Characteristics of gravity wave horizontal phase velocity spectra in the mesosphere over the Antarctic stations, Syowa and Davis.

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Abstract

Mesospheric gravity-wave (GW) phase velocity spectra and total powers at two Antarctic stations, Davis and Syowa, were derived using OH airglow image data from March to October in 2016. The total powers have similar seasonal variation, that is, maxima in winter at both stations. However, the power at Davis was one standard deviation larger in winter and three times smaller in September than at Syowa. The total power at Davis in winter was mainly attributed to GWs with high eastward (phase velocity. On the other hand, the higher total power at Syowa in September was attributed to GWs with omnidirectional phase velocity. These differences between Syowa and Davis can not be explained by the wind filtering effect, and other factors are needed. To further explorer the origin of the difference in winter, we focused on an event on August 29, 2016, in which GWs with ~100 ms^{-1} southeastward phase velocity appeared at Davis. The raytracing method was applied, and its result indicated that those GWs with high southeastward phase velocity propagated from ~45 km altitude over the southern ocean (~ where GWs with high amplitude and southeastward propagating emitted from the tropospheric jet appeared. These jet GWs were probably saturated in 45-50 km altitudes. Therefore, the GWs with eastward phase velocity were probably secondary gravity waves. This result suggests that the higher power in the eastward high phase velocity domain at Davis was contributed to by secondary GWs.

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Key Points:

- Mesospheric gravity-wave activity at two Antarctic stations was larger in winter than in spring and fall.
- The wave activity at Davis was larger in winter than Syowa but smaller in fall, which cannot be explained by the wind filtering effect.
- We demonstrated that gravity waves with ~100 ms^{-1} eastward phase velocity at Davis were generated by primary gravity wave breaking.

Abstract

Mesospheric gravity-wave (GW) phase velocity spectra and total powers at two Antarctic stations, Davis and Syowa, were derived using OH airglow image data from March to October in 2016. The total powers have similar seasonal variation, that is, maxima in winter at both stations. However, the power at Davis was one standard deviation larger in winter and three times smaller in September than at Syowa. The total power at Davis in winter was mainly attributed to GWs with high eastward ($\geq \sim 50\ \mathrm{ms}^{-1}$) phase velocity. On the other hand, the higher total power at Syowa in September was attributed to GWs with omnidirectional phase velocity. These differences between Syowa and Davis can not be explained by the wind filtering effect, and other factors are needed. To further explorer the origin of the difference in winter, we focused on an event on August 29, 2016, in which GWs with ~100 ms^{-1} southeastward phase velocity appeared at Davis. The raytracing method was applied, and its result indicated that those GWs with high southeastward phase velocity propagated from ~45 km altitude over the southern ocean (~$43°\mathrm{E}, \sim 58°\mathrm{S}$) where GWs with high amplitude and southeastward propagating emitted from the tropospheric jet appeared. These jet GWs were probably saturated in 45-50 km altitudes. Therefore, the GWs with eastward phase velocity were probably secondary gravity waves. This result suggests that the higher power in the eastward high phase velocity domain at Davis was contributed to by secondary GWs.

1 Introduction

Gravity waves (GW) play an essential role in transporting momentum vertically and driving the circulation in the middle atmosphere (Fritts and Alexander, 2003). Current global circulation models (GCMs) cannot resolve the full spectrum of GWs so that GW influences (GW drags) are parametrized in GCMs. Although the GW drag parameterization schemes improve the ability of GCMs to reproduce observations, they still poorly represent GW sources, spectra, local variability, and intermittency, which causes deviations between GCMs and the real atmosphere (Alexander et al., 2010). Inadequate knowledge of GWs is a key element of this limitation.

In order to understand GW sources and features, many studies have investigated GWs in the real atmosphere by using in-situ and remote sensing instruments. Airglow imaging is one remote sensing method that has high sensitivity to GWs with small horizontal wavelengths (~10s km) and short periods (several minutes). Such GWs deposit substantial momentum around the mesopause (Fritts and Vincent, 1987).

Matsuda et al. (2017) investigated the spatial variability of GW horizontal phase velocity around the edge of Antarctica by using observations from four OH airglow imagers during 1.5 months (April to May) at Syowa (69°S,40°E), Davis (69°S, 78°E), McMurdo (78°S, 167°E), and Halley (76°S, 26°E). These instruments which are operated by the Antarctic Gravity Wave Instrument Network (ANGWIN) showed that GW intermittency can be caused by the wind filtering effect, and the GW activity over the Antarctic was variable in latitude (the GW activities at Syowa and Davis were ~5 times larger than those at McMurdo and Halley). They also found that GW power spectral density (PSD) in eastward phase velocity less than $\sim 70\ \mathrm{ms}^{-1}$ was ~10 times larger at Davis than at Syowa. This large PSD at Davis was exceptional because the GWs with small eastward phase velocity could not reach the airglow layer from the troposphere at Davis, as well as Syowa. The paper mentioned that a possible source of those GWs may be secondary generation above the stratosphere, although it does not robustly demonstrate the possible source. Also, it is uncertain that the large PSD in the eastward velocity is a common feature at Davis because of the short (1.5 month) dataset.

The purpose of this study is to clarify differences in phase velocity spectra between Davis and Syowa and to show statistically significant results by using a whole Austral winter season (March to October) in 2016. Chapter 2 introduces OH airglow imagers at Syowa and Davis, and data analysis. In Chapter 3, we show the differences of GW activities between Syowa and Davis, and compare our results with Matsuda et al. (2017). We discuss the effects of wind filtering on the GWs in Chapter 4 and investigate a source of GWs with eastward phase velocity at Davis.

2. Observation and Data analysis.

Two OH airglow imagers at Syowa (40°E, 69°S) and Davis (78°E, 69°S) are operated by National Institute of Polar Research (NIPR) and the Utah State University, respectively. Both imagers are equipped with indium gallium arsenide (InGaAs) detectors, which are sensitive to wavelegths between 0.9-1.7μm, avoiding a large part of the auroral contamination (0.4-0.8 μm) seen by CCD detectors (Matsuda et al., 2017). The detectors are incorporated with a Fujinon FE185C086HA-1 C-mount fisheye lens to observe the whole sky. The specification of both

Station	Latitude	Longitude	Operation Institution	Airglow[a]	Sampling Interval[b]	Exposure	Detector Size
Syowa	69° S	40° E	NIPR	OH (0.9-1.7μm)	5 s	2 s	320×256
Davis	69° S	78° E	USU	OH (0.9-1.7μm)	10 s	3 s	320×256

[a] The OH layer altitude is assumed as 87 km.
[b] OH intensity images are averaged in 1 min.

imagers is summalized in Table 1.

