e., critical level) in this scenario (Booker and Bretherton 1967). A valve layer, however, is defined as a layer of reduced wind speed with no wind reversal or critical level (e.g., Fig. 9). Such a layer might allow small-amplitude mountain waves to be transmitted unaffected but might also cause larger-amplitude mountain waves to steepen, become nonlinear, and attenuate (e.g., between 23 June and 5 July in Figs. 8a,e,g).

The valve-layer concept may also be applied to nonstationary gravity waves (i.e., c_p ≠ 0) by considering profiles of wind speed through the wave relative to the wave phase speed (U^⊥ - c_p). Wave attenuation within this frequently present lower-stratospheric valve layer above New Zealand is quantitatively investigated below.

b. GW attenuation and potential vorticity

Another useful attenuation diagnostic is potential vorticity (PV), defined by

\[ P = \frac{\omega \cdot \nabla \times \mathbf{u}}{\rho}, \]  

where \( \omega \) is the angular frequency of the gravity wave, \( \nabla \times \mathbf{u} \) is the vorticity of the flow, and \( \rho \) is the density of the fluid. This diagnostic helps to identify regions of strong wave attenuation and can be used to evaluate the impact of orographic and nonorographic perturbations on the propagation of mountain waves.
where $\mathbf{\omega}$ is the 3D vorticity vector, $\rho$ is air density, and $\chi$ is some scalar function of only density and pressure conserved in adiabatic, inviscid flow. Here, the planetary vorticity in $\mathbf{\omega}$ has been neglected, as it is much smaller than the relative vorticity generated during wave attenuation in the WRF simulations (not shown). The conservation equation for $P$ is

$$\frac{DP}{Dt} = \mathbf{\omega} \cdot \nabla \chi + \nabla \cdot (\nabla \times \mathbf{F}),$$

(4)

where $\dot{\chi} = D\chi/Dt$ represents sources and sinks of $\chi$ and $\mathbf{F}$ is the frictional force per unit mass (Vallis 2006). When potential temperature $[\theta = T(p_0/p)^{R_d/c_p}]$ is chosen for $\chi$, $P$ is referred to as Ertel’s PV (EPV).

EPV is conserved assuming the flow is adiabatic ($\dot{\theta} = 0$) and inviscid ($\mathbf{F} = 0$). Within the stratosphere, the flow can be regarded as adiabatic given little latent heating and radiative divergence over the short time scales considered. The stratosphere is also often considered to be inviscid, implying EPV conservation. However, wave attenuation will cause turbulent heat and momentum fluxes, invalidating EPV conservation. Within WRF, these turbulent fluxes are not resolved, but are instead represented by the planetary boundary layer (PBL) scheme. In the selected PBL scheme (ACM2; Pleim 2007), subgrid-scale mixing is local above the PBL. That is, vertical fluxes of some quantity are proportional to the vertical gradient of the quantity multiplied by an eddy diffusivity. This eddy diffusivity is primarily a function of the Richardson number ($R_i$), increasing with decreasing $R_i$ when $R_i < 0.25$. Steepening mountain waves may locally reduce $R_i$, causing the turbulence parameterization to rapidly restore static and dynamic stability and invalidate local EPV conservation. Within the stratosphere, EPV generation is a useful GW attenuation diagnostic as there are few other phenomena that might invalidate EPV conservation.

However, EPV has a strong dependence on altitude, making it difficult to quantitatively compare EPV perturbations at different altitudes. Here, we use the modified potential vorticity suggested by Lait (1994) by choosing

$$\chi = \frac{\theta}{-\alpha + 1} \left( \frac{\theta}{\theta_0} \right)^{-\alpha}.$$  

(5)

Substituting Eq. (5) into Eq. (3) gives

$$\Pi = \text{EPV} \left( \frac{\theta}{\theta_0} \right)^{-\alpha},$$

(6)

where $\theta_0 = 350$K and $\alpha = 1 + c_p/R_d \approx 9/2$. $c_p$ is the specific heat capacity, and $R_d$ is the gas constant for dry air. The scaling $\theta_0$ allows this modified PV to have the same units as EPV and $\alpha$ is chosen to reduce the dependence of EPV on altitude assuming an isothermal atmosphere [see Lait (1994) and Müller and Günther (2003) for details on choices for $\alpha$]. Unfortunately, $\Pi$ still increases significantly with height across the tropopause. This choice for $\chi$ satisfies the condition that $\chi = \chi(p, \rho)$ so that $\Pi$ is materially conserved for adiabatic, inviscid flow similar to EPV [i.e., Eq. (4) is still valid].

