The Momentum Budget in the Stratosphere, Mesosphere, and Lower Thermosphere. Part II: The In Situ Generation of Gravity Waves

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ABSTRACT

The contributions of gravity waves to the momentum budget in the mesosphere and lower thermosphere (MLT) is examined using simulation data from the Ground-to-Topside Model of Atmosphere and Ionosphere for Aeronomy (GAIA) whole-atmosphere model. Regardless of the relatively coarse model resolution, gravity waves appear in the MLT region. The resolved gravity waves largely contribute to the MLT momentum budget. A pair of positive and negative Eliassen–Palm flux divergences of the resolved gravity waves are observed in the summer MLT region, suggesting that the resolved gravity waves are likely in situ generated in the MLT region. In the summer MLT region, the mean zonal winds have a strong vertical shear that is likely formed by parameterized gravity wave forcing. The Richardson number sometimes becomes less than a quarter in the strong-shear region, suggesting that the resolved gravity waves are generated by shear instability. In addition, shear instability occurs in the low (middle) latitudes of the summer (winter) MLT region and is associated with diurnal (semidiurnal) migrating tides. Resolved gravity waves are also radiated from these regions. In Part I of this paper, it was shown that Rossby waves in the MLT region are also radiated by the barotropic and/or baroclinic instability formed by parameterized gravity wave forcing. These results strongly suggest that the forcing by gravity waves originating from the lower atmosphere causes the barotropic/baroclinic and shear instabilities in the mesosphere that, respectively, generate Rossby and gravity waves and suggest that the in situ generation and dissipation of these waves play important roles in the momentum budget of the MLT region.

1. Introduction

The climatological structure of the temperature in the middle atmosphere is much different from that expected for a radiative equilibrium. The cold summer mesopause and the warm winter stratosphere are particularly outstanding (e.g., Becker 2012). Such a difference is attributed to the forcing on the mean fields from atmospheric waves, particularly those from gravity waves (GWs) (Plumb 2002). However, in the mesosphere and lower thermosphere (MLT), thermal tides (TWs) and Rossby waves (RWs) also have large amplitudes. Thus, it is important to examine each wave’s contribution to the momentum budget, including the wave–mean flow interaction and the wave–wave interaction, to better understand the climatology and variability in the MLT region.

GWs are mainly generated by various sources, such as topography, spontaneous balance adjustment (Plougonven and Zhang 2014; Yasuda et al. 2015), fronts (Snyder et al. 1993), convection (Sato 1993; Alexander et al. 1995), and the baroclinic jet–front systems in the moist atmosphere (Wei and Zhang 2014; Wei et al. 2016) in the troposphere. GWs may also be radiated from the shear instability of large-scale flows. Bühler et al. (1999) examined the GW emissions from shear instability using linearized f-plane
Boussinesq equations and showed that GWs are radiated in four directions, namely, eastward upward, eastward downward, westward upward, and westward downward, from the unstable region. Such a process may be an important source of GWs in the summer lower stratosphere because the GWs with eastward-upward group velocities can propagate into the summer mesosphere without encountering a critical level in the westward vertical shear (Bühler and McIntyre 1999). A condition of the shear instability sometimes also holds when wave breaking occurs. Using a two-dimensional numerical model, Satomura and Sato (1999) showed a secondary generation of GWs via convective instability and/or shear instability in the region where mountain waves break. However, the possibility of such a secondary generation of GWs has not been well examined in the MLT region. It is necessary to confirm these secondary GW generations in the MLT region using models.