Table 1. Summary of the two airglow imager specifications.

This study used the data without clouds, moon, and auroral contamination lasting for more than 1 hour from March to October 2016. Table 2 shows the distribution of the data windows, and this study analyzed 39 days at Syowa and 55 days at Davis.

		Mar.	Apr.	May	Jun.	Jul.	Aug.	Sep.	Oct.	Total
Syowa	Num. of night	1	8	9	3	4	7	6	1	39
	Total length (hour)	1.0	17.9	35.7	11.4	14.8	23.5	12.2	1.3	117.5
Davis	Num. of night	6	6	7	2	8	11	10	5	55
	Total length (hour)	15.4	12.3	15.8	12.8	26.0	56.0	27.6	10.6	176.5

Table 2. The number of data windows used in this study. The criteria for choosing the data are (1) clear-sky, (2) without the moon, and (3) aurora-free images continued for > 1 hour.

To derive a GW phase velocity spectrum, an M-transform was applied to the OH airglow image data. Since the M-transform is described in Matsuda et al. (2014; 2017) and Perwitasari et al. (2018) in detail, only preprocesses are explained here. The raw image data were averaged to 1 min for compatibility between measurement cadences at Syowa and Davis and star and dark count removal was applied. To derive relative OH airglow intensity perturbations, the average night time intensity is subtracted from each 1 min image, the residual intensity (I') was divided by the nightly mean (I_0), and then I'/I_0 was defined as the relative intensity perturbation. The I'/I_0 images were projected onto geographical coordinates, assuming a mean emission height of 87 km altitude (Baker and Stair, 1988). An OH emission height can vary between 79-88 km altitudes (Grygalashvyly et al., 2014; Nishiyama et al., 2021), but this causes less than 10 % phase speed error and the error can be neglected in this study. The M-transform was applied to each 256×256 km^2 area image, centered on the zenith, with 1×1 km^2 pixel size . The spectral components with 5-100 km horizontal wavelengths, 8-60 min periods, and 0-150 ms^{-1} phase speeds were extraced and regarded as representing GW intensities.

3. Comparison of GW activities between Syowa and Davis

Figure 1 shows total powers of the GWs, obtained using the PSD integrated in direction and phase speed from 0-150 ms^{-1}, over Davis and Syowa stations. The GW activity over both stations increased from March-April (fall) to May-August (winter) and decreased in September-October (spring). The total powers have similar seasonal variations, although the power at Davis was enhanced in late June to early July and late July to early August. On the other hand, the total power over Davis in spring was three times smaller than that over Syowa.

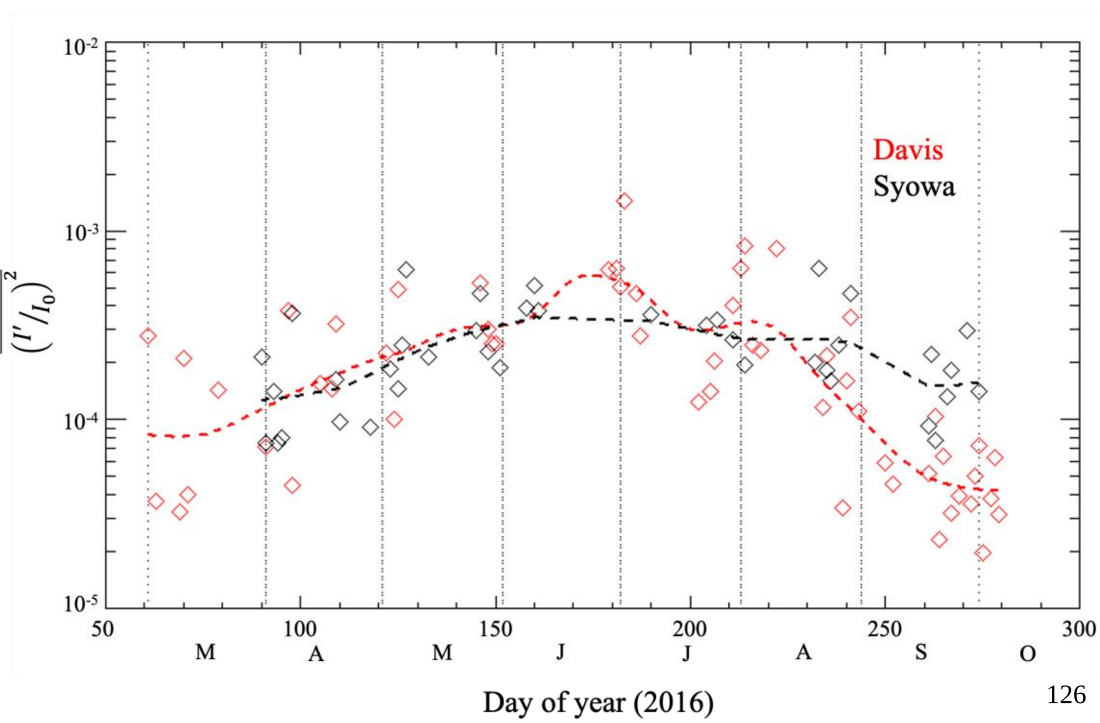


Figure 1. The red and black diamonds indicate total powers of GWs at Davis and Syowa, respectively. The red and black dashed lines indicate the smoothed total powers with logarithmic Gaussian weighting function ($\sigma = 15$ days, $width = 61$ days), respectively. It should be noted that the smoothed values have large uncertainty because of the sparse data.

Figure 2 follows Matsuda et al. (2017), but the total powers averaged in May-August and September over Davis and Syowa have been added. The powers over both stations were largest in May-August 2016, followed by April-May 2013 and September 2016. The total powers over Syowa and Davis in May-August, and over Syowa in September are larger than those over Halley and McMurdo in April-May, but the one over Davis in September is comparable.

The average power in May-August 2016 over Davis was larger than that over Syowa by 7×10^{-5}. This difference was larger than its one standard deviation (5×10^{-5}), which was calculated from one standard deviation of the two average total powers over Syowa and Davis in May-August. However, the powers in April-May 2013 were similar between Davis and Syowa, and Matsuda et al. (2017) concluded that the powers over Syowa and Davis were not different significantly. This contradiction could be attributed to different sampling numbers. The numbers of sampling days in May-August 2016 (23 days at Syowa and 29 days at Davis) are about three times larger than those in April-May 2013 (9 days at Syowa and 8 days at Davis). The confidence intervals in May-August 2016 should be about $\sqrt{3}$ times smaller than those in April-May 2013 in Figure 2. Also, the total power in September over Davis was almost three times (9×10^{-5}) smaller than that over Syowa, and this difference was larger than its standard deviation (3×10^{-5}). Thus, the GW total powers have significant differences between Syowa and Davis.