The horizontal variance of $\Pi$ is shown as a function of height and time in Fig. 8i. Within the stratosphere, the vertical and temporal structure of $\Pi$ variance closely follows that of the zonal GWD (Fig. 8e). Regions of GWD, diagnosed from the filtered $u'$ and $w'$, are also regions of enhanced $\Pi$ variance, a dynamical quantity representing local flow response to wave attenuation.
FIG. 8. (left) Height–time and (right) time-averaged plots of (a),(b) area-averaged zonal wind, (c),(d) area-averaged MF, (e),(f) zonal GWD, (g),(h) area of NLR > 0.25, and (i),(j) modified PV variance within the long run over the boxed region in Fig. 1b.
The frequent peak in II variance near 12 km (Figs. 8i and 8j) is associated with vertical GW motions and PV advection at the tropopause. The local characteristics of stratospheric wave attenuation are investigated in the following subsection.

c. Local mountain wave attenuation characteristics

In this subsection, the local characteristics of wave attenuation are investigated within a 2-km-resolution event run for the 24 June 2014 mountain wave event. The event run presented here differed from those presented in section 2 in that a 5-arc-min (=10-km resolution) DEM was used to derive model terrain instead of the 30-arc-s DEM in order to reduce stationary resolution-scale (2–5Δx) spurious II perturbations over the highest model terrain (see the appendix for details).

The 24 June event is characterized by low-level winds increasing with height, a tropopause-level jet, and winds decreasing to near 10 m s\(^{-1}\) in the lower stratosphere (Fig. 10a). There was little mountain wave attenuation near 12 km within the tropopause layer (Figs. 10b and 10c). Mountain waves were strongly and nonlinearly attenuated in the valve layer above this jet, with zonal GWD peaked near 16.5-km altitude (Figs. 10b–d). Stratospheric II variance (Fig. 10e) closely followed both zonal GWD and regions of nonlinearity.

In Figs. 11a and 11b, II fields at 12 and 16 km, respectively, are shown, valid at 2300 UTC 24 June 2014. At z = 12 km, where there is little attenuation, there is also little PV generation and variance over the South Island of New Zealand. At z = 16 km, however, many PV banners (elongated filaments of anomalous PV; e.g., Schär et al. 2003) originating above the New Zealand terrain are apparent. Positive and negative II is generated to the south and north of regions of low Ri, respectively. Low Ri indicates regions of high eddy diffusivity (i.e., significant subgrid-scale mixing and transport of heat and momentum) and mountain wave momentum deposition through the turbulence parameterization. Mountain wave attenuation is horizontally inhomogeneous, which is likely a result of the horizontal inhomogeneity of the underlying terrain forcing horizontally varying wave amplitudes.

The vertical and cross-mountain distribution of II variance and minimum Ri is shown in Fig. 12. Regions of high PV variance are collocated with regions of low minimum Ri. As the mountain waves steepen at these altitudes, the mountain waves induce upstream tilted regions of low Ri, wave attenuation, and PV generation. Note that a patch of turbulence upstream of a cold updraft, thought to be due to an overturning mountain wave, was observed at z = 13.5 km and t = 1200 UTC by the NGV [Fig. 17 of Smith et al. (2016)], validating to some extent the lower region of wave breaking in Fig. 12.

d. What controls deep propagation?

Mountain wave events within the well-validated long run are frequently strongly attenuated within the valve layer between 15 and 25 km. Qualitatively, this attenuation is associated with wind minima (Fig. 8), leading to wave steepening and nonlinear attenuation. However, the incident wave amplitude is important, as large-amplitude vertically propagating waves may be close to breaking, making them more likely to break down at a given altitude than a smaller-amplitude wave. Kaifler et al. (2015) investigated how both of these factors influence the propagation of mountain waves into the mesosphere, finding that mesospheric GW potential energy densities increase with increasing low-level winds and with increasing minimum stratospheric winds to a point, beyond which GW energy densities decreased. Both controls are investigated within the long run here.

Area-averaged MF\(_x\) at z = 12 and 25 km are plotted versus area-averaged 4-km zonal wind in Figs. 13a and 13b, respectively. There is some relation between the low-level winds and 12-km MF\(_x\), but it is not linear as expected from linear theory and there is significant
scatter. This situation is worse for the 25-km MF; low-level winds may be useful in predicting the maximum MF, but there is more scatter due to valve-layer attenuation.