The important role of GWs in the MLT momentum budget was indicated directly and/or indirectly by previous studies using observations by satellites and radars, and a limited number of GCMs that include the MLT region. Ern et al. (2011) examined the distributions of GW momentum flux in the stratosphere and mesosphere using Cryogenic Infrared Spectrometers and Telescopes for the Atmosphere (CRISTA), the High Resolution Dynamics Limb Sounder (HIRDLS), and the Sounding of the Atmosphere Using Broadband Emission Radiometry (SABER). It was shown that the GWs propagate poleward in the summer mesospheric jet and equatorward in the polar night jet. On the other hand, the ground-based radar observations allow us to estimate the GW forcing and the propagating direction. Reid and Vincent (1987) showed using medium-frequency (MF) radar located near Adelaide, Australia (35°S, 138°E), that GW forcing is 100 m s$^{-1}$ day$^{-1}$ near $z = 82$ km, and decreases with height and reaches about $-100$ m s$^{-1}$ day$^{-1}$ at $z \sim 98$ km in July. Seasonal variation in the GW forcing were also shown by using the meteor radar and the MF radar (e.g., De Wit et al. 2015; Placke et al. 2015). The global distribution of GW forcing has been examined by high-resolution GCMs. Using a GW-resolving GCM without including the GW parameterizations that cover the altitudes up to the upper mesosphere, Watanabe et al. (2008) separately showed the Eliassen–Palm (EP) flux and its divergence (EPFD) associated with planetary waves, synoptic waves, and GWs. The EPFDs in the summer and winter mesosphere and the summer stratosphere are mainly attributed to GWs, while the planetary waves are most dominant in the winter stratosphere. McLandress et al. (2006) showed that the eastward (westward) wave forcing with the zonal wavenumber $s > 4$ is observed from $z = 110$ to $120$ km in the winter (summer) MLT region using the extended Canadian Middle Atmosphere Model (CMAM). Note that the large wavenumber components in CMAM may be due to resolved GWs, as shown by the present study, but they did not indicate the possibility of GWs. Karlsson and Becker (2016) also showed that the wave forcing with wavenumber $s > 6$ is large in the MLT region, and this forcing includes the resolved GW forcing. These resolved GWs may be secondarily generated by the resolved wave drag. In the thermosphere, Miyoshi et al. (2014) examined wave forcing by the resolved GWs propagating upward from the lower atmosphere using a high-resolution whole-atmosphere model in detail. The resolved GW forcing has large positive values in the summer northern thermosphere; this forcing reaches $230$ m s$^{-1}$ day$^{-1}$. Miyoshi et al. (2015) also showed that this forcing modifies the thermospheric meridional circulation and is changed by a stratospheric sudden warming (SSW) event.

The mean flow is also modulated by momentum deposition through wave–wave interactions. Miyahara and Forbes (1994) showed that GWs propagating from the lower atmosphere break in the strong-shear region, which is caused by diurnal migrating tides, and that the magnitude of the GW forcing is modulated by the phases of migrating tides. Similar results for diurnal and semidiurnal tides are shown by Liu et al. (2014).

In a companion paper (Sato et al. 2018, hereafter referred to as Part I), we examined the momentum budget for the stratosphere to the lower thermosphere, focusing on the RW, GW, and TW contributions during all seasons using 11 years of simulation data from a whole-atmosphere model. For RWs, as is well known, EP flux is upward and equatorward oriented, and its divergence is negative in the winter stratosphere. One of the important findings is that RWs are radiated through the barotropic (BT) and/or baroclinic (BC) instabilities in the mesosphere and contribute significantly to the momentum budget in the MLT region. These characteristics are in good agreement with the satellite observations [e.g., using SABER observations (Ern et al. 2013) and Aura MLS observations (Part I, their Fig. 6)]. The BT/BC instability in the mesosphere is almost always observed and maintained by parameterized gravity wave forcing (hereafter referred to as GWF$P$). Another important finding is that, in spite of the relatively coarse horizontal resolution of the model, a significantly resolved GW forcing in the MLT region is observed. Very recently, Karlsson and Becker (2016), using the Kühlingsborn Mechanistic General Circulation Model (KMCM), also suggested small resolved GW forcing in the MLT region.

Our study examines the characteristics and generation mechanisms of the resolved GWs in the MLT region.
using simulation data from the whole-atmosphere model called the Ground-to-Topside Model of Atmosphere and Ionosphere for Aeronomy (GAIA) (e.g., Jin et al. 2011). The most probable generation mechanism will be shown to be shear instability in the MLT region. The roles of GWs originating from the lower atmosphere and the tides in the MLT region are discussed.

The remainder of this paper is organized as follows. A brief description of the data from GAIA is given in section 2. A method of analysis is described in section 3. The results of the momentum budget evaluation when dividing the resolved GWs into eastward- and westward-propagating components are shown in section 4. The in situ generation mechanisms of the resolved GWs in the MLT region are examined in section 5. Section 6 presents confirmation of the in situ GW generation in the MLT region by using data from higher-resolution model simulations over a limited time period.