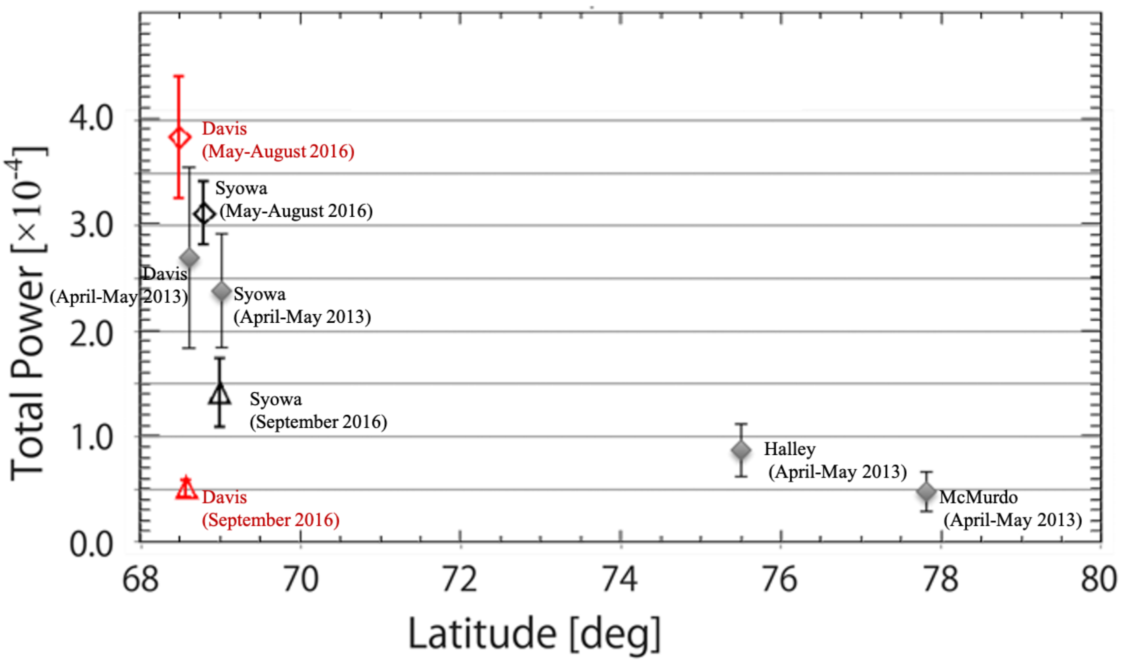


Figure 2. The red and black diamonds (triangles) indicate the mean total power in May to August (September) in Davis and Syowa, respectively. The gray diamonds indicate the mean total variance in April to May, 2013 [Matsuda et al., 2017]. These diamonds from right to left indicate the variances in Davis, Syowa, Halley, and McMurdo, respectively. The error bar indicates confidence interval of the average total power.

To further explore the different characteristics of the GWs between Davis and Syowa, the average phase velocity spectra for May-August and September are compared in Figure 3. The spectra for Syowa and Davis in Figures 3a, b, d, and e have the same directionality; that is, for phase speeds $\leq \sim 50\ \mathrm{ms}^{-1}$, the PSD between $180 - 360°$ clockwise from the north (westward) is larger than that in the same speed range between $0 - 180°$ clockwise (eastward). For phase speeds $\geq 80\ \mathrm{ms}^{-1}$, the result is reversed, and the PSD magnitude is for higher eastward waves. However, the spectral ratio $(PSD_{Davis}/PSD_{Syowa})$ in Figure 3c shows the PSD averaged in May-August around $45 - 180°$ clockwise from the north (northeast to southward), especially

$\geq \sim 50\ \mathrm{ms}^{-1}$, is larger over Davis than that over Syowa by one standard deviation in the PSD difference. This result implies that the larger PSD for azimuth between $45 - 180°$ clockwise from the north contributed to the higher winter mean total power at Davis. On the other hand, for Septmber (Figure 3f), the PSD at Davis in September is smaller in almost the whole phase velocity domain than that in Syowa by one standard deviation, implying the larger PSD in almost the whole domain contributed to the larger total power at Syowa during the Spring. These differences can be caused by the two possible mechanisms (1) wind filtering (2) wave generation. In the following section, we will discuss these mechanisms.

