Climatology and ENSO-related interannual variability of gravity waves in the Southern Hemisphere subtropical stratosphere revealed by high-resolution AIRS observations

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Abstract A new temperature retrieval from Atmospheric Infrared Sounder with a fine horizontal resolution of 13.5 km was used to examine gravity wave (GW) characteristics in the austral summer at an altitude of 39 km in the subtropical stratosphere over 8 years from 2003/2004 to 2010/2011. Using an S transform method, GW components were extracted, and GW variances, horizontal wave numbers, and their orientations were determined at each grid point and time. Both climatology and interannual variability of the GW variance were large in the subtropical South Pacific. About 70% of the interannual variation in the GW variance there was regressed to El Niño–Southern Oscillation (ENSO) index. The regression coefficient exhibits a geographical distribution similar to that of the precipitation. In contrast, the regression coefficient of the GW variance to the quasi-biennial oscillation of the equatorial lower stratosphere was not significant in the South Pacific. These results indicate that the interannual variability of GW variance in the South Pacific is controlled largely by the convective activity modulated by the ENSO. An interesting feature is that the GW variance is maximized slightly southward of the precipitation maximum. Possible mechanisms causing the latitudinal difference are (1) dense distribution of islands, which effectively radiate GWs with long vertical wavelengths, to the south of the precipitation maximum; (2) selective excitation of southward propagating GWs in the northward vertical wind shear in the troposphere; and (3) southward refraction of GWs in the latitudinal shear of background zonal wind in the stratosphere.

1. Introduction

It is well known that the meridional circulation in the middle atmosphere is driven by atmospheric waves, which maintain a temperature structure that is significantly different from that expected from radiative equilibrium [e.g., Holton, 1983]. Synoptic-scale waves are important to form the shallow branch of the Brewer-Dobson circulation (BDC), which is the meridional circulation in the stratosphere, both in the summer and winter hemispheres, while planetary waves are a main driver of the deep branch of the BDC in the winter hemisphere [e.g., Plumb, 2002]. In the mesosphere, gravity waves (GWs) are primary waves providing wave force to drive the meridional circulation [e.g., Andrews et al., 1987]. However, GWs play an important role to drive the BDC as well, particularly for the summer hemispheric part of the winter circulation where dominant westward mean winds prohibit upward propagation of planetary waves and for the shallow branches of the BDC through the westward forcing deposited in the weak wind layer above the middle latitude jet [Okamoto et al., 2011; Butchart, 2014; Stephan et al., 2016]. Studies using recently available high-resolution satellite observations and general circulation models suggest that the origins of GWs in the summer hemisphere are convection in the subtropical regions, particularly summer monsoon regions, while those in the winter hemisphere are topography and jet-front systems [Sato et al., 2009; Geller et al., 2013].

Satellites can detect GWs globally. However, the observable range of horizontal and vertical wavelengths by satellites are limited, and the limitations largely depend on the viewing geometry [Alexander and Barnet, 2007]. Limb-viewing satellite instruments such as the Limb Infrared Monitor of the Stratosphere (LIMS), the Cryogenic Infrared Spectrometers and Telescopes for the Atmosphere (CRISTA), and the High Resolution Dynamics Limb Sounder (HIRDLS) are able to detect GWs with relatively short vertical but long horizontal wavelengths. Nadir-viewing or sublimb-viewing satellite instruments and such as the Advanced Microwave Sounding Unit (AMSU) and the Atmospheric Infrared Sounder (AIRS) can observe GWs with relatively short horizontal but long vertical wavelengths. Such limitation in the detectable wavelength and/or frequency range is called the observational filter [Alexander, 1998].

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Several previous studies estimated absolute momentum flux associated with GWs using satellite data. Geller et al. [2013] conducted the first comparison among absolute momentum fluxes estimated using satellite, superpressure balloon, and radiosonde observations; those simulated by high-resolution general circulation models (GCMs); and those parameterized in climate models. They showed that the parameterized GW momentum flux is largely different from those estimated by satellite observations and those explicitly simulated in high resolution GCMs and indicated that particularly, nonorographic GWs are not sufficiently well expressed in the GW parameterizations and that a principal problem in models with parameterized gravity waves was in specifying the source distribution. Such differences may cause systematic model biases that are observed in the jet structure in middle atmosphere models. Thus, the characteristics of GWs originating from nonorographic sources need to be further investigated using high-resolution observations. According to the Geller et al. [2013] study, the GW momentum flux shows two peaks latitudinally: one is at subtropical latitudes in the summer hemisphere, and the other is at high latitudes in the winter hemisphere. The former is considered to be due to GWs originating from monsoon convection [e.g., Sato et al., 2009].

The horizontal distribution of GW variance at an altitude of 38 km with short horizontal and long vertical wavelengths has been investigated using high horizontal resolution data from the Microwave Limb Sounder (MLS) [Wu and Waters, 1996; McLandress et al., 2000; Jiang et al., 2004]. MLS detected fluctuations with vertical wavelengths longer than 10 km. Hence, part of the observed GW distributions, such as those around the polar night jet where strong winds refract waves to long vertical wavelengths, were attributable to the observational filter. However, the longitudinal distribution of GW variances in the summer subtropical regions may reflect the real nature of GWs, because large GW variance regions compare well with small outgoing longwave radiation and because the background wind that can modify GW vertical wavelengths is zonally almost uniform. These GWs are likely originating from the convection in the subtropical region. Enhancement of the GW activity over the summer subtropical monsoon regions was also observed by HIRDLS, which can detect GWs with short vertical wavelengths and long horizontal wavelengths [Wright and Gille, 2011]. GWs originating from convection are expected to have short horizontal wavelengths comparable to individual convection and/or convective systems. Thus, it is important to examine nadir-view satellite observation data as well. Currently, AIRS has the highest horizontal resolution, which is 13.5 km across and 18 km along the satellite orbit at nadir. Several previous studies using the AIRS radiance data examined the GW characteristics by applying a wavelet analysis method for a specific height level in the stratosphere [Alexander and Barnet, 2007; Alexander and Teitelbaum, 2007, 2011]. In this paper, we analyze high-resolution AIRS temperature data from a new retrieval [Hoffmann and Alexander, 2009] focusing on GWs in the tropical region.