2. Data description

a. GAIA simulation data

The GAIA simulation data are briefly introduced here. Details are given in Part I. GAIA is a coupled neutral and ionized atmosphere model that covers an altitude range from the ground to the thermosphere/ ionosphere. In this study, data from the neutral atmosphere model part (i.e., the GCM part) (e.g., Miyoshi and Fujiwara 2003) are used. The resolution of the GCM is T42L150, which has a horizontal grid point at every 1.1° and a vertical grid point at every 0.2 scale height for the range from the surface to a height of $z \sim 600 \text{km}$. The height range analyzed in this study is $z = 0 \sim 113 \text{km}$. Accuracy of the model data below $z \sim 97 \text{km}$ was confirmed by comparing the data with the Aura MLS observations (see Part I). The structure of the zonal-mean temperature and zonal wind in the GAIA results is quantitatively similar to the satellite observations (e.g., Ern et al. 2013; Part I). See Part I for the details. A time interval of the model output is 1 h. The momentum deposition of the subgrid-scale GWs is represented using the GW parameterizations of McFarlane (1987) and Lindzen (1981) for orographic and nonorographic GWs, respectively. The GWF$_P$, which has a horizontal grid point at every $1.1^\circ$ and the same vertical grids as those of the standard GAIA (Miyoshi et al. 2014, 2015). Note that nonorographic GW parameterization is not included, but the orographic GW parameterization by McFarlane (1987) is implemented in HRGAIA. The distributions of the water vapor and clouds are predicted in the model. As the lower boundary condition of this HRGAIA, climatologies are given for the mean sea surface temperature, ground wetness, and sea ice distribution. The analyzed height region is the same as that of the standard GAIA data, and the analyzed period is the month of January. Note that the HRGAIA is not nudged by reanalysis data. The HRGAIA is used here only for the purpose of validation for the existence of gravity waves as simulated in the standard GAIA.

3. Methods of analysis

To examine the characteristics of the resolved GWs and the mechanisms of the in situ generation of GWs in the MLT region, the resolved GWs are extracted from the model simulation data as follows. First, the TWs defined as a sum of the migrating tides with $(s, \tau) = (1, 24), (2, 12),$ and $(3, 8) \text{ h}$, where $\tau$ denotes a period, were removed. Next, the resolved GWs are extracted as disturbances with periods of $\tau \leq 24 \text{ h}$. Note that nonmigrating tides are included in the resolved GWs because at least a part of the nonmigrating tides are regarded as global-scale inertia GWs (Sakazaki et al. 2015). The resolved GWs are divided into GWs with eastward- and westward-propagating components using a Fourier transform. The contributions of the eastward- and westward-propagating resolved GWs to the momentum budget are analyzed separately for EP flux and EPFD in the transformed-Eulerian-mean equation (Andrews et al. 1987) (see Part I for details).
4. Resolved gravity wave contribution to the momentum budget in the MLT region

Figures 1a and 1b show the latitude–height sections of the zonal-mean temperature (color shading) and zonal wind (contours) climatologies in January and July, respectively. As shown in Part I, most characteristics of the mean fields are consistent with the *Aura* MLS observations, except for the missing wind reversal in the vertical around \( z \approx 86 \text{ km} \) in the summer hemisphere, although there exists a weak wind layer (see Part I, their Fig. 1). This difference may be due to the limitation of the representation of GWF\(_P\) in the model. The differences in the zonal-mean temperature and zonal wind between GAIA and *Aura* MLS observations were described in Part I in detail.

In Part I (their Fig. 3), we have already shown the characteristics of the EP flux and EPFD associated with the resolved GWs: the EP flux is oriented downward (upward), and the EPFD is positive (negative) in the summer (winter) MLT region. Note that the downward (upward) EP flux implies the dominance of the vertical flux of the eastward (westward) momentum for GWs propagating energy upward. A pair of negative and positive EPFD regions is observed in the summer hemisphere near \( \phi = 60^\circ \) and \( z = 80 \text{ km} \) and above, respectively. This result shows that the resolved GWs are likely generated in the summer MLT region. The GWs can propagate eastward and westward, and each propagating component may contribute differently to the EP flux and EPFD. Thus, to examine this pair of positive and negative EPFDs in detail, the EP flux and EPFD are divided into two namely that by resolved GWs propagating eastward and by those propagating westward.