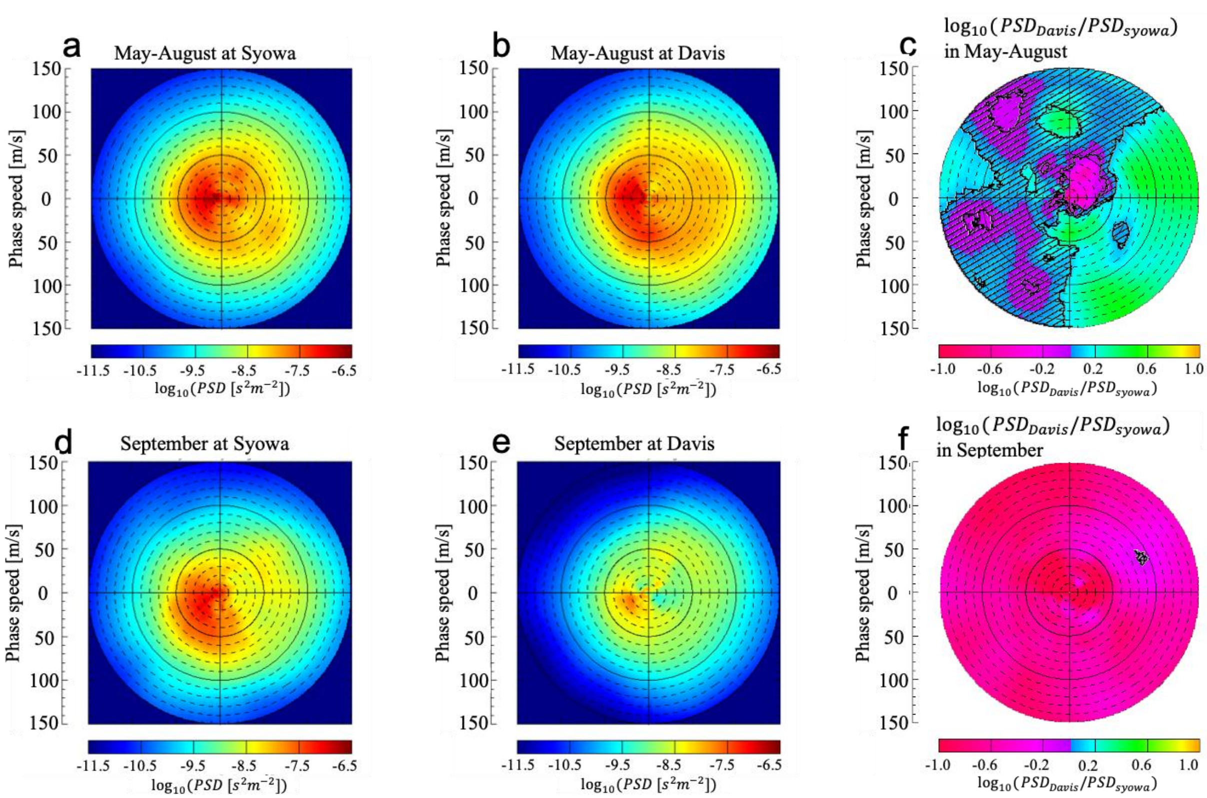
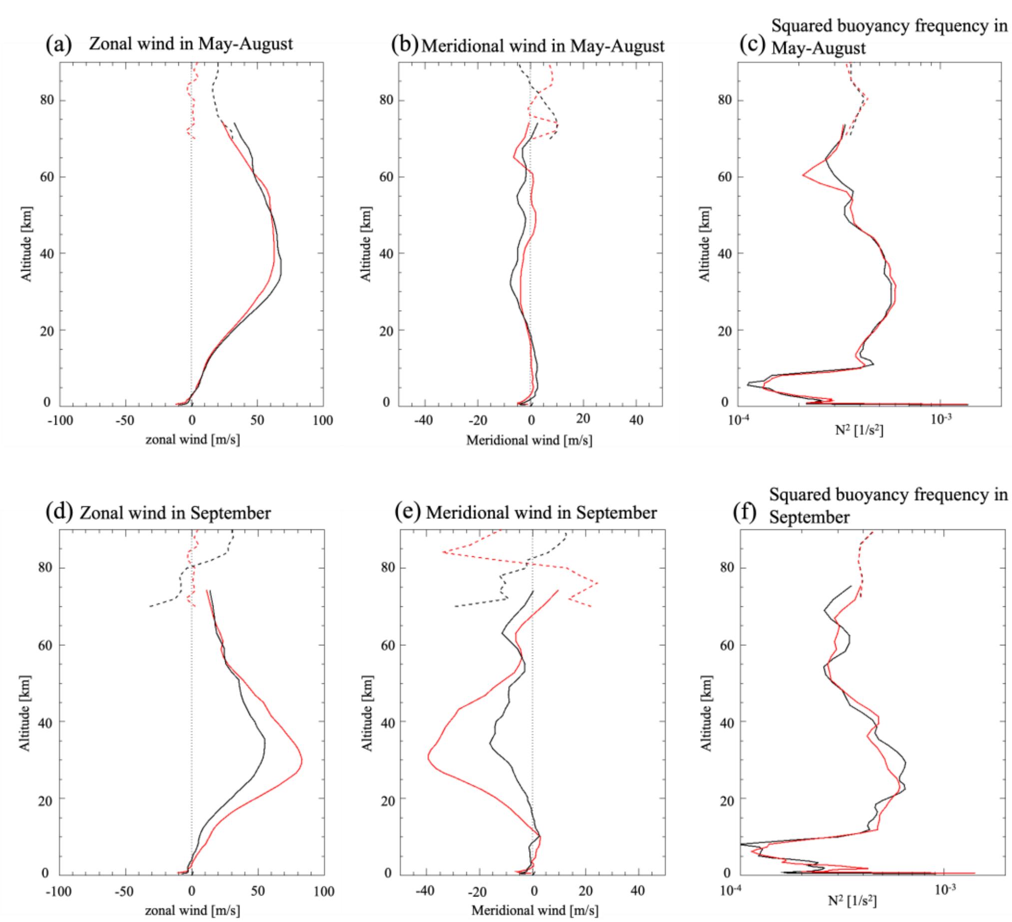


Figure 3. PSDs averaged in May-August and September at Syowa (a and b) and Davis (d and e), respectively. Phase speed lines are separated by 10 m s^{-1}. Panels c and f show the ratio between

the PSDs at Syowa and Davis in May-August and September, respectively. The hatch areas indicate that the ratio is within the confidence interval.

4. Comparison between GW spectra and transmission diagrams

To evaluate the wind filtering effect, the GW spectra were compared with transmission diagrams. The transmission diagram shows the possibility of vertical propagation, for a GW on a phase velocity spectrum, from the altitude of a wave emission up to the airglow layer altitude (Tomikawa, 2015). The wave source and airglow altitudes are assumed to be 10 km and 87 km, respectively. The diagrams were calculated from the winds and buoyancy frequencies averaged



in May-August and September at each station (Figure 4).

Figure 4. (a) Mean zonal and (b) meridional wind, and (c) squared buoyancy frequency profiles in May to August in Syowa (black) and Davis (red), respectively. The solid lines indicate re-analysis, MERRA-2. The dashed lines indicate observations (MF radars and MLS/Aura). (d), (e), and (f) are the same as (a), (b), and (c), respectivity, but in September.

The winds and buoyancy frequencies below 70 km altitude are from The Modern-Era Retrospective Analysis for Research and Applications, Version 2, i.e., MERRA-2, (Gelaro et al., 2017), and those above the top of MERRA-2 were from observations, MF radars (Tsutsumi et al., 2001) and Aura Microwave Limb Sounder (MLS) (Waters et al., 2006). The values between 70 km and the top of MERRA-2 are averages between MERRA-2 and the observations calculated with a linear weighting function. Tomikawa (2015) pointed out that the propagation possibility significantly depends on a GW period so that the spectra in Figure 3 were divided into 6-period bands in accordance with Matsuda et al. (2017).