There are few studies on the interannual and intraseasonal variability of the GW activity except for the relation to the equatorial quasi-biennial oscillation (QBO), using radiosondes [Sato et al., 1994; Sato and Dunkerton, 1997], satellites [Ern and Preusse, 2009; Gong et al., 2012; John and Kumar, 2012; Zhang et al., 2012], and high-resolution numerical models [Kawatani et al., 2010; Evan et al., 2012]. However, the interannual and intraseasonal variability of GW variance can be affected by other dominant phenomena in the tropical and subtropical regions such as the El Niño–Southern Oscillation (ENSO) and Madden-Julian Oscillation (MJO), having characteristic horizontal structure in precipitation and horizontal winds. As AIRS started its observation in 2002, the observation duration is sufficient to examine such interannual and intraseasonal variations. In the present study, AIRS data over 9 years from 2003 to 2011 were used to examine the climatology of GWs in the summer subtropical region and the interannual variability of GWs in terms of ENSO. The intraseasonal variability in terms of MJO is investigated in a companion paper [Tsuchiya et al., 2016].

In section 2, details of the AIRS observation data and the method of analysis are described. The climatology of GWs in the tropical and subtropical regions in summer is presented in section 3. In section 4, the interannual variability of GWs and its relation to ENSO are shown focusing on the SH subtropical region. In section 5, modulation of GWs by the QBO as another factor causing the interannual variability and possible mechanism of the latitudinal difference between GW and convection maxima that is elucidated in the present study are also examined and discussed. Summary and concluding remarks are given in section 6.

2. Data Description and Method of Analysis
2.1. Data Description

AIRS [Aumann et al., 2003] is one of the six instruments on board the Aqua satellite [Parkinson, 2003]. Aqua was launched on 4 May 2002. It has a Sun-synchronous nearly polar orbit with 98° inclination at 705 km altitude. Aqua crosses the equator at 01:30 (descending orbit) and 13:30 (ascending orbit) local time.
AIRS measures the thermal emissions from atmospheric constituents in the nadir and sublimal directions. The scan angle across the measurement track is ±49.5°, corresponding to a distance on the ground of 1765 km [Hoffmann et al., 2013]. Each across-track scan consists of 90 footprints. The extent of a granule, which consists of 135 scans, is about 2400 km along the track. Thus, the highest horizontal resolution is 13.5 km across and 18 km along the satellite orbit at nadir. Kernel functions of CO₂ channels with radiances of 15 and 4.3 μm typically have a peak in the stratosphere and a depth of about 12 km [Alexander and Barnet, 2007; Hoffmann and Alexander, 2009].

The AIRS operational level 2 temperature product has a horizontal resolution coarser than the original radiances data by a factor of 3 × 3 (namely, 39 km × 54 km), which corresponds to the horizontal resolution of Advanced Microwave Sounding Unit (AMSU) on board Aqua. However, the level 2 temperature data may be not sufficient to detect such short horizontal wavelengths as convectively generated GWs have. To overcome this shortage, Hoffmann and Alexander [2009] developed a new retrieval of atmospheric temperature, which provides data with a native high resolution of the AIRS radiance measurements. They used 23 channels of 4.3 μm radiance and 12 channels of 15 μm radiance for retrievals at the nighttime when the solar zenith angle is larger than 96°, while only 12 channels of 15 μm radiance are used for retrievals at the daytime, because the assumption of local thermodynamic equilibrium for 4.3 μm radiance is not valid. Thus, the noise level of the retrievals at nighttime is lower than that at daytime. For this reason, the present study used the new retrieval of temperature at nighttime only. In addition, the noise of the AIRS high-resolution retrieval of temperature is minimized for an altitude range 25–45 km. Following previous studies, this study focused on a specific height level of 39 km approximately corresponding to 3 hPa. Vertical resolution is about 9 km at that level. The analyzed time period is 9 years from 2003 to 2011. A validation of the new AIRS retrieval is presented by Meyer and Hoffmann [2014].

A reanalysis data, NASA’s Modern-Era Retrospective Analysis for Research and Applications (MERRA) [Rienecker et al., 2011], is used for the analysis of the background field of GWs. MERRA data are generated with the Goddard Earth Observing System atmospheric model and data assimilation system where AIRS data are also assimilated. Although the original MERRA data are available three hourly, daily mean temperature and horizontal wind values are used for the analysis. The ocean fraction at each grid point in the numerical model used for MERRA is also used to see the surface condition.

In addition, we used daily 1 × 1° gridded precipitation data from the Global Precipitation Climatology Project version 1.2 [Huffman et al., 2001] as an index of convection. The NINO.3 index, which is defined as sea surface temperature (SST) anomalies averaged over the region which is 5°S to 5°N and 150°W to 90°W, from the Japan Meteorological Agency (http://www.data.jma.go.jp/gmd/cpd/data/elnino/nino3irm.html) is used as an ENSO time series. We also used daily 0.25 × 0.25° gridded SST values from the Optimum Interpolation Sea Surface Temperature data [Reynolds et al., 2007] to examine a horizontal distribution of SST.

2.2. Method of Analysis

In this section, the method used in the present study to analyze the horizontal propagation characteristics of GWs is described. S transform [Stockwell et al., 1996] is a one-dimensional wavelet-type analysis and is suitable for the estimation of local characteristics of GWs. Several previous studies [Alexander and Barnet, 2007; Alexander et al., 2008; Alexander et al., 2009; Alexander and Grimsdell, 2013] applied the S transform to the satellite data of AIRS and HIRDLS to detect localized GW packets.

First, a large-scale field was obtained as follows: The data scans of temperature across the orbit were regressed to a second-order polynomial function and the regressed data scans were further smoothed along the orbit by the 31-point (i.e., 558 km) running mean. The deviation of the original data from the large-scale field was designated as the GW components. Original sampling interval of the data scans across the orbit varies from 13.5 km at nadir to 39.6 km at its edge. So as to make the analysis easier, the GW components are interpolated at the same interval of 13.5 km by a spline fit across the track. As a consequence, the number of data series across the track becomes 130 (=1755 km).