Figure 2 shows the latitude–height sections of the EP flux and EPFD of the resolved GWs propagating eastward and westward during January and July. The net forcing to the mean zonal wind from the resolved GWs, which is in the same direction as GWF\(_P\) in the MLT region (see Part I), is found to be mainly due to the eastward (westward) GWs in the summer (winter) MLT region. It is interesting that the westward (eastward) GWs provide wave forcing in the opposite direction of the GWF\(_P\) in the summer (winter) hemisphere. The upward EP flux associated with the westward GWs in the summer hemisphere is observed above the region with a strong mean wind shear from 10ºS, \( z = 50 \text{ km} \) to 50ºS, \( z = 100 \text{ km} \) in January and with that from 0º, \( z = 50 \text{ km} \) to 45ºN, \( z = 95 \text{ km} \) in July. It is also important
FIG. 2. Latitude–height sections of the climatologies of the EP flux and its divergence of the resolved GWs propagating eastward during (a) January and (c) July and propagating westward during (b) January and (d) July. Arrows indicate EP flux (m$^2$ s$^{-1}$). Color shading represents EP flux divergence (m s$^{-1}$ day$^{-1}$). Note that the contour intervals are not uniform.
that the EPFD for the westward GWs along the wind shear is positive. These features indicate the possibility that the westward GWs are in situ generated in the mesosphere via shear instability. It is also possible that eastward-resolved GWs are in situ generated in the mesosphere, although a part of the large-scale GWs originates from the latent heat release related to diurnal variations from convection at the planetary scale in the troposphere (Sakazaki et al. 2015).

5. Gravity wave generation in the MLT region

a. Occurrence frequency of Richardson numbers smaller than 1/4

To examine the possibility of shear instability in the MLT region, the Richardson number \( R_i = N^2/u_z^2 \) is calculated. Here, \( N^2 \) and \( u_z \) denote the static stability and zonal wind, respectively. Figure 3 shows the latitude–height sections of the occurrence frequencies of shear instability (i.e., \( R_i < 1/4 \)) during January and July. The occurrence frequency is defined as the percentage of the cases in which \( R_i < 1/4 \) in January and July at each longitude, latitude, level, and time. In the middle latitudes of the summer hemisphere, the occurrence frequency of \( R_i < 1/4 \) is maximized at \( \phi = 40.5^\circ \text{S} \) and \( z = 89 \text{ km} \) (≈ 7.3%) in January and at \( \phi = 37.7^\circ \text{N} \) and \( z = 86 \text{ km} \) (≈ 5.2%) in July. The regions with large occurrence frequencies of \( R_i < 1/4 \) in the summer hemisphere tilt poleward with height. Figure 4 shows the time–latitude sections of the occurrence frequencies of \( R_i < 1/4 \) at \( z = 90–100, 80–90, \) and \( 65–75 \text{ km} \). A smoothing of a 10-day running mean was applied to more clearly identify the characteristics. The seasonal variability is different at each level. As seen in Fig. 4c, the maximum is much clearer in the middle latitudes of the summer hemisphere at this level, and another maximum occurs at approximately \( 70^\circ \text{S} \) in the middle latitudes of the winter hemisphere in June and July.

On the other hand, in Fig. 3, it is interesting that the region with high-frequency occurrences extends toward the high latitudes of the winter hemisphere but that there is a secondary maximum at \( \phi = 4.2^\circ \text{S} \) and \( z = 98 \text{ km} \) during both January and July in the summer hemisphere. These secondary maxima are observed not only in January and July but also in other months. For \( z = 90–100 \text{ km} \) (Fig. 4a), the occurrence frequencies are high in the low and middle latitudes of the summer hemisphere and in the low latitudes near the equator in the spring and autumn. Similar features are observed at
$z = 80–90$ km, but the maximum around the latitudes of $30^\circ–40^\circ$ is clearer (Fig. 4b). The maximum around the equator is not observed at $z = 65–75$ km (Fig. 4c).

In the regions of the maxima in the middle latitudes of the summer hemisphere, there is strong vertical wind shear, as seen in Fig. 1. This feature suggests that a part of the resolved GWs in the summer MLT region is likely generated in situ by shear instability. As the region with large vertical wind shear in the summer mesosphere coincides well with strong GWF$_P$, as indicated in Part I, the large occurrence frequency of Ri $< 1/4$ (i.e., strong vertical wind shear) is likely formed by the strong GWF$_P$. In contrast, there is no mean wind structure corresponding to the secondary maximum in the low latitudes and its extension into the winter hemisphere. This secondary maximum is related to the TWs, as discussed in detail in section 5c. Note that the occurrence frequency of the convective instability ($N^2 < 0$) was also examined, but it was negligible ($< 0.6\%$) (not shown).