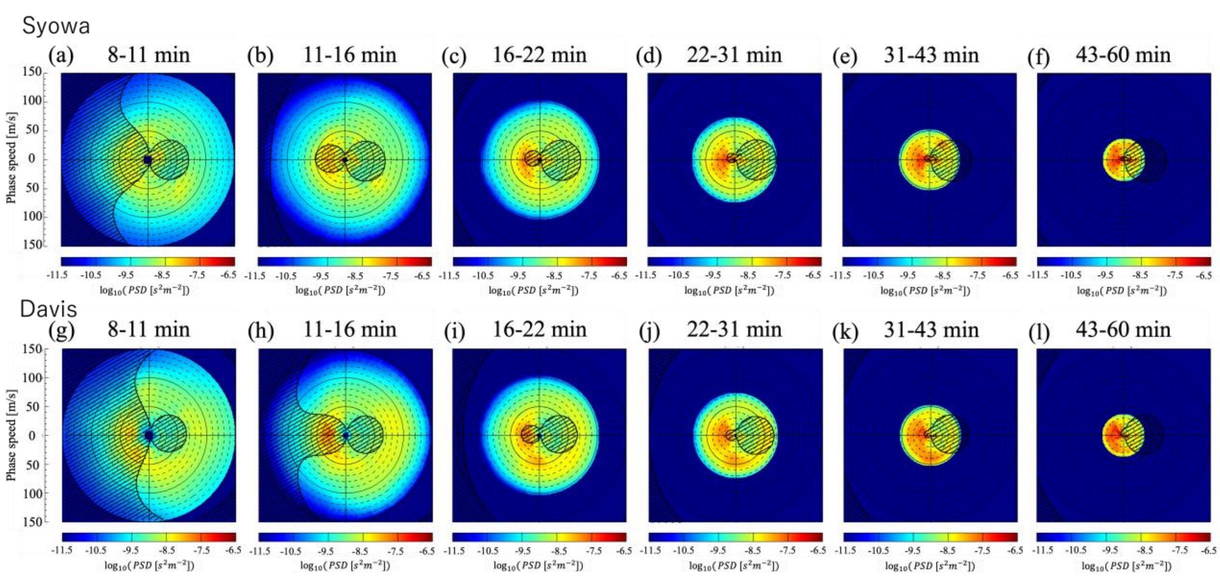


Figure 5. Comparison between the average spectra and transmission diagrams in May-August. The color plots indicate the mean spectra in Syowa and Davis in period ranges of 8-11, 11-16, 16-22, 22-31, 3-43, and 43-60 min, respectively. The shaded areas indicate prohibited areas for GWs to propagate into the mesopause region.

Figure 5 shows the comparison between the averaged spectra and transmission diagrams in May-August. The wind filtering effects at each period are almost the same at both stations; that is, the GWs with very short periods tended to be filtered in the westward phase velocity spectra, and the ones with the periods closer to an hour were filtered on the eastward phase velocity spectra less than ~70 ms^{-1}. This result shows that the wind filtering effect cannot explain the larger PSDs in eastward phase velocity in Davis. It should be noted that PSDs in the westward phase velocity spectra less than ~50 ms^{-1} and 8-16 min periods were enhanced at both stations even though they should be filtered. This result suggests that those GWs were not generated below the assumed 10 km source altitude (i.e., the troposphere). Bossert et al. (2017) and Heale et al. (2017) showed that the saturation of large-scale (long period) GWs can generate small-scale secondary GWs in the mesosphere. These saturated waves should also have higher harmonic Fourier components in frequency and horizontal wavenumber domains. The PSDs in period range 31-60 min and westward phase velocity less than ~50 ms^{-1} were also enhanced suggesting those GWs were saturated such that they might generate the small scale GWs and have higher harmonic Fourier components.

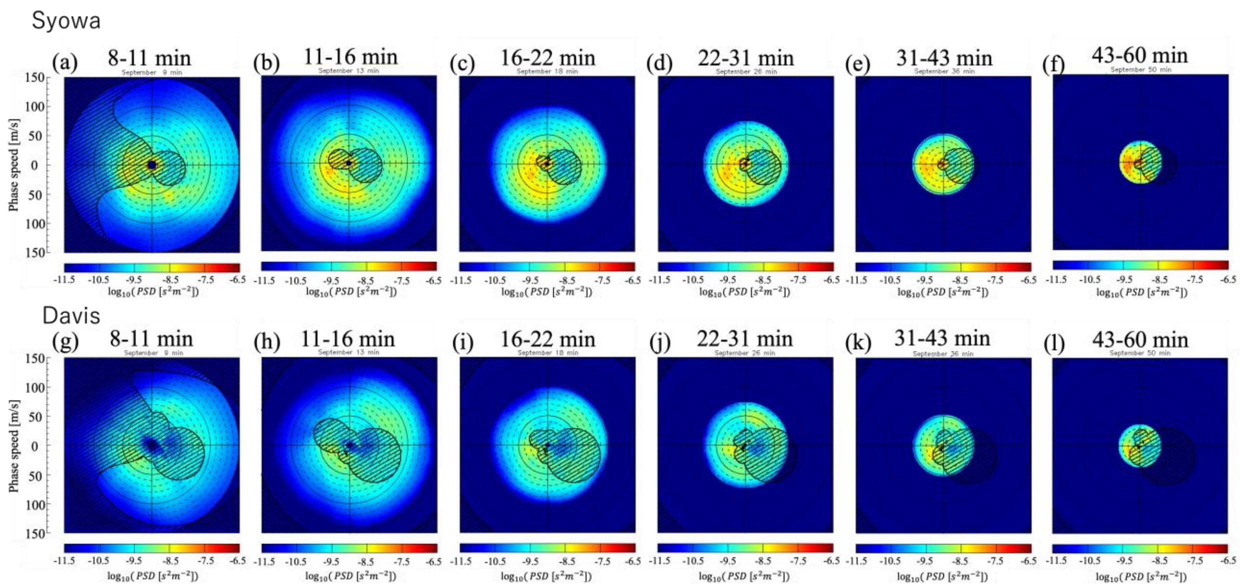


Figure 6 Comparison between the average spectra and transmission diagrams in September. The color plots indicate the mean spectra in Syowa and Davis in period ranges of 8-11, 11-16, 16-22, 22-31, 31-43, and 43-60 min, respectively. The shaded areas indicate prohibited areas for gravity waves.

Figure 6 is the same as Figure 5 but in September. The prohibited spectra in Davis were broadened into ~70-90 ms^{-1} southeastward more widely than those in Syowa because of the stronger southward wind. This difference between the filtering effects can explain the larger PSD in ~70-90 ms^{-1} southeastward phase velocity area in Syowa but cannot explain the PSD in the rest of the phase velocity area.