Area-average MF at the top of the valve layer ($z = 25$ km) is plotted versus the minimum zonal wind speed $U_{\text{min}}$ within it in Fig. 13c. This quantity strongly controls MF transmitted through the valve layer, with MF well approximated ($R^2 = 0.81$) by a function of $U_{\text{min}}^3$:

$$\text{MF} = -1.74 \times 10^{-3} (\text{g s}^{-4}) U_{\text{min}}^3,$$  \hspace{1cm} (7)

where the units of MF are millipascals. This result is consistent with the Lindzen (1981) and similar (e.g., McFarlane 1987) GWD parameterizations, which propose that momentum fluxes of saturated waves (i.e., waves with amplitudes constrained to lie at the amplitude threshold for neutral stability) are most strongly constrained by wind speed and are proportional

Fig. 10. As in Fig. 8, but for the 24 Jun 2014 event simulated at 2-km resolution. Note that color scales here are different from those in Fig. 8.
to the wind speed cubed. Note that this is also true in the Palmer et al. (1986) parameterization, where amplitudes are constrained such that wave-induced Ri are not less than 0.25. The lack of weak \( |\text{MF}_x| \) when valve layer winds are strong in Fig. 13c suggest that even under these favorable deep propagation conditions, mountain wave amplitudes are typically large enough to become attenuated within the valve layer.

Quantitatively, the saturated (i.e., maximum) \( \text{MF}_{x,\text{sat}} \) within the McFarlane (1987) parameterization is given by [from his Eq. (2.31)]

\[
\text{MF}_{x,\text{sat}} = -\frac{F_c^2 e k \bar{p} U^3}{N},
\]

where \( F_c \) is a critical local Froude number specifying when a wave saturates (i.e., breaks), \( e \) is an efficiency or intermittency factor, \( k \) is the angular horizontal wavenumber, and \( N \) is the buoyancy frequency; \( F_c, e, \) and \( k \) are free tuning parameters, with \( F_c = 0.5 \) and \( e k = 1.6 \times 10^{-5} \text{m}^{-1} \). This relation provides area-averaged momentum flux, representative of some mountainous area. To compare the parameterization [Eq. (8)] to the fit within WRF [Eq. (7)], the parameterization relation and the fit were multiplied by the area of the Southern Alps (approximated as one-third the area of the South Island, 50 146 km\(^2\)) and the averaging area (426 414 km\(^2\)), respectively, to compute area-integrated \( \text{MF}_x \). Also, values of 0.1 kg m\(^{-3}\) and 0.02 s\(^{-1}\) were used for \( \bar{p} \) and \( N \) in Eq. (8). This produced coefficients (i.e., \( a \) in \( \text{MF}_{x,\text{sat}} = a U^3 \)) of \(-5.01 \times 10^5 \text{kg m}^{-2}\) and \(-7.42 \times 10^5 \text{kg m}^{-2}\) for the parameterization and fit within WRF, respectively. Despite the crude approximations used within the parameterization and assumptions made here, these values are quite similar, validating the saturation hypothesis and tuning methodology to some extent. In the following section, parameterized \( \text{MF}_x \) and GWD\(_x\) within MERRA and MERRA2 are more rigorously compared against the long run.

Initial wave amplitude may also play a role in valve layer transmission. In Fig. 13d, the transmission ratio,

\[
\text{TR} = \frac{\text{MF}_x(z = 25 \text{km})}{\text{MF}_x(z = 12 \text{km})},
\]

is plotted versus 12-km \( \text{MF}_x \) and minimum lower-stratospheric zonal wind in Fig. 13d. The highest TR (\( \approx 85\% \)) are seen in the weakest events (12-km \( \text{MF}_x \) > \(-40 \text{mPa}\)), with TR decreasing with increased amplitude. In the stronger events (12-km \( \text{MF}_x \) \( \approx -150 \text{mPa}\)), TR is a strong function of minimum stratospheric wind speed, with TR varying from less than 10\% to \( \approx 50\% \) with minimum stratospheric wind speeds from \( \approx 10 \) to \( \approx 30 \text{m s}^{-1} \). While fractional attenuation depends on both wave amplitude and minimum stratospheric wind

Fig. 11. Horizontal sections of modified PV [Eq. (6)] plotted at (a) 12- and (b) 16-km levels of little and maximum mountain wave attenuation (Fig. 10c), respectively. Regions where Ri < 0.25 are shown in green. Both panels are valid at 2300 UTC 24 Jun 2014.
speed, the tight cubic relation shown in Fig. 13c suggests minimum stratospheric wind speed plays the primary role in governing the amount of wave momentum transmitted through the valve layer.