![Fig. 4. Time–latitude sections of the occurrence frequencies of Ri $< 1/4$ at z = (a) 90–100, (b) 80–90, and (c) 65–75 km, respectively. The variabilities of the occurrence frequencies of Ri $< 1/4$ data are filtered by a 10-day running mean.](image-url)
FIG. 5. (a) Latitude–height section of the climatology of the energy flux ($\Phi'w'$) for the resolved GWs during January. (b) The latitude–height section of the ratio of $\Phi'w' < 0$ for the resolved GWs, namely, the ratio of the GWs propagating energy downward during January. (c),(d) As in (b), but for the ratio of $\Phi'w' < 0$ of the resolved GWs propagating eastward and westward, respectively. The red lines denote the occurrence frequencies of $\text{Ri} < 1/4$ in Fig. 3a.
b. Resolved GWs propagating energy downward from the region with strong vertical wind shear in the MLT

According to recent theoretical studies by Bühler et al. (1999) and Bühler and McIntyre (1999), GWs are likely radiated eastward, westward, upward, and downward relative to the background winds from a region with strong vertical wind shear. To find evidence of the GWs propagating energy downward, the zonal mean vertical energy flux ($\Phi'w'$) of the resolved GWs was examined in the latitude–height section. Figure 5a shows the result for the Southern Hemisphere (SH) in January. The energy flux is positive (i.e., upward) in most regions, except for in the high latitudes below $z = 45$ km. The ratio of negative $\Phi'w'$, namely, $R_d = \sum |\Phi'w'|/\left(\sum|\Phi'w'| + \sum|\Phi'w'|\right) \times 100$ (%), is shown in Fig. 5b as a function of the latitude and height of the SH in January. Here, $|\cdot|$ and $\sum |\cdot|$, respectively, denote the absolute values of the negative $\Phi'w'$ and those of the positive $\Phi'w'$ at each time, latitude, longitude, and height. The summation was performed over time and longitude. The color shade represents the region with $R_d \geq 18\%$. The value of $R_d$ is large below the regions with large occurrence frequencies of $R_i < 1/4$ in the low and middle latitudes. This result is consistent with our inference that the resolved GWs propagating energy downward are radiating from the strong wind shear, although the major part of the vertical energy flux is upward. Next, $R_d$ is examined for the eastward and westward GWs in Figs. 5c and 5d, respectively. The $R_d$ values are large for both the eastward and the westward GWs below the region with large occurrence frequencies of $R_i < 1/4$. However, $R_d$ of the eastward GWs is slightly smaller than that of the westward GWs. This is likely because a part of the eastward GWs is due to the GWs propagating upward without encountering their critical levels from the lower atmosphere. The $R_d$ maximum is observed below the maximum of the vertical wind shear for both the eastward and westward GWs. This fact means that at least part of both the eastward and westward GWs is generated in situ in the mesosphere.

To examine the radiation of GWs with downward group velocities from shear instability in more detail, a cospectrum of $\Phi'$ and $w'$ of the resolved GWs is calculated as a function of the ground-based phase speed at each height and is compared with the zonal mean zonal winds at the latitude of $\phi = 40.5^\circ S$ in Fig. 6a. The vertical profiles in Fig. 6b are shown for the climatologies of $N^2$, the square of the vertical wind shear, the median of $R_i$, and the occurrence frequencies of $R_i < 1/4$ at $40.5^\circ S$ during January.
of the cospectrum. The smaller downward energy flux for positive phase speed is likely because a major part of the resolved GWs has upward energy flux and hides the downward energy flux of GWs. This inference is consistent with relatively small $R_d$ for eastward GWs (Fig. 5). This fact reinforces our inference that the resolved GWs in the middle latitudes of the summer mesosphere are generated in situ by shear instability. The most likely candidate to cause the shear instability in the middle latitudes of the summer mesosphere is the forcing of GWs propagating from the lower atmosphere, which is expressed by parameterizations in the model. Note again that these results do not differ much from the results using GAIA without nudging. This study showed that the most likely candidate for the secondary generation of GWs is the shear instability in the MLT region. However, there are other possible mechanisms. For example, GWs are radiated through a spontaneous adjustment around a possible westward jet imbalance in the summer hemisphere. Direct GW emission from the given wave forcing is also likely (i.e., Zhu and Holton 1987; Becker and Vadas 2018). The relative importance of these possible generation mechanisms in the momentum budget in the MLT region is left for future studies.