In summary, the filtering effect cannot explain the difference in May-August but can somewhat explain that in September. This result implies that other physical mechanisms should contribute more to the difference in the PSDs between Syowa and Davis. The possible mechanisms are (1) Horizontal propagation and (2) Source activity. The next section focuses on an event for GWs with eastward phase velocity at Davis to further explorer the differences in May-August.

5. Event for GWs with high southeastward phase velocity at Davis on August 29, 2016

5.1. Origin altitudes of GWs on August 29, 2016.

Intense GWs with east to southward ($90 - 180°$ clockwise from the north) phase velocity higher than $\sim 50\ \mathrm{ms}^{-1}$ were observed at Davis on August 29, 2016, but not observed at Syowa. Figure7 shows the one-night mean spectra on August 29 at Syowa and Davis. Powers at Syowa and Davis were intense in southwestward phase velocity spectra with ~20-30 ms^{-1}, while the powers at Davis were larger in southeastward phase velocity spectra between 80-120 ms^{-1} than those at Syowa. The features of the two PSDs are similar to those of the winter averages shown in Figure 3.

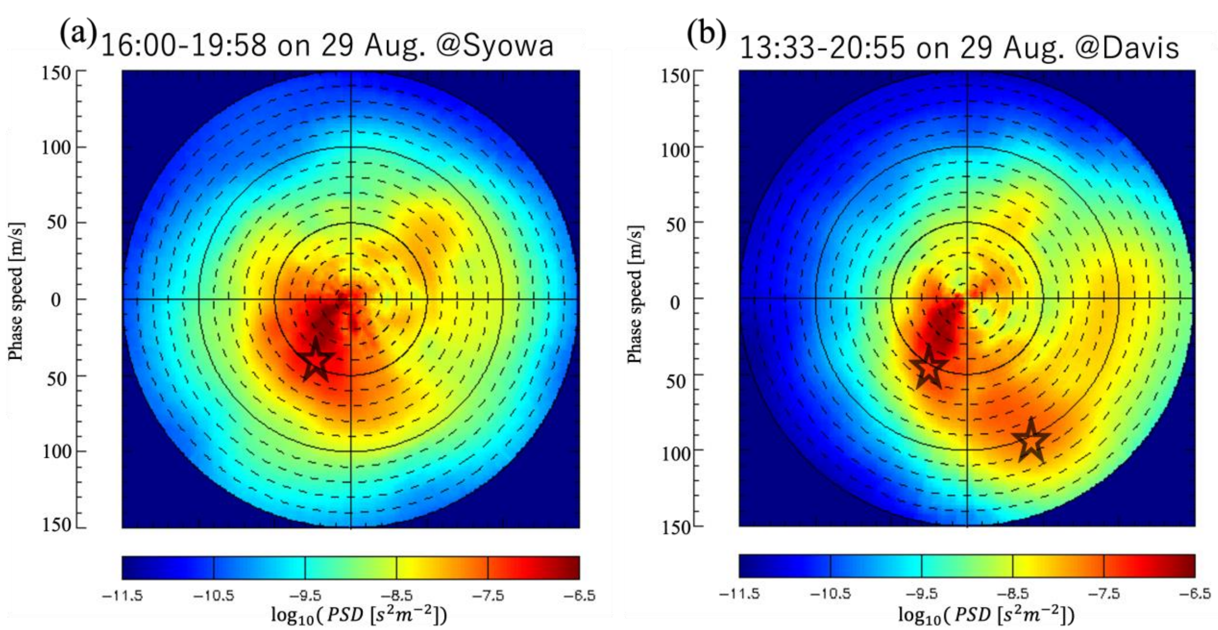


Figure 7. PSDs averaged on August 29, 2016, at Syowa (a) and Davis (b), respectively. The stars indicates initial phase velocities for raytracing method.

To explore where the GWs with ~100 ms^{-1} southeastward phase velocity came from, we applied a backward raytracing method (Kogure et al., 2018; Kogure et al., 2020). The initial λ_H, τ, C_h are selected as values corresponding to the peaks of the power in ~30 ms^{-1} southwest velocity spectra at Syowa and Davis, and ~100 ms^{-1} southeast velocity spectra at Davis. Table 3 shows the initial parameters. The background wind and temperature below 70 km altitude were obtained from MERRA-2. Above the top of MERRA-2 (~74 km altitude), the background wind was measured by MF radars, and the temperature was measured from Aura MLS. Between 70 km and the top of MERRA-2, a linear weighting function was used to average between MERRA-2 and the observations. It should be noted that horizontal wind shear effects on wavelength were neglected above the top of MERRA-2.

	No.	λ_H[km]	τ[min]	C_h[m/s]	Direction [degree]	Terminal height [km]
Syowa	S1	84	30	46	209	0.1
Davis	D1	28	9	52	207	9.2
	D2	88	14	103	156	44.5

Table 3. Initial raytracing parameters for hydroxyl layer gravity waves

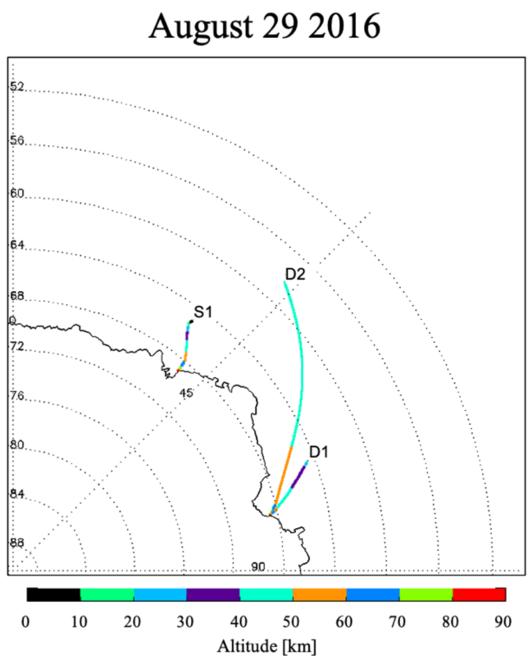


Figure 8. The backward raytracing method results for the GWs described in Table 3.

The backward raytracing results in Figure 8 show the three GWs propagated from the southern ocean. While the S1 and D1 (southwestward phase velocities) can propagate from 0.1 and 9.2 km altitudes, respectively, the D2 (southeastward phase velocity) back trace terminated at 44.5 km altitude because of a critical level. This result indicates that D2 could not have originated from below that level and must have been generated in the upper stratosphere. Rourke et al. (2017) observed such gravity waves with eastward phase velocity over Davis. This work also showed that those waves originated from the upper stratosphere by using the raytracing method, although it used empirical model background fields and did not determine the wave sources.