GW parameters such as two-dimensional horizontal wave numbers were estimated in two ways. First, the estimation was performed using an S transform method for cross-track data series and cross spectra for adjacent cross-track data by a method developed by Alexander and Barnet [2007]. With this method, cross-track wave numbers are more accurately estimated than along-track wave numbers. Second, the same method...
was applied to along-track data series which usually provides more accurate along-track wave numbers than cross-track wave numbers. We took better estimates between the two depending on the direction of the two-dimensional wave number vector for further analysis. An example of satellite data are shown in Figures 1a–1c.

The detailed method is as follows.

Figure 1. (a) An example of AIRS temperature observation 20 January 2007. (b) A cross-track data series (black) with an adjacent cross-track data series (blue). The blue curve is shifted upward by 3 K. (c) An along-track data series (black) with an adjacent along-track data series. A blue curve is shifted upward by 3 K. (d) The angle $\alpha$ between the directions of the data series across and along the satellite orbit at the daytime (ascending orbit) shown by a red curve and at the nighttime (descending orbit) by the blue curve as a function of the latitude, calculated using data on 24 December 2003. (e) A schematic illustration of directions and angles describing a satellite track and a gravity wave. See text for details.
The S transform spectra are calculated using the Fourier transform and hence could be affected by the difference between the first and last ends of the data series. So as to avoid this edge effect and to obtain statistically stable S transform spectra, a window function, which has a cosine shape at both ends for one tenth of the total length (=1755 km), was multiplied to the GW data series across the track. The S transform spectra were calculated at respective GW data series across the track. Cross spectra for respective two adjacent data series were obtained using the S transform spectra. A wave number vector \((k, \ell)\) of the GWs is estimated at each grid point. Here \(k\) and \(\ell\) are the wave number components in the direction of the cross-track data series and perpendicular to that direction, respectively. First, the wave number \(k\) is determined as the wave number at which the magnitude of the cross-spectral density is maximized in the meaningful wave number range. Here the meaningful wave number range was estimated at \(2 \pi/(70 \text{ km})\) to \(2 \pi/(700 \text{ km})\) based on an analysis of noise spectra (see Appendix A for details). The wave number component across the data series (i.e., along the track), \(\ell\)', is estimated from the phase shift of the cross spectra at \(k\). Note that \(\ell\)' is different from \(\ell\), because the two data series which are, respectively, across and along the track are not right angled as seen in Figure 1a which shows the angle between the two data series \(a\) as a function of the latitude. The relation among \(k\), \(\ell\), \(\ell\)' and \(a\) is illustrated in Figure 1b. The wave number component \(\ell\) is estimated using \(k\), \(\ell\)' and \(a\) as

\[
\ell = \frac{\ell' - kc\sin a}{\sin a}.
\]

The wave number vectors were estimated for a central cross-track data series with a width of 1215 km, because the above mentioned window function was applied to each one tenth end of the cross-track data series. Similar analysis can be made by applying the S transform to the data series “along” the track with a length of 196 (i.e., 3528 km), and the estimation of wave number vector was made for the central 135 grid points (i.e., 2430 km). Note that the analysis using cross-track data series (along-track data series) provides better estimates for waves with \(|\ell| < |k|\) (\(|\ell| > |k|\)). In this way, we obtained a pair of horizontal wave number vectors for each grid point for the same GW, i.e., one from the two adjacent data series across the track and the other from that along the track. Through tests with idealized wave patterns, a better estimate of \((k, \ell)\) vector was selected with a criteria based on the angle of the horizontal wave number vector \((\phi)\), where a positive (negative) \(\phi\) value means an angle counterclockwise (clockwise) from the cross-track direction (see Figure 1e). We selected the estimate from the along-track data series when \(-45^\circ < \phi < 45^\circ\) and that from the cross-track one when \(-90^\circ \leq \phi \leq -45^\circ\) or \(45^\circ \leq \phi \leq 90^\circ\). The direction of the horizontal wave number vector in the Cartesian coordinate, namely, the angle with counterclockwise rotation from the eastward direction, \(\phi_n\), is then estimated using the cross-track direction in the Cartesian coordinate, \(\phi_\theta\), as \(\phi_n = \phi + \phi_\theta\). Note that there is an ambiguity of \(180^\circ\) in \(\phi_n\). According to a GW-resolving general circulation model study by Sato et al. [2009], dominant GWs tend to have negative (positive) vertical flux of zonal momentum in the eastward (westward) background wind. This means that the zonal component of horizontal wave number vector has opposite sign to the background zonal wind. Thus, based on this fact, we determined the direction of horizontal wave number using the zonal wind from MERRA at each grid point. The horizontal wavelength is calculated as \(\lambda_n = 2\pi/(k^2 + \ell^2)^{1/2}\).

GW amplitude squared was estimated as the absolute value of the cross spectra at \(k\) with a unit of K^2, which is hereafter referred to as the GW variance. Note that this GW variance is equal to twice as much conventional variance. In addition, it was seen that the data were quite noisy and temperature perturbation signals were quite weak in the regions with weak background winds. This is probably because in such weak background winds, vertical wavelengths of GWs are not sufficiently long to be detected by AIRS. Thus, we simply omitted the data in regions where the background wind is slower than 10 m s^{-1} for the analysis. This threshold for the background wind is somewhat arbitrary; however, it was confirmed that the results are not sensitive to slight changes of the threshold.

### 3. Climatology of GWs in the Summer Subtropics

Figures 2a and 2b show maps of the climatology of GW variance in the summer subtropics for the Southern Hemisphere (SH) averaged over December to February (DJF) and for the Northern Hemisphere (NH) over June to August (JJA), respectively. The GW variances are large over continents such as South Africa, Australia, South America, North Africa, South and Southeast Asia, North America, and over the western to central South Pacific. This feature is consistent with MLS observations [McLandress et al., 2000; Jiang et al., 2004].
Figure 2c (2d) shows maps of the standard deviation of seasonal mean GW variance showing interannual variability in the SH (NH). The numbers of years to obtain the interannual variability is eight and nine for Figures 2c and 2d, respectively.

An interesting feature is that the interannual variability of the GW variance in the summer subtropics is larger in the SH than in the NH, although the climatological GW variances are comparable. The standard deviation of the DJF-mean GW variance in the Australian monsoon region amounts to about 20% of the climatology (Figure 2c), while in JJA Asian monsoon region it is about 12% (Figure 2d). Thus, in the following, we mainly analyze the climatology and interannual variability of DJF-mean GW characteristics in SH subtropics.