FIG. 7. Latitude–height sections of the amplitudes of the (a) $T$, (b) $u$, and (c) $v$ components of DW1 and the (d) $T$, (e) $u$, and (f) $v$ components of SW2 during January as obtained from the GAIA simulation data.
c. Causes of shear instability in the summer low latitudes and winter middle latitudes of the MLT region

The occurrence frequency of $R_i^{1/4}$ is also large in the summer low latitudes and winter middle latitudes of the MLT region (Fig. 3). However, the vertical shear of the zonal-mean zonal wind is weak (e.g., Fig. 1), and $N^2$ is not as small as in the middle latitudes. Thus, another mechanism must be present for the shear instability in this region. The most likely candidate is TWs.

The vertical shear of the horizontal winds in some phases of the TWs may be sufficiently large to cause shear instability. Figure 7 shows the latitude–height sections of the amplitudes of the $T$, $u$, and $v$ fluctuations of diurnal...
westward-migrating tides with $s = 1$ (DW1) and semi-diurnal westward-migrating tides with $s = 2$ (SW2) in January obtained by the model data. The $T$ amplitude of DW1 (Fig. 7a) is maximized (~5.7 K) in the low latitudes for the height region above $z \sim 50$ km. Both the $u$ and $v$ amplitudes of DW1 (Figs. 7b,c) are latitudinally maximized ($u \sim 13.8$ m s$^{-1}$; $v \sim 20.2$ m s$^{-1}$) around $\phi = 20^\circ$ in the height region above $z \sim 60$ km. The SW2 amplitude maximum is observed in the middle latitudes and the maximum values increase with height above $z = 100$ km (Figs. 7d–f). Thus, the DW1 and SW2 strongly influence the short-period variations in the low and middle latitudes of the MLT region, respectively. Because both the $u$ and $v$ amplitudes are large for DW1 and SW2, $R_i$ is estimated using the following formula:

$$R_i = \frac{N^2}{(dV/dz)^2} \left( \frac{dV}{dz} \right)^2 + \left( \frac{du}{dz} \right)^2 + \left( \frac{dy}{dz} \right)^2.$$  

Animation S1 in the online supplemental material for this paper shows the latitude–height section of the occurrence frequency of $R_i < 1/4$ as a function of local time. Regions with large occurrence frequencies at each local time are observed in the summer middle latitudes and low latitudes of the MLT region. The maximum in the middle latitudes of the summer hemisphere is almost steady. In contrast, the large occurrence frequency region in the low latitudes propagates downward with time over a period of 24 h, suggesting that the large occurrence frequency is caused by DW1. Similar time variations of the region with the occurrence frequency of $R_i < 1/4$ but with a period of 12 h are seen in the winter MLT. This variation is attributable to SW2, which has large amplitudes in the middle latitudes above $z = 100$ km (Figs. 7d–f). Note that the $T$, $u$, and $v$ amplitudes of the diurnal tides simulated in the GAIA are in good agreement with those observed by SABER (e.g., Zhu et al. 2008), but the amplitudes of the semidiurnal tides are larger above $z = 105$ km. Thus, the occurrence frequency of shear instability associated with SW2 would be smaller than in the real atmosphere.

Figure 8a shows the latitude–height section of the variance of the diurnal variation of the occurrence frequency of $R_i < 1/4$ during January in the model. The variance is quite small in the summer MLT region. This is consistent with our inference that the shear instability in this area is due to the climatological-mean zonal wind shear caused by the parameterized GW forcing. In contrast, the variance of the local time variation of occurrence frequency of $R_i < 1/4$ is large in the low
latitudes above $z = 80$ km and in the middle latitudes of the winter lower thermosphere. These large variances are attributable to DW1 and SW2, as mentioned before. Figure 8b shows the local time and height section of the occurrence frequency of $R_i < 1/4$ (contour) and the zonal wind of DW1 (color) at 20.9$^\circ$S. The regions with maximum occurrence frequencies of $R_i < 1/4$ propagate downward at a rate of $\sim 0.6$ km h$^{-1}$. This downward speed accords well with the phase speed of the $u$ component of the DW1 in the region above $z = 90$ km. Figure 8c shows the local time and height section of the ratio of the downward energy flux $R_d$ at 20.9$^\circ$S, where the occurrence frequency of $R_i < 1/4$ has a secondary maximum in January. It is clear that $R_d$ is large below the height of the maxima of the occurrence frequency of $R_i < 1/4$ and varies with the local time. This result indicates that large-amplitude DW1 significantly contributes to the formations of shear instability in the low latitudes of the MLT region, which radiates (resolved) GWs. Similarly, SW2 contributes to the shear instability in the winter hemisphere. Figure 8d shows the $u$ component of SW2 and the occurrence frequency of $R_i < 1/4$ at 54.4$^\circ$N as a function of the local time. The occurrence frequency of $R_i < 1/4$ varies over a semi-diurnal period. The value of $R_d$ is large below the region where the occurrence frequency is maximized depending on the local time. Thus, the shear instability caused by SW2 is a likely mechanism for generating GWs in the middle latitudes of the winter hemisphere in the upper mesosphere and lower thermosphere. Note again that because the amplitudes of the semidiurnal tides simulated in the GAIA are larger than those in the observations above $z = 105$ km, it is possible that the maximum of the occurrence frequency of $R_i < 1/4$ above this level is overestimated.