5.2. The possible GW sources in the stratosphere.

The possible sources for the D2 GW (southeastward phase velocity) in the stratosphere are (1) spontaneous adjustment (Plougonven and Zhang, 2014), and (2) secondary wave generation (Vadas & Becker, 2019; Kogure et al., 2020). The residual of the nonlinear balance equation (ΔNBE) from Plougonven and Zhang (2014) using MERRA-2 fields was not enhanced around the termination point so that the spontaneous adjustment mechanism can be ruled out (not shown). On the other hand, the Atmospheric Infrared Sounder (AIRS) aboard NASA's Aqua satellite captured two intense GWs near the termination point (~42E, ~58S) at ~9:40 UT in the 15 μm high channel, which, in the polar region, is most sensitive to GW temperature peturbations at ~38 km altitude (Figure 9 (a)). (Channel response data can be acquired from https://datapub.fz-juelich.de/slcs/airs/gravity_waves/ (Hoffman et al., 2013; 2017)). The 15 μm low channel of AIRS, which is most sensitive to GW temperature peturbation at ~23 km altitude, also captured two intense GWs in a similar area (not shown). These observed GWs appeared in the leeward of the geopotential height ridge at 200 hPa suggesting the waves were emitted from the tropospheric jet through spontaneous adjustment (Plougonven and Zhang, 2014).

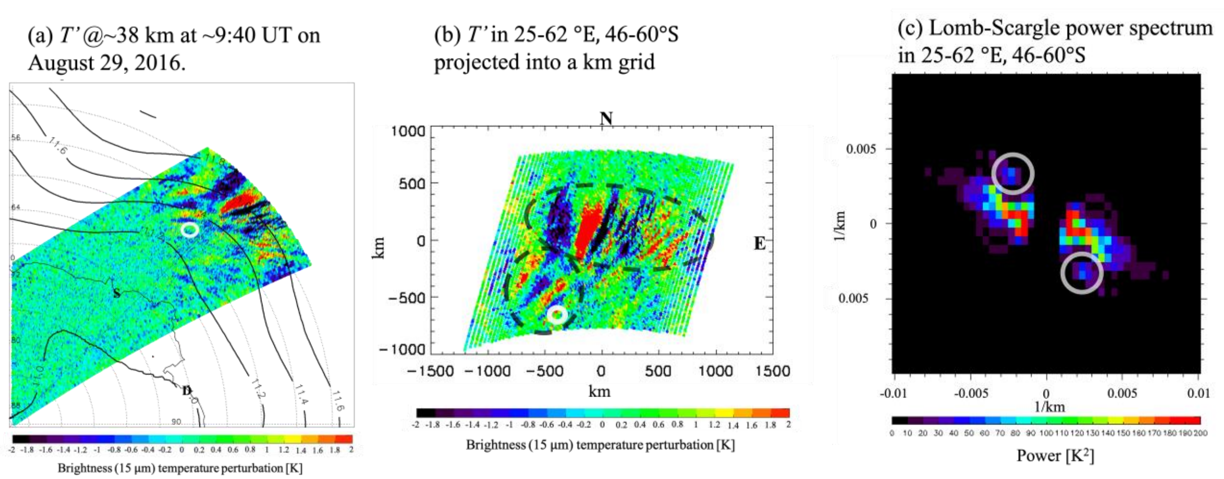


Figure 9. (a) Temperature perturbations observed by AIRS at ~38 km altitude at ~9:40 UT on August 29, 2016. The contour shows geopotential heights at 200 hPa. The white circlecorresponds to the termination of the D2 wave. (b) The Temperature perturbations in 25-62°E, 46-60°S projected onto a km grid. The white circle indicates the terminal point of the D2

GW (with southeastward phase velocity observed over Davis). (c) The Lomb-Scargle power spectrum of the temperature perturbations in 25-62°E, 46-60°S .

The AIRS data were projected onto a km grid in Figure 9 (b). There are two GWs apparent, with ~500 km and ~200 km horizontal wavelengths. The GW with ~200 km horizontal wavelength is situated near the termination of the D2 ray. Such intense GWs observed by AIRS could be saturated and have generated secondary gravity waves in the upper stratosphere and the mesosphere due to shear and/or convective instabilities (Bossert et al., 2015; Heale et al., 2020; Kogure et al., 2020). To investigate the saturation of the GWs, we calculate the local Richardson number, *Ri*, along a forward ray path. This analysis is similar to Kogure et al. (2020) with the exception that this study assumes wave action conservation; previously it was assumed that the growth rate of wave amplitude was the square root of density. To determine an initial horizontal vector for the forward raytracing, the AIRS data in Figure 9(b) is subject to 2D Lomb-Scargle spectral analysis. Figure 9(c) shows the Lomb-Scargle power spectrum where two local maxima in the spectrum can be seen. The local power maxima with 241 km horizontal wavelength and a direction of 146° or 214° clockwise from north correspond to the GW with ~200 km horizontal wavelength in Figure 9(b). The 2D Lomb-Scargle analysis has 180° ambiguity, but 146° is adopted here because the phase speed of a GW emitted from a jet is expected to match the phase speed of a baroclinic wave, i.e., eastward (Plougonven and Snyder, 2007). Murphy et al.(2014) also frequently observed GWs in radiosonde data with eastward phase velocity, which were potentially emitted from the jet, in the lower stratosphere over Davis but rarely observed westward GWs.

The measured amplitude of the GW with a 241 km horizontal wavelength is ~2 K in Figure 9(a) and (b), but this drastically underestimates the true amplitude (Hoffman et al., 2013; 2017). A response rate of the AIRS high 15 μm kernel is estimated using the AIRS vertical kernel investigation (https://datapub.fz-juelich.de/slcs/airs/gravity_waves/data/kernel.pdf), and the amplitude of the raytraced GW was compensated accordingly. However, the response rate and peak of the AIRS kernel varies in location and season, which causes ambiguities of the amplitude and the altitude used to launch the ray. Due to these ambiguities, the response rate and the initial altitude are varied by 3% and 2 km, respectively. The AIRS instrument cannot observe wave period, but the GWs emitted from jets tend to have a long ground-based period. Here, the period was trialled at 3, 5, and 10 hours.