Figure 3 shows DJF-mean climatology of (a) precipitation and zonal winds at 100 hPa, (b) GW variance and zonal winds at 3 hPa, (c) GW horizontal wavelength averaged with a weight of the GW variance, and (d) GW horizontal wave number direction \( \phi_h \) averaged with a weight of the GW variance in the SH tropical and subtropical region. The South Pacific Convergence Zone (SPCZ) is defined as the latitudes of the precipitation maxima for respective bins from 150°E to 140°W and denoted by a red curve in all maps of Figure 3.

As expected, the GW variance maxima are observed in strong precipitation regions such as in South Africa, Australia, and South America continents and in the South Pacific. This indicates that the GWs at 39 km observed by AIRS are originating from strong convection in the troposphere. It is interesting that the GW variance maxima are located southward of the precipitation maxima by a few degrees at respective longitudes. Similar differences in the locations of precipitation and GW variance maxima are also seen in South Africa and South America.

The zonal wind at 3 hPa is mainly zonally uniform in Figure 3b, although it is slightly stronger southward of the SPCZ. This fact indicates that the characteristic longitudinal distribution of GW variance observed in Figure 3b (or Figure 2a) is not solely due to the observational filter of AIRS but is reflecting true differences in GW properties. The mean GW horizontal wavelengths are long (>200 km) over southeastern Africa, Australia, and southwestern America where the GW activity is high, while those in the other regions are ~150 km (Figure 3c). The mean horizontal wave number direction is eastward or slightly southward in most regions (Figure 3d). The direction tends to more southward to the west of the precipitation. It was confirmed by a nonparametric method [Fisher, 1993] that the observed directional departures of several degrees from the zonal direction are statistically significant at 95% confidence intervals.
4. ENSO-Related Interannual Variability of GW Variance

As a possible cause of the GW interannual variability observed by AIRS in the austral summer season, we examined the relation with ENSO. DJF-mean GW variance and precipitation were made in respective years and binned at each 2.5° × 10° latitude-longitude box area. To see the interannual variability of ENSO, the NINO.3 index is used (Figure 4) [e.g., Trenberth, 1997]. Note that a five-monthly mean was applied to the NINO.3 index by its definition. Values of the NINO.3 index in January of 2004 to 2011 were used as a reference time series for our analysis. It is seen that DJF periods of 2003/2004, 2004/2005, 2006/2007, and 2009/2010 (2005/2006, 2007/2008, 2008/2009, and 2010/2011) are in El Niño (La Niña) or similar conditions, which are hereafter referred to as the El Niño (La Niña) years.

Figures 5a (5c) and 5b (5d) show composite maps for precipitation (GW variance) in the El Niño and La Niña years, respectively. Both the precipitation and GW variance distributions are largely different over the subtropical South Pacific between the El Niño and La Niña years. They are large northward (southward) of climatological SPCZ shown by a red curve in the El Niño (La Niña) years. The GW variance is stronger in Maritime Continent, North Australia, and Indochina in the La Niña years than in the El Niño years. In contrast, the difference is not significant in other regions such as African and American continents and western Indian Ocean. It is also worth noting that the GW variance is large over the continents and not necessarily very large along the SPCZ where significant precipitation is observed, as consistent with the fact recently indicated by Geller et al. [2015]. The contrast in the precipitation and GW variance distribution between the El Niño and La Niña years is more clearly seen in the results from a regression analysis in the following.

Figure 6a (6b) shows correlation coefficients of the DJF-mean precipitation (GW variance) with the NINO.3 index. Figure 6c (6d) represents regression coefficients of the DJF-mean precipitation (GW variance) to the NINO.3 index in the region where the magnitude of correlation coefficients with the NINO.3 index is larger than 0.62 corresponding to a confidence level of 90%. Positive correlation and regression coefficient values indicate an increase in the precipitation and GW variance in the El Niño years.

Magnitudes of the correlation coefficients are greater than 0.83 over the subtropical South Pacific for both precipitation and GW variance. The regressed component accounts for about 70% of the GW interannual variability at most. The precipitation has positive regression and correlation coefficients eastward of SPCZ and in the central equatorial South Pacific, and negative regression and correlation coefficients westward of SPCZ and in the eastern South Indian Ocean. The coefficients for the GW variance exhibit similar
distributions. An exception is seen at the Maritime Continent where the correlation and regression coefficients are significantly negative for the precipitation, while they are small for the GW variance.

So as to examine the cause of the difference in precipitation and hence likely the GW variance, we computed correlation coefficients with and regression coefficients to the NINO.3 index of the DJF-mean sea surface temperature (SST) and horizontal wind divergence at 850 hPa (Figure 7). Only values with a confidence level greater than 90% are plotted. The SST has positive regression and correlation coefficients eastward of SPCZ and in the central equatorial South Pacific. However, significant values are not seen westward of SPCZ. The correlation and regression coefficients of the horizontal wind divergence are negative eastward of SPCZ and positive westward of SPCZ and in the central equatorial South Pacific. These characteristics are consistent with the distributions observed in the precipitation shown in Figure 6.

The characteristic modulation of the GW variance by ENSO observed in Figures 6b and 6d is likely due to modulation of GW sources (i.e., convection). However, we need to scrutinize carefully the possibility of virtual modulation by the observational filter. For example, GWs, which have zonal wave number vectors and roughly zero ground-based phase speeds like mountain waves, tend to have longer (shorter) vertical wave-lengths in stronger (weaker) zonal winds which are more detectable (undetectable) by AIRS. As seen in Figure 3d, horizontal wave number vectors are oriented mainly zonally. Thus, even if the horizontal phase speed spectra of GWs propagating into the middle stratosphere are the same, the GW variance observed by AIRS may exhibit virtual interannual variability by interannually varying background zonal winds.

To examine this possibility, further analysis is made for three regions where characteristic interannual variability of the GW variance is observed, synchronized with ENSO: (A) the equatorial western South Pacific, (B) the equatorial eastern South Pacific, and (C) the Maritime Continent.