### 6. Features of resolved GWs in HRGAIA

Structures of GWs resolved in the standard GAIA data may be largely distorted by relatively coarse horizontal model resolutions, although their generation mechanism is likely present in the real atmosphere. Thus, the characteristics of the GWs are also examined using the HRGAIA simulation data, although this simulation period is limited. Figure 9a shows the latitude–height sections of the zonal-mean temperatures and zonal winds in HRGAIA for January. The zonal-mean temperature field of HRGAIA is quite similar to that of the standard GAIA data. A notable difference is seen in the zonal-mean zonal wind field. The eastward jets in the SH stratosphere and mesosphere in HRGAIA are weaker, and their core is located at a higher latitude than that in the GAIA dataset. Moreover, a vertical reversal of the zonal winds near the summer mesopause is clear in HRGAIA, as is consistent with the observations (see Part I, their Fig. 1b).

Next, the EP flux and EPFD associated with the resolved GWs in HRGAIA are examined (Fig. 9b). The EP flux is downward and equatorward, and the EPFD is positive in most regions of the summer mesosphere. The EP flux is upward and poleward, and the EPFD is negative in the winter mesosphere, which is generally consistent with the features seen in the standard GAIA. A difference is observed in the lower thermosphere, where the EP flux is upward (downward) and poleward and the EPFD is negative (positive) in the summer (winter) hemisphere. This difference is probably related to the presence of the clear zero-wind layer in the MLT region in HRGAIA. However, it was confirmed that the total wave forcing due to resolved GWs and parameterized orographic GWs in HRGAIA is comparable to that due to resolved GWs and parameterized orographic and nonorographic GWs in the mesosphere in GAIA (see Fig. 3 in Part I) with an exception at high latitudes causing from the deficiency of the GW parameterization (see Part I).

The possibility of the shear instability is also analyzed as a GW generation mechanism in the MLT region in HRGAIA. Figure 10 shows the latitude–height section of the occurrence frequency of $R_i < 1/4$ for HRGAIA in
January. The distribution of the large occurrence frequency of $R_i < 1/4$ is quite similar to that in the standard GAIA with a slight difference: the maximum values of the summer middle latitudes and the low latitudes observed in the standard GAIA dataset are located, respectively, in higher and lower latitudes, in HRGAIA. The occurrence frequencies of $R_i < 1/4$ have maxima at $\phi = 56.6^\circ S$ and $z = 87$ km ($\sim9.5\%$), and $\phi = 0.6^\circ N$ and
z = 93 km (~8.1%) in the MLT region. This result suggests that the shear instability and associated in situ generation of the GWs also occurs in HRGAIA.

Next, the energy flux of the resolved GWs ($F \Phi \omega'$) is examined in the summer MLT region for January using the data from HRGAIA (Fig. 11a). The value of $F \Phi \omega'$ is positive, which is consistent with the features observed in the standard GAIA data. Figure 11b shows the latitude–height section of $R_d$. Similar to the results for the standard GAIA data analysis, large $R_d$ values are observed below the regions of the large occurrence frequencies of $R_i < 1/4$. The distribution of $R_d$ is shown separately for the eastward and westward GWs in Figs. 11c and 11d, respectively. The value of $R_d$ is large below the regions of the large occurrence frequencies of $R_i < 1/4$ in the low and middle latitudes of the summer hemisphere for both components, which is also consistent with the results for the standard GAIA.