### e. Global context

This valve layer plays an important role in determining the amount of MF transmitted to the upper stratosphere and middle atmosphere over New Zealand, but does a valve layer exist elsewhere? Zonal-averaged, time-averaged zonal winds and parameterized total (orographic plus nonorographic) zonal GWD within MERRA are shown in Fig. 14a. The averaging period is June and July over the years 2011–15. Between 15° and 50°S and between 15- and 25-km altitude, there is a maximum in zonal GWD associated with a wind minimum above the subtropical jet. This GWD maximum is primarily due to the orographic GWD parameterization, as this region is not far above the nonorographic source level of 400 hPa (Molod et al. 2015) and non-orographic waves are typically launched with amplitudes small enough such that they break in the upper stratosphere and mesosphere. That is, this region of weak winds above the subtropical jet is a mountain wave valve layer, causing mountain waves to attenuate and deposit momentum. This valve layer is also present in the Northern Hemisphere during boreal winter (not shown).
shown), suggesting a valve layer similar to that over New Zealand exists above the wintertime subtropical jet (i.e., between latitudes of $15^\circ$ and $50^\circ$ and between $z = 15$ and $25$ km) over major mountain ranges in both hemispheres.

4. Comparison of resolved and parameterized

GWD

Here total (orographic plus nonorographic) parameterized $\mathbf{M}_x$ and zonal GWD within MERRA and MERRA2 are quantitatively compared with those in the well-validated long run. $\mathbf{M}_x$ was computed from zonal GWD (Fig. 15b) using Eq. (2) by assuming zero $\mathbf{M}_x$ at the uppermost model level ($0.1$ hPa) and integrating downward. The general circulation model (GCM) in both MERRA and MERRA2 uses the same orographic (McFarlane 1987) and nonorographic (Garcia and Boville 1994) GWD parameterizations; however, the tuning of both parameterizations was changed in MERRA2 (Molod et al. 2015). The most relevant change was to the efficiency or intermittency factor,  in Eq. (8), which was kept the same as MERRA in the Northern Hemisphere ($\varepsilon = 0.125$), increased by a factor of $2.5$ south of $40^\circ S$ and ramped in between (M. J. Suarez 2016, personal communication).

MERRA $\mathbf{M}_x$ and zonal GWD are shown in Fig. 15 for the same period and area considered in the long run in Fig. 8. Similar figures for MERRA2 are omitted as they were qualitatively similar to those for MERRA (though with larger, i.e., more negative, values). As in the long run, the parameterized momentum flux and GWD are primarily orographic given the small area of interest centered on the South Island (boxed region in Fig. 1b). The vertical and temporal structure of both $\mathbf{M}_x$ and zonal GWD are quite similar between WRF and MERRA (cf. Figs. 8c and 8e with Figs. 15a and 15b). Both deep events and valve-layer attenuated shallow events in WRF are also evident in the parameterized fields. Note that the agreement is only qualitative, as color contours in Fig. 15 are a fourth of those in Figs. 8c and 8e. The troposphere is a region of notable disagreement between the simulated and parameterized zonal GWD.

The time-averaged profiles of long run, MERRA, and MERRA2 area-averaged $\mathbf{M}_x$ and zonal GWD are shown in Fig. 16. Here, the resolved quantities in MERRA2 are also provided, computed by interpolating MERRA2 fields onto the long-run grid and repeating the method used to compute quantities in the long run. This was not possible in MERRA, as vertical velocity is only provided on a reduced-resolution grid ($1.25^\circ \times 1.25^\circ$); however, MERRA- and MERRA2-resolved quantities should be similar given the similar resolutions. MERRA2-resolved $\mathbf{M}_x$ was smaller than that parameterized in both MERRA and MERRA2. The MERRA2-resolved GWD was larger than that parameterized in both MERRA and MERRA2 in the troposphere.
but still significantly smaller than that in the long run. In the stratosphere, however, resolved MF$_x$ and zonal GWD were negligible.

In the troposphere, parameterized MF$_x$ in both MERRA and MERRA2 is significantly smaller than in the long run and nearly constant with height. This may be due to the constraint that orographic GW source amplitude is limited to the saturation value (McFarlane 1987). As the wind speed typically increases with height (e.g., Fig. 8b), GWs launched at the saturation amplitude might not saturate and attenuate in the troposphere [Eq. (8)], leading to the near conservation of parameterized MF$_x$ with height in Fig. 16a. Given little parameterized momentum deposition in the troposphere, parameterized GWD is significantly smaller than that simulated in the long run.