**Figure 4.** Time series of SST anomaly from the 30 year climatology in the NINO.3 region (5°S to 5°N, 150°W to 90°W). Black dots show NINO.3 data used for the regression analysis. See text for details.

**Figure 5.** Same as Figures 3a and 3b but for composite maps of (a, b) precipitation and (c, d) GW variance at 39 km for (Figures 5a and 5c) El Niño and (Figures 5b and 5d) La Niña years.
Paciﬁc region (0°S to 10°S, 150°E to 150°W), (B) the subtropical region (10°S to 30°S, 150°E to 110°W) to the east of the SPCZ, and (C) the subtropical region to the west of the SPCZ that are denoted in Figure 6c. We calculated DJF-mean background wind zonal wind at 3 hPa for respective years and averaged over respective A, B, and C regions. Results are shown in Figure 8 together with the time series of DJF-mean GW variance and precipitation averaged over respective regions. Regional dependence is clear for the GW variance and precipitation: both the GW variance and precipitation values are large (small) in the A and C regions and small (large) in the B region in the El Niño years except 2003/2004 (the La Niña years). In contrast, the mean zonal wind exhibits similar variation for all regions and does not seem to be modulated much by ENSO. Thus, we can exclude the possibility of the observational ﬁlter alone causing the interannual variability observed in the GW variance. It is therefore concluded that the interannual variability of stratospheric GWs in the SH summer subtropical region is largely due to the modulation of tropical convective GW sources by ENSO.

The regression to the NINO.3 index is also performed for the mean horizontal wavelengths and the horizontal wave number direction (Figure 9). The correlation and regression coefﬁcients for the horizontal wavelengths exhibit similar patterns to those for the GW variance (Figures 6b and 6d): They are largely positive in the A and C regions and negative in the B region. The rate of change in the horizontal wavelength is about 20 km per 1 K NINO.3 SST at most. In contrast, the regressed pattern of horizontal wave number direction shows different

Figure 6. Maps of correlation coefﬁcients of NINO.3 time series with (a) DJF-mean precipitation and (b) GW variance. Red (blue) contours show positive (negative) correlation. Thin contours show ±0.62 corresponding to the 90% signiﬁcant level, and thick contours show ±0.83 corresponding to the 99% signiﬁcant level. Regression coefﬁcients for (c) DJF-mean precipitation and (d) GW variance are shown by colors only in regions with correlation coefﬁcient magnitudes larger than 0.62. Thick black lines show the regions of (A) the equatorial central South Paciﬁc (from 150°E to 150°W, from 0°S to 10°S) and (B and C) the regions, respectively, to the east and west of SPCZ (from 150°E to 110°W, from 10°S to 30°S). The SPCZ is denoted by a thick red curve.

Pacific region (0°S to 10°S, 150°E to 150°W), (B) the subtropical region (10°S to 30°S, 150°E to 110°W) to the east of the SPCZ, and (C) the subtropical region to the west of the SPCZ that are denoted in Figure 6c. We calculated DJF-mean background wind zonal wind at 3 hPa for respective years and averaged over respective A, B, and C regions. Results are shown in Figure 8 together with the time series of DJF-mean GW variance and precipitation averaged over respective regions. Regional dependence is clear for the GW variance and precipitation: both the GW variance and precipitation values are large (small) in the A and C regions and small (large) in the B region in the El Niño years except 2003/2004 (the La Niña years). In contrast, the mean zonal wind exhibits similar variation for all regions and does not seem to be modulated much by ENSO. Thus, we can exclude the possibility of the observational ﬁlter alone causing the interannual variability observed in the GW variance. It is therefore concluded that the interannual variability of stratospheric GWs in the SH summer subtropical region is largely due to the modulation of tropical convective GW sources by ENSO.

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Figure 7. Same as Figure 6 but for (a, b) SST and (c, d) horizontal wind divergence. The SPCZ is denoted by a thick orange (red) curve for Figures 7a–7d.
features: Significant negative correlation and regression coefficients are observed along the SPCZ. The rate of change in the direction is about 2° per 1 K NINO.3 SST. This means that the GWs over the SPCZ propagate slightly more southward relative to the mean wind in the El Niño phase than in the La Niña phase.

As described in section 3, the DJF-mean GW variance climatology is maximized slightly to the south of the precipitation maximum (Figures 3a and 3b). This feature is further examined by making a composite separately for the El Niño years and for the La Niña years. Figure 10a shows composite profiles of the GW variance (black) and precipitation (blue) as a function of the latitude relative to the climatological SPCZ latitude that are averaged over longitudes from 150°E to 150°W for all years (i.e., climatology), while Figures 10b and 10c represent the same composite profiles but for the El Niño and La Niña years, respectively. A profile of a mean ocean fraction in each grid box for the same longitude region is also plotted by a green curve in Figure 10, which will be referred to in the discussion in section 5.2.

The precipitation maximum shifts northward (southward) in the El Niño (La Niña) years compared with the climatology. However, it is commonly seen for both phases that the GW variance is maximized southward of the precipitation maximum. It is interesting that the latitudinal difference between the precipitation and GW variance maxima is larger in the El Niño years than in the La Niña years. This feature is at least qualitatively consistent with the fact that the mean horizontal wave number vector direction is more southward in the El Niño years (Figures 9b and 9d).

5. Discussions
5.1. Possibility of Interannual Variability Modulation by the QBO

As described in section 1, the interannual variability of stratospheric GWs in the tropical region has been discussed in terms of the QBO in previous studies. It seems, however, that the QBO does not largely modulate the interannual variability of the GWs observed by AIRS over this subtropical South Pacific region, as shown below.
As a QBO index, we used a time series of DJF-mean zonal-mean zonal wind at 10 hPa at the equator from MERRA. Figures 11a and 11b respectively show correlation and regression coefficients between the GW variance at respective locations and the QBO index. Regression coefficients of the GW variance time series to the QBO index are only shown in regions where the correlation coefficient magnitudes are greater than 0.62 corresponding to a confidence level of 90%. Significant modulation by the QBO is observed in longitudes from 120°W eastward to 60°E at latitudes lower than 10°S. The negatively large regression coefficients in this region mean that the GW variances are larger in the westward phase of the QBO at 10 hPa than the eastward phase. In contrast, significant modulation by the QBO is not observed in the west and central South Pacific region even near the equator which is the focus in the present study. Similar results were obtained for the correlation and regression analysis performed using zonal-mean zonal wind at the equator at 30, 40, 50, and 70 hPa (not shown). Thus, this result also strongly suggests that the interannual variability of stratospheric GWs over the western and central parts of the subtropical South Pacific in austral summer is largely affected by ENSO.