Figure 12a shows a co-spectrum of the $\Phi'$ and $w'$ fluctuations associated with the resolved GWs and zonal-mean zonal winds at $\phi = 40.9^\circ$S. The vertical profiles in Fig. 12b show the climatologies of $N^2$, the square of the vertical wind shear, the median of $R_i$, and the occurrence frequency of $R_i < 1/4$. While $N^2$ is minimized near $z = 90$ km, the vertical wind shear is maximized near $z = 77$ km. Thus, the median of $R_i$ is minimized and the occurrence frequency of $R_i < 1/4$ is maximized around $z = 80$ km. The negative co-spectrum of $\Phi'$ and $w'$ for the eastward-resolved GWs with phase speeds $C_x = \sim 20$ m s$^{-1}$ is observed below $z = 70$ km. This feature of a negative co-spectrum of $\Phi'$ and $w'$ for eastward-resolved GWs in HRGAIA is similar to that observed in the standard GAIA data, although the phase speeds of the corresponding GWs are slower than those in the standard GAIA. This difference in the phase speed of GWs propagating energy downward between standard GAIA (Fig. 6a) and HRGAIA (Fig. 12a) data is probably related to the difference in the mean zonal wind and the simulated time period as well as the horizontal resolution. The mean zonal wind is about $-35$ m s$^{-1}$ at the height of the maximum occurrence frequency of shear instability in the standard GAIA data. In contrast, it is $\sim 0$ m s$^{-1}$ in HRGAIA. If the GWs are generated from the shear instability, the phase speeds of the GWs are expected to be distributed around the background wind speed (e.g., Fritts 1984). The distribution of phase speeds of GWs propagating energy downward is consistent with this expectation both for the standard GAIA and the HRGAIA. Note that the mean zonal wind is around zero at the level with the highest occurrence frequency of the shear instability for HRGAIA. To estimate small phase speed near 0 m s$^{-1}$, a long time period of simulated data is necessary. However, the time period analyzed for HRGAIA is only a month. These reasons might explain why GWs propagating energy downward are not clearly identified for HRGAIA.

The feature of an $R_d$ value that is large in the low latitudes of the MLT region (Fig. 11b) is consistent with the result of standard GAIA. The feature also confirms that the large $R_d$ values are related to the large occurrence frequencies of $R_i < 1/4$ caused by DW1 (not shown in detail). These results from the HRGAIA data also strongly support the in situ GW generations from the shear instability in the MLT region.

7. Summary and concluding remarks

This study examined the characteristics and generation mechanisms of the resolved GWs in the MLT region using simulation data from GAIA for over approximately 11 years, from August 2004 to June 2015,
by separating the GWs into eastward- and westward-propagating components.

According to the analysis of the occurrence frequency of Ri < 1/4, the main mechanism causing the resolved GW generation in the MLT region is likely shear instability. There are two possible causes of the shear instability. In the middle latitudes of the summer hemisphere, a large climatological vertical shear of the zonal wind is observed, satisfying the shear instability. This shear is likely formed by GWF, more specifically speaking, by nonorographic GW forcing. The dominance of the nonorographic GW forcing is confirmed by the fact that in the summer MLT region the orographic GW forcing is quite weak, and the zonal distribution of the area with high frequency of shear instability does not match that of the orographic GW forcing (not shown). This is related to the existence of the critical level for orographic GWs in the summer lower stratosphere at middle and high latitudes. The large number of GWs propagating energy downward Rd below the strong vertical wind shear supports this inference. The occurrence frequency of the shear instability is also large in the low latitudes of the summer hemisphere and in the middle latitudes of the winter hemisphere. Large-amplitude TWs, in particular, DW1 and SW2, are responsible for the shear instability in these regions. Following the DW1 phase variation, the occurrence frequency of Ri < 1/4 has strong diurnal variability in the low latitudes of the summer hemisphere. In the diurnal variation, Rd is enhanced below the regions with high occurrence frequencies of Ri < 1/4 for each local time. Similar features are observed in the middle latitudes of the winter hemisphere, but the variation is semidiurnal because of the dominant SW2. The in situ GW generations were confirmed by the analyses using simulation data over a limited period from the higher-resolution GAIA (HRGAIA).

The resolved GWs and RWs, including the QTDWs and the 4-day waves in the MLT region, the latter of which were analyzed in detail in Part I, are generated through shear and BT/BC instabilities caused by GWF, respectively. This means that the climatologies of the mean fields are strongly affected by the GWF directly and/or indirectly through in situ GW and RW excitations. Thus, it is very important to improve the GW parameterizations. Moreover, these in situ generated GWs in the MLT region may propagate into the thermosphere and contribute to the formation of the thermospheric circulation (e.g., Miyoshi et al. 2014), ionospheric disturbances (e.g., traveling ionospheric disturbances; Kelley 2009), and/or initial disturbances of the equatorial spread F (Huang and Kelley 1996). Further studies are necessary to confirm these viewpoints using whole-atmosphere models.

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