In the valve layer ($z = 12$–$27$ km), wave attenuation is represented in both the long-run-resolved and MERRA-/MERRA2-parameterized fields. Here, MF$_x$ is 2–4 times smaller in MERRA than in the long run (Fig. 16a). At $z = 12$ km, where long-run MF$_x$ was nearly unbiased relative to aircraft observations (section 2b), the momentum fluxed into the stratosphere in MERRA was about 3.5 times smaller than in the long run. MERRA zonal GWD was 3–6 times smaller than the long run in the valve layer. In MERRA2, MF$_x$ and zonal GWD are similar to the long run or slightly underrepresented in the valve layer. These results suggest the mountain wave momentum fluxed into the stratosphere and the lower-stratospheric zonal GWD, at least over New Zealand, may be significantly underrepresented within the MERRA GCM and potentially other GCMs using similar parameterizations and tuning criteria. Increasing the parameterized source MF$_x$, as performed in MERRA2 by increasing $\varepsilon$ in Eq. (8), may resolve these issues in the valve layer.

Further evidence of this underrepresentation of valve layer GWD in MERRA is provided by Fig. 14b, showing the zonal, time mean zonal wind analysis increments from MERRA for June and July of 2011–15. Analysis increments are model errors (i.e., the difference between a 6-h forecast by the underlying GCM and an analysis of observations at the same time) divided by the time integrated to express the errors as a tendency.

![Fig. 15. Height–time plots of area-averaged total (orographic plus nonorographic) (a) parameterized MF$_x$ and (b) zonal GWD from MERRA over the same region in Fig. 8. Note that the color contours in Fig. 8c and 8e are 4 times those in this figure.](image-url)
These tendencies are assumed constant and are applied as an additional forcing term in the governing equations to dynamically correct the GCM predictions. Similar analysis increments within the stratosphere were interpreted by McLandress et al. (2012) as missing GWD. In the midlatitude, lower-stratospheric valve layer, negative analysis increments are roughly collocated with negative zonal GWD (e.g., at 30°S, z = 18 km in Figs. 14a and 14b). There are other regions outside of the valve layer where this is also true. These increments also suggest valve-layer GWD is underrepresented within the MERRA GCM. Further, the analysis increments within this region are 4–8 times the MERRA GWD (e.g., increments and zonal GWD of $-2^{-1}$ and $-2^{-4}$ m s$^{-1}$ day$^{-1}$, respectively, near 30°S and $z = 17$ km), roughly consistent with the GWD comparison between MERRA and the long run.

Above the valve layer ($z = 30–35$ km), MERRA2 parameterized MF is rather sharply deposited, resulting in very large zonal GWD ($\sim-12$ m s$^{-1}$ day$^{-1}$) relative to long run and MERRA (Fig. 16). This strong deviation from MERRA may be partially due to the increase in $\varepsilon$ in MERRA2, as the nonorographic parameterization was unchanged poleward of $\pm 20^\circ$ latitude (Molod et al. 2015). This tuning parameter modification appears to have resolved the underrepresentation of valve layer GWD in MERRA but also introduced an overrepresentation of GWD by an order of magnitude near $z = 35$ km. We speculate that this increase in GWD is associated with the fact that increasing $\varepsilon$ increases both the source and saturated parameterized momentum fluxes. This modification results in more MF into the valve layer and more GWD within it but also increases the MF and wave amplitude transmitted through the valve layer. These larger momentum fluxes and wave amplitudes then allow for more GWD aloft. Modifying $\varepsilon$ is inconsistent with Fig. 13c, which suggests MF transmitted through the valve layer is independent of MF below it. The overrepresentation of GWD above the valve layer might be addressed by removing the dependence on $\varepsilon$ from the saturated MF relation in GWD parameterizations.

The representation of GWD in GCMs has important implications for simulations of stratospheric and, indirectly, tropospheric climate. For example, increasing the Southern Hemisphere source and saturated orographic momentum flux in the MERRA2 GCM was reported to hasten the springtime breakdown of the stratospheric polar vortex (Molod et al. 2015), bringing the timing of this breakdown closer to observations. Further, Sigmond and Scinocca (2010) found that similarly increasing orographic GWD significantly decreased the strength of the stratospheric polar vortex by altering planetary wave propagation and increasing planetary wave drag on the vortex jet. In addition to addressing the missing GWD near 60°S presented by McLandress et al. (2012), this previous work suggests careful tuning and modification of orographic GWD parameterizations may also help in reducing the common cold-pole problem in chemistry–climate models, and the results presented here provide some justification and guidance for doing so.