Figure 9. The same as Figure 6 but for (a and c) horizontal wavelengths averaged with a weight of the GW variance and (b and d) direction of horizontal wave number vector averaged with a weight of the GW variance.

Figure 10. Composites of the GW variance (black), precipitation (blue), and mean ocean fraction (green) at longitudes from 150°E to 150°W as a function of the latitude relative to the climatological SPCZ. (a) Composites are made for the climatology. (b) Composites are made for the seasonal mean in the El Niño years such as 2003/2004, 2004/2005, 2006/2007, and 2009/2010. (c) The same as Figure 10b but for the La Niña years such as 2005/2006, 2007/2008, 2008/2009, and 2010/2011.
5.2. Possible Mechanisms of Latitudinal Difference Between GW and Convection Maxima

An interesting result from the analysis of climatology in section 4 is that the DJF-mean GW variance takes its maximum southward of the precipitation maximum by about 3°. In this section, we discuss three possible mechanisms causing the latitudinal gap of stratospheric GWs and tropical convection in the South Pacific. They are (1) island distribution, (2) selective excitation of southward propagating waves in the troposphere, and (3) southward refraction due to background wind shear in the stratosphere. Other mechanisms and a combination of these mechanisms are also discussed.

5.2.1. Island Distribution

Island distribution may affect the GW climatology because the occurrence frequency of deep convection over the land is higher than that over the ocean [Takayabu, 2002]. Thus, it is expected that GWs with long vertical wavelengths are generated more effectively over the islands. Such GWs with long vertical wavelengths have fast intrinsic phase speed and hence less frequently encounter their critical levels compared with those with short vertical wavelengths. In addition, such GWs with long wavelengths are more easily detectable by AIRS.

As a proxy of the existence of islands, we used mean ocean fraction for each bin, which is hereafter referred to as MOF. A map of MOF is plotted in Figure 12a. MOF values are zero over continents and one over oceans by its definition. Small but nonzero MOF values are observed around and in particular southward of SPCZ indicating that a number of islands and/or islands with large areas are distributed there.

Figure 12b shows a histogram as a function of the DJF-mean climatology of GW variance versus that of precipitation for a region of (0°S to 30°S, 160°E to 160°W) which is denoted by a rectangle on the map in Figure 12a. It is clear that the two quantities are positively correlated. Figure 12c shows the mean of MOF values at respective bins of this plot by the same color scale as used for Figure 12a. The mean MOF values are smaller at larger GW variance for a particular precipitation value. This result indicates that GWs are effectively generated from convection over islands.

![Figure 11.](image-url)  
(a) Correlation coefficients of DJF-mean GW variance with the DJF-mean zonal mean zonal wind at 10 hPa at the equator (QBO time series). Positive (negative) values are shown by red (blue) contours. Thin contours show ±0.62 (a significant level of 90%), and a thick contours show ±0.83 (a significant level of 99%). (b) Regression coefficients of DJF-mean GW variance to the QBO time series are shown by colors only in the regions where the correlation coefficient magnitudes are larger than 0.62.

![Figure 12.](image-url)  
(a) A map of mean ocean fraction from MERRA. (b) Histogram and contours for the precipitation versus the GW variance at 39 km in the region of (160°E to 160°W, 0° to 20°S). Contour interval is 2. (c) Mean ocean fraction as a function of precipitation and the GW variance at 39 km for the same region as for Figure 12b.
In Figure 10, composite MOF values were shown as a function of the latitude relative to the climatological SPCZ. It is seen for the climatology in Figure 10a that the MOF takes its minimum slightly southward of the precipitation maximum and slightly northward of the GW variance maximum. It is also seen from Figures 10b and 10c that the GW variance maximum does not move much and remains close to the MOF minimum, although latitudinal movement of the precipitation maximum is largely depending on the ENSO phase. This result is consistent with our inference that convection over islands effectively generates GWs with long vertical wavelengths and suggests that the island distribution is partly attributable to the difference in the dominant latitude between the observed GW variance and precipitation. In addition, it is worth noting that the diurnal cycle of convection near islands has a peak in evening [Mori et al., 2004; Ichikawa and Yasunari, 2008], while convection over the tropical Pacific and Atlantic Oceans is maximized in the morning [Serret and McPhaden, 2004]. Nighttime observations by AIRS which are used in the present study maybe more apt to detect GWs originating from convection near islands rather than those over the ocean. However, as the MOF minimum is always located slightly northward of the GW variance maximum (Figures 10a–10c), additional mechanisms causing southward shift of the GW variance maximum are necessary.

5.2.2. Selective Excitation of GWs in the Background Wind Shear

Beres et al. [2002] showed from a series of numerical simulation using a two-dimensional model that GWs propagating opposite to the upper tropospheric wind shear are effectively excited by convection in squall lines. Figure 13a shows vertical profiles of composite meridional winds for longitudes of 160°E to 160°W as a function of the latitude relative to SPCZ. Composites of precipitation and GW variance for the same longitude region are respectively shown in Figures 14b and 14c as a function of the latitude relative to the SPCZ. As was also shown in Figure 10a, the latitude of GW variance maximum is observed southward of the precipitation maximum by 3°.

Northward wind is observed in the upper troposphere with a maximum around 200 hPa and hence the vertical wind shear in the troposphere is northward. This wind structure suggests that southward propagating GWs should be more effectively excited from the SPCZ. This implication is qualitatively consistent with the relative location of the GW variance and precipitation maxima.