The altitudes at which Ri < 0.25 for the trial GWs, which corresponding to potential start of saturation, are given in Table 4. Figure 10 shows the forward raytracing result for the GW with 3 hour ground-based period from 38 km altitude. This GW began to be saturated at almost the same point (diamond mark) as the D2 termination (white circle) on the horizontal surface although the saturation altitude is ~3 km higher than the D2 termination altitude. Precise agreement between the D2 termination height and the saturation height is not needed because the D2 wave could have been launched above the backward ray tracing termination height (but not below). GWs launched from similar horizontal positions but with the other parameters also propagated in the same direction (i.e., Davis ward) but the propagation paths and saturation altitudes varied. The saturation altitude of the GWs with the 3 hour ground-based period was closer to altitudes of the D2 path than those for 5 and 10 hours. It should be noted that Murphy et al. (2014) occasionally observed GWs with eastward and 22 ms^{-1} phase speed (from 241 km/3 hours) in the lower stratosphere over Davis that are consistent with this wave. Making the assumption that the southoeast propagating GW in the AIRS data had ~3 hour period, it can be asserted that the wave saturated at ~45-50 km. Since characteristic scales and directions of a secondary GW depend on a half scale and direction of a local body force, respectively (Vadas et al., 2003; Heale et al., 2020), the jet GW could cause a local body force with southeastward direction and ~44 km horizontal scale (a half of D2 GW horizontal wavelength). This local body force could create the GWs with ~100 ms^{-1} southeastward phase velocity. Otherwise, the GWs over Davis could be generated through nonliner fluid interactions (Andreassen et al., 1998; Bossert et al., 2015; Fritts et al., 1994, 1998; Heale 2017).Thus, the GWs with ~100 ms^{-1} southeastward phase velocity over Davis were very likely secondary GWs generated by the

Period (hour)	Initial altitude	Response rate of AIRS kernel (%)	Initial amplitude (K)	Bottom height Ri < 0.25 (km)	Initial vertical wavelength (km)
3	36	10 (7, 13)	20 (29, 15)	46 (39, 49)	16
3	38	10 (7, 13)	20 (29, 15)	48 (41, 59)	16
3	40	10 (7, 13)	20 (29, 15)	49 (46, 51)	16
5	36	14 (11, 17)	14 (18, 12)	50 (48, 51)	18
5	38	16 (13, 19)	13 (15, 11)	51 (51, 53)	19
5	40	16 (13, 19)	13 (15, 11)	53 (52, 56)	19
10	36	17 (14, 20)	12 (14,10)	52 (51,53)	20
10	38	19 (16, 22)	11 (13, 9)	55 (53, 57)	21
10	40	19 (16, 22)	11 (13, 9)	57 (55, 59)	21

intense GW activity above the southern ocean.

Table 4. Parameters for GW with a 241 km horizontal wavelength in the AIRS data on August 29, 2016. The initial amplitudes are 2 K (which is observed by AIRS) × the inverse of the response rate of the AIRS vertical kernel based on Figure 1 in Hoffmann et al. (2017). The values in parentheses are corresponding to the response rate −3 % and +3 %, respectively.

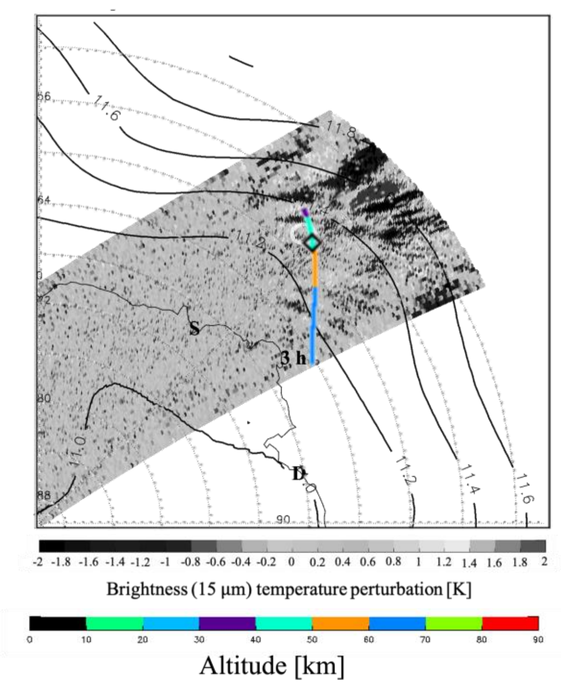


Figure 10. Raytracing result for the GWs with southeast horizontal wavevector superimposed in Figure 9(a). Its ground-based period, initial height, and initial temperature amplitude were assumed as 3 hour, 38 km, and 20 K, respectively. The diamamond indicates a bottom altitude with Ri < 0.25.

6. Conclusion

We compared the GW total powers and phase velocity spectra in the OH layer at two Antarctic stations, Davis and Syowa, from March to October 2016. The total powers showed maxima in winter and have similar seasonal variation at both stations. The powers at Davis were one standard deviation higher in winter and three times smaller in September than those at Syowa, respectively, although Matsuda et al. (2017) concluded that the powers at both stations were not significantly different at those times in 2013. This contradiction could be attributed to different sampling numbers, as our sampling numbers are about three times larger than those in Matsuda et al. (2017). The larger total power at Davis in winter was attributed to GWs with high eastward ($\geq \sim 50\ \mathrm{ms}^{-1}$) phase velocity. On the other hand, the larger total power at Syowa in September was attributed to GWs with omnidirectional phase velocity. To evaluate background meteological field impacts on the GWs at both stations, we compared the spectra and transmission diagrams and then concluded that filtering effects cannot explain the differences in May-August but somewhat explain them in September. To further explore the causes of the differences between Syowa and Davis in May-August, we focused on a GW event with eastward phase velocity at Davis. The raytracing method indicated that those GWs propagated from ~45 km altitude over the southern ocean (~$43°\mathrm{E}, \sim 58°\mathrm{S}$), where GWs emitted from the tropospheric jet were apparent. These tropospheric jet GWs were probably saturated in 45-50 km altitudes and caused a ~44 km horizontal scale (a half of the GW horizontal wavelength over Davis) local southeastward body force. Otherwise, the GWs over Davis could be generated through nonliner fluid interactions. Therefore, the GWs with eastward phase velocity observed in the mesosphere were secondary gravity waves. This result suggests that the larger PSD in eastward phase velocity at Davis in May-August was attributed to secondary GWs, that secondary GWs may contribute to GW phase velocity spectra in the mesosphere and that they cause non-negligible GW longitudinal variations. The larger total power in Syowa in September might be attributed to differences in a GW source activity or/and generation mechanisms between Syowa and Davis, but futuher study is needed to verify this.

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