5. Conclusions

New Zealand mountain wave propagation into the upper stratosphere is controlled by wave attenuation in a lower-stratospheric mountain wave “valve layer.” A valve layer is distinct from a critical level and is defined as a weak-wind layer containing no wind reversal. In a valve layer, mountain wave momentum flux transmission is primarily controlled by ambient wind speed. During the DEEPWAVE field campaign in May–July 2014, a valve layer was typically present in the lower stratosphere over New Zealand allowing a detailed study of wave attenuation within such a layer. High-resolution, deep ($z \approx 0–45$ km) WRF simulations, constrained by operational ECMWF analyses and extensively validated by research aircraft, radiosonde, and satellite observations, were used to reproduce observed mountain wave
events and study mountain wave dynamics within this valve layer.

Locally, mountain waves induce regions of low Richardson number (Ri), which cause momentum deposition through the subgrid-scale turbulence parameterization representing unresolved heat and momentum transport in attenuating GWs. Attenuation is horizontally inhomogeneous, likely as a result of horizontal variations of terrain height forcing horizontally varying wave amplitudes. On either side of the many localized regions of low Ri, positive and negative PV anomalies are generated. These anomalies are conservatively advected downstream generating numerous PV banners. Wave attenuation is also inhomogeneous in the vertical and is mostly confined to wave-induced upstream-tilted regions of low Ri.

Within the long run, some mountain wave events are deep (>30 km), while many events are shallow with mountain waves strongly attenuated in a lower-stratospheric valve layer. This valve layer is a climatological feature, present above the wintertime subtropical jet (between latitudes of 15° and 50° and between z = 15 and 25 km) in both hemispheres. Both minimum stratospheric wind, which causes waves to steepen and attenuate, and wave amplitude play a role in determining the percentage of MF transmitted through the valve layer. However, MF above the valve layer is primarily controlled by the minimum wind speed within the valve layer alone (i.e., MF ~ U_{min}^3), consistent with predictions of linear saturation theory. The 20–21 June 2014 event was the most significant deep event during the DEEPWAVE project (Figs. 8c and 7) and had the strongest valve-layer wind speeds along with appreciable low-level winds forcing the waves (Fig. 8a).

The McFarlane (1987) gravity wave drag (GWD) parameterization used in MERRA and MERRA2 was able to qualitatively reproduce the vertical and temporal structure of MF and zonal GWD over New Zealand within the well-validated long run. However, the long-run-simulated zonal GWD was 3–6 times stronger than that parameterized in MERRA in the valve layer. MERRA zonal wind analysis increments were consistent with this result, suggesting zonal GWD was underrepresented by a factor of 4–8 in the lower-stratospheric valve layer above the wintertime subtropical jet. This underrepresentation of valve-layer momentum flux and GWD in MERRA was significantly reduced in MERRA2 by increasing both the source and saturated momentum fluxes by a factor of 2.5 south of 40°S. However, a significant overrepresentation of GWD was introduced aloft (z = 30–35 km). Future tuning and modification of GWD parameterizations is warranted and should be constrained and guided by direct observations and high-resolution modeling.

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APPENDIX

Spurious Resolution-Scale PV Perturbations

Within event runs using the 30-arc-s digital elevation model, large-amplitude stationary resolution-scale (∼2–5Δx) potential vorticity (PV) perturbations were typically produced on all model levels below 25 km over the highest New Zealand terrain (not shown). These PV perturbations were interpreted as spurious and attributed to the added discretization error introduced when high-amplitude terrain features are poorly resolved and retained in the terrain following model levels (e.g., Schar et al. 2002). The amplitude of the topography retained in the terrain following model levels decreases with height but was only reduced by 10% in the lower stratosphere in these deep (z ≲ 0–45 km) simulations. Using the smoother 5-arc-min DEM largely removed the spurious II anomalies. Unfortunately, this reduced the model terrain heights and removed an important part of the terrain spectrum. However, stratospheric momentum fluxes and zonal GWD were nearly unaffected.

REFERENCES


Schröter, C., D. Leuenberger, O. Fuhrer, D. Lüthi, and C. Girard, 2002: A new terrain-following vertical coordinate formulation for