Figure 13. A composite latitude and height cross section of (a) the mean meridional wind at a contour interval of 0.5 m s$^{-1}$. Composite latitudinal profiles of (b) precipitation and (c) GW variance averaged for 160°E to 160°W. The reference latitude is the latitude of the precipitation maximum between 0° and 30°S.
A rough but quantitative estimation is made for a possible latitudinal propagation distance of GWs using typical wave parameters obtained from the S transform analysis: The mean horizontal wave number $k_h = \frac{2\pi}{225 \text{ km}}$ (Figure 3c) and a mean horizontal wave number direction of $-4.5^\circ$ (Figure 3d) around the location (15°S, 180°E) near the SPCZ. A typical observable vertical wave number $m$ is assumed as $\frac{2\pi}{15 \text{ km}}$. The inertial frequency $f$ at 15°S is $\frac{2\pi}{46.4 \text{ h}}$, and a typical stratospheric Brunt-Väisälä frequency $N$ is assumed as $\frac{2\pi}{5 \text{ min}}$. From the linear theory of nonhydrostatic internal inertia-gravity wave, vertical ($c_{gz}$) and meridional group velocity components ($c_{gy}$) are expressed as

$$c_{gz} = \frac{-k_h^2 m (N^2 - f^2)}{(m^2 + k_h^2)^{3/2} (N^2 k_h^4 + f^2 m^2)^{1/2}},$$
(2)

$$c_{gy} = \frac{k_h^2 m^2}{(m^2 + k_h^2)^{3/2} (N^2 k_h^4 + f^2 m^2)^{1/2}} \frac{k_h \sin \phi_h m^2}{(m^2 + k_h^2)^{3/2} (N^2 k_h^4 + f^2 m^2)^{1/2}},$$
(3)

Figure 14. Latitude and height cross sections of background zonal winds at (a) 160°E and (d) 160°W. Contour intervals are 2.5 m s$^{-1}$. Latitudinal distributions of the precipitation at (b) 160°E and (e) 160°W and those of GW variance at (c) 160°E and (f) 160°W. Thin vertical lines denote the latitudes of the maximum precipitation (Figures 14b and 14e) and the maximum GW variance (Figures 14c and 14f).
where $k_y$ is meridional wave number \((=k_x \sin \phi_y)\) and estimated using the above mentioned parameters at $c_{gr} = 3.31 \text{ m s}^{-1}$ and $c_{gr} = 3.89 \text{ m s}^{-1}$. Thus, the time period needed for propagation from the upper troposphere ($z = 9 \text{ km}$) to the stratosphere ($z = 39 \text{ km}$) is estimated at about 2.52 h, and the latitudinal distance over which the GWs migrate is 35.3 km. This value is not sufficient to explain the observed distance of $3^\circ \sim 330 \text{ km}$.

Thus, a preference for excitation of southward waves is not the only mechanism causing the large latitudinal distance between the GW variance and precipitation maxima, although it is qualitatively consistent.

### 5.2.3. Refraction Due To the Latitudinal Gradient of Zonal Wind

GWs tend to propagate meridionally by refraction in a background zonal wind having latitudinal shear \cite{Dunkerton, Sato et al.}. Meridional cross sections of the mean zonal wind are shown for two longitudes of 160°E and 160°W (Figure 14) instead of a composite, because changes largely depend on longitude (see Figure 3a). The latitudinal gradient of zonal wind, $\partial U/\partial y$, is mainly positive in the stratosphere in the latitude range between the GW variance and precipitation maxima. Assuming that the zonal wave number ($k_x$) is positive in the westward background wind, the $k_y$ tendency is negative from the ray tracing theorem,

$$\frac{dk_y}{dt} = -k_x \frac{\partial U}{\partial y} \tag{4}$$

This means that the GW packets would tend to propagate southward. This fact is at least qualitatively consistent with the difference in the latitude between the GW variance and precipitation maxima. A rough but quantitative estimation is next made. Acceleration of the GW packet in the latitudinal direction is written as

$$\frac{d^2 y}{dt^2} \frac{dc_{gr}}{dt} = \frac{m^2}{(m^2 + k_y^2)^2} \frac{N^2 - f^2}{(N^2 k_y^2 + f^2 m^2)^2} \frac{dk_y}{dt} = \frac{k_x m^2}{(m^2 + k_y^2)^2} \frac{N^2 - f^2}{(N^2 k_y^2 + f^2 m^2)^2} \frac{\partial U}{\partial y} \tag{5}$$

To isolate the refraction effect on the meridional propagation direction, $dy/dt$ is set to zero at the initial time. The zonal wind shear $\partial U/\partial y$ is simply set to a constant value of 4.5 m s$^{-1}$ per 3° at 30 hPa and (160°E, 7.5°S) during the propagation (Figure 14). The latitudinal propagation distance is estimated at 28 km for GWs using $k_x \sim k_h = 2\pi/(225 \text{ km})$. This value is again not sufficient to explain the observed distance of $\sim 330 \text{ km}$.

The latitudinal propagation due to refraction is not the only mechanism to cause the difference in the latitudes between the GW variance and precipitation maxima.

### 5.2.4. Other Possible Mechanisms and a Combination of Multiple Mechanisms

We considered other mechanisms such as advection by the southward background wind and critical level filtering at the latitudes of the precipitation maximum. However, none of them can explain the southward shift of the GW variance maximum. Meridional background wind is almost zero in the stratosphere below 3 hPa (Figure 13a) and hence cannot cause much advection. The background zonal winds are not very different between the latitudes of the GW variance and precipitation maxima (Figure 14). The eastward wind around the tropopause is rather stronger at higher latitudes (Figure 14d), which means that GWs at higher latitudes can be more effectively filtered. An effect of the observational filter by the stratospheric zonal wind is also possible. However, this is not the case at least around 160°W as seen in Figure 14d where the zonal wind around 3 hPa does not vary much with the latitude around the precipitation maximum.

In conclusion, the most important mechanism explaining the latitudinal distance about 330 km between the GW variance and precipitation maxima is the island distribution which is dense (sparse) southward (northward) of SPCZ. The selective GW excitation in the vertical shear of mean meridional wind and the latitudinal propagation by refraction due to the latitudinal shear of mean zonal wind have secondary contribution (about 63 km in total). Probably, a combination of these mechanisms is likely responsible for the latitudinal difference.

### 6. Summary and Concluding Remarks

The present study first examined the climatology and interannual variability of GW variance in the subtropical region in the summer middle stratosphere based on satellite nadir sounding data by AIRS over 8 years. High-resolution temperature data at 39 km made from the \textit{Hoffmann and Alexander} [2009] retrieval algorithm were used for the analysis. An $S$ transform method was applied to extract GW parameters such as temperature...
variance and the magnitude and direction of horizontal wave number. In a climatology, large GW variance is observed over continents and the tropical Maritime Continent in both hemispheres. Precipitation is also dominant over the continents, but there is a systematic latitudinal difference between the GW variance maximum and precipitation maximum by about 3°.

The interannual variability in the summer subtropics is larger in the SH than in the NH. Thus, further analysis was focused on the SH. Horizontal wavelengths are longer (>200 km) over continents and the Maritime Continent and shorter (about 150 km) over the ocean. Assuming that the zonal phase speeds are opposite to the background zonal wind as is consistent with previous studies, the waves propagate primarily eastward, but the latitudinal component of the wave number vectors is negative (i.e., southward) for most GWs.

An interesting and important feature is that the interannual variability of the GW variance in the western and central South Pacific region in summer is closely related to the ENSO which accounts for 70% of the variation. This variation of GW variance follows the SPCZ latitudinal movement in association with the ENSO. The distribution of both horizontal wavelengths and propagation direction also vary following the ENSO. The contribution of the equatorial QBO is minor in that region.

Last but not least, we examined possible mechanisms causing the systematic latitudinal difference by 3° between the maxima of GW variance and precipitation climatology. An important mechanism is the distribution of islands which are dense southward of SPCZ. It is expected that deep convection excited over islands effectively generates GWs with long vertical wavelengths, which are more easily detectable by AIRS. Selective GW excitation due to vertical shear of the upper tropospheric wind, and GW refraction in the latitudinal shear of the background wind are secondary but important mechanisms for the southward component of propagation of GWs. By using typical GW parameters estimated from AIRS data, the sum of the two mechanisms might account for about 20% of the latitudinal distance. Combination of the three mechanisms are likely responsible for the latitudinal difference.

This study showed a significant interannual modulation of stratospheric GW activity by ENSO in the SH subtropical region. It is seen from comparison between Figures 3b and 6d that ENSO modulation of the GW variance is more than 10% depending on the location. This fact means that the meridional circulation in the middle and upper atmosphere may be also modulated by ENSO. Changes in the meridional circulation also modify the thermal structure and affect the structure of tides, which are dominant in the upper mesosphere and thermosphere. Thus, climate models using a parameterization with constant, prescribed GW sources versus a convection-based GW scheme may not represent the GW interannual variability related to ENSO and its effect on the meridional circulation.

For a more quantitative discussion, it is necessary to examine the momentum flux associated with GWs. To do this, the estimation of vertical wavelengths is needed using data from at least two altitudes in addition to the temperature variance. However, generally speaking, this is difficult to derive from nadir-viewing satellite observations with low vertical resolution like AIRS. The momentum flux is expressed using a formula $\frac{1}{2} \rho f \frac{d}{dt} \left( \frac{\mathbf{v} \cdot \mathbf{k}}{\mathbf{k}} \right)^2$ from observed temperature variances [Ern et al., 2004; Alexander, 2015]. Thus, we assume a typical detectable vertical wavelength of 15 km for a rough estimation. Using a climatological mean GW variance of 1 K², a background temperature of 250 K, a typical horizontal wavelength of 225 km, and damping due to limited vertical resolution of AIRS retrieval of about 15% in variance [Hoffmann and Alexander, 2009], the climatological momentum flux observed by AIRS is estimated at about 0.5 mPa and the interannual variability related to ENSO is about 0.05 mPa. This climatological momentum flux value of the GWs observed by AIRS is comparable to the estimate (about 0.5 mPa) around the SPCZ at 40 km in January 2006 from observations of Sounding of the Atmosphere using Broadband Emission Radiometry and HIRDLS, which are sensitive to GWs with short vertical wavelengths unlike AIRS [Geller et al., 2013].

In addition, it is also worth noting that the GW variance dependence on longitude has an interannual variability. This means that the Lagrangian mean circulation in the middle atmosphere may have significant three-dimensional structure, although it has mainly been examined in the two-dimensional meridional cross section so far. It should be interesting to examine the three-dimensional structure of the interaction of GWs with the mean flow [e.g., Kinoshita and Sato, 2013; Sato et al., 2013] and the interaction between GWs and
Appendix A: Noise Spectra

It is expected that there are few significant GW sources such as topography, jet-front systems, and convection in the winter subtropical Pacific. In addition, the background wind there is generally weak at 3 hPa, and hence, GWs originating from convection are not significantly Doppler shifted. Such GWs have short vertical wavelengths that are hardly detected by AIRS. Thus, we regarded the magnitude of the $S$ transform cross spectra of adjacent data series in such regions as “noise” spectra which is a function of the location and wavelength.

Figure A1 shows the noise spectra obtained from adjacent two data series across the track for the region of $10^\circ$N to $30^\circ$N, $150^\circ$E to $120^\circ$W on 12 February and the region of $0^\circ$S to $25^\circ$S, $180^\circ$W to $90^\circ$W on 12 July 2003–2011. The spectral densities, which we call variances, are larger at shorter wavelengths. At nadir where the cross-track location is 0 km, the variance is maximized at a wavelength of about 30 km. Similar maxima are observed at longer wavelengths for larger distances from the nadir. Such dependence of the maximum wavelength can be explained by the coarser resolution at larger distances from the nadir. These maxima are likely due to the random noise that appears in the temperature retrievals and hence should be removed. A weak peak is also observed around 1000 km wavelength. The reason of this peak is not clear but may be due to the detrending method used in the present study. This peak should also be removed as noise. Thus, we examined $S$ transform spectra in the range of wavelengths $70$–$700$ km. Note that the variances are diminished near the edge of a cross-track scan. This reflects to the cosine-shaped window function applied to the original data before the $S$ transform calculation. Thus, the edge regions are not examined for the analysis either.

Two examples of the $S$ transform spectra including GW signals are shown in Figure A2. Figures A2a and A2b respectively show the results over convection in Australia on 15 January 2007 and over the Andes on 8 May 2006. The latter corresponds to a significant GW event examined by Alexander and Teitelbaum [2011]. Clear GW signals are observed in both examples, occurring at a wavelength of 300 km and a distance of $200$ km in Figure A2a and at a wavelength near $100$ km at nadir in Figure A2b.