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Gravity wave variances and propagation derived from AIRS radiances

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Abstract

As the first gravity wave (GW) climatology study using nadir-viewing infrared sounders, 50 Atmospheric Infrared Sounder (AIRS) radiance channels are selected to estimate GW variances at pressure levels between 2–100 hPa. The GW variance for each scan in the cross-track direction is derived from radiance perturbations in the scan, independently of adjacent scans along the orbit. Since the scanning swaths are perpendicular to the satellite orbits, which are inclined meridionally at most latitudes, the zonal component of GW propagation can be inferred by differencing the variances derived between the westmost and the eastmost viewing angles.

Consistent with previous GW studies using various satellite instruments, monthly mean AIRS variance shows large enhancements over meridionally oriented mountain ranges as well as some islands at winter hemisphere high latitudes. Enhanced wave activities are also found above tropical deep convective regions. GWs prefer to propagate westward above mountain ranges, and eastward above deep convection. AIRS 90 field-of-views (FOVs), ranging from +48° to −48° off nadir, can detect large-amplitude GWs with a phase velocity propagating preferentially at steep angles (e.g., those from orographic and convective sources). The annual cycle dominates the GW variances and the preferred propagation directions for all latitudes. Quasi-biennial oscillation (QBO) signals are also found in the tropical lower stratosphere despite their small amplitudes.

From 90 AIRS FOV radiance measurements, we are able to clearly identify measurement noises, high-frequency internal GWs, and low-frequency inertia GWs. Even though the vertical wavelengths of inertia GWs are shorter than the thickness of instrument weighting functions, simulations support the AIRS sensitivity to these waves. The novel discovery of AIRS capability of observing shallow inertia GWs will expand the potential of satellite GW remote sensing and provide further constraints on the GW drag parameterization schemes in the general circulation models (GCMs).

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1 Introduction

Gravity waves (GWs) are known to play a key role in global climate and weather dynamics by transporting energy and momentum from the lower to upper atmosphere, transport which is essential for determining the general circulation and temperature structure in the stratosphere and mesosphere (e.g., mesosphere wind reversal, quasi-biennial oscillation, etc.). GWs are of particular importance for wave dynamics in the summer hemisphere in the stratosphere where planetary waves are weak, and are important in the entire mesosphere. As an example, model simulation results suggest that the mesosphere summer easterlies would arrive one month late without including gravity wave drag (GWD) in the model (Scaife et al., 2002). GWs also affect weather and chemistry. For example, temperature fluctuations associated with GWs can lead to formation of polar stratospheric clouds (PSCs) in a synoptically warm condition and susequently affect ozone depletion (Hamill and Toon, 1991).

The importance of gravity waves on climate and weather became widely recognized in recent decades due to progress in observational techniques and model improvements. Two issues remain as major concerns, however. Firstly, most of the GWs cannot be resolved in GCMs and hence need to be parameterized except those with long wavelengths and low frequencies. These prescribed parameters, formulated to be dependent on detailed characteristics of GWs and GW sources, however, are poorly constrained by observations, and are actually heavily "tuned" in most of the cases to attain a realistic atmosphere. For example, one of the major GW sources, convection, is often assumed to be uniformly distributed all over the globe in GCMs with spectral GWD parameterization schemes, which is obviously far from reality (Alexander and Rosenlof, 2003). The other important non-stationary GW source – jet imbalances – is not included in most of the GCMs (Kim et al., 2003). Secondly, different observational instruments can only see partial gravity wave spectra. Although solving this jigsaw puzzle is underway (e.g., Alexander et al., 2010), the high-frequency portion of the spectrum is invisible to most of the instruments (Wu et al., 2006; Alexander et al.,

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2010), and the momentum flux associated with these waves remain uncertain. According to the gravity wave dispersion relationship, high-frequency GWs tend to have more vertical fluxes of the horizontal momentum, and hence may exert appreciable impacts on the mean flow with momentum deposition. Therefore, the underestimation of these waves might result in serious shortfall for constraining the gravity wave drag (GWD) parameterization schemes in current climate models.

Among various GW observations, satellite instruments have advantages over others (e.g., radiosonde, radar, aircraft measurements) in terms of high and regular sampling density and global coverage. As an example, the Aqua satellite, which carries the Atmospheric Infrared Sounder (AIRS), samples each $2^\circ \times 2^\circ$ latitude-longitude gridbox at least 24 times per month. In addition, intercomparisons and cross-validations can therefore be readily carried out among different satellite instruments on a global basis. Based on their viewing geometry, satellite GW observational instruments can be categorized into three groups: limb, sub-limb, and nadir sounders. The limb sounders (e.g., LIMS, CRISTA, HIRDLS, GPS) are sensitive to GWs with small λ_z/λ_h ratio, i.e., low-frequency GWs. Here λ_z and λ_x denote vertical and horizontal wavelength, respectively. The nadir sounders (e.g., AIRS, AMSU-A, SSMIS) are mostly sensitive to high-frequency GWs, and the sub-limb sounders (e.g., MLS) are sensitive to mid-frequency GWs. More details can be found in Wu et al. (2006).

As a nadir viewing sounder, AIRS has four supurb properties for GW studies. Firstly, its measured radiances are very sensitive to high-frequency GWs, which are likely to be underestimated so far in both observations and models. Secondly, AIRS radiances are a direct measure of GW induced air temperature perturbations, and hence provide a more accurate measurement of GWs than retrieved temperatures. Thirdly, AIRS has a high horizontal resolution (~ 13 km at nadir) which makes it more attractive in detecting the high-frequency, short λ_h GWs. Lastly, unlike the limb sounders, AIRS sees GWs selectively depending on its viewing angles. This allows us to estimate preferred GW propagation directions from the viewing-dependent variance difference between the two outmost off-nadir views.

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This paper is organized as follows. The following section gives a brief introduction of the AIRS instrument, and a description of the methods we use to retrieve GW properties. In Sect. 3, a climatology of wave variance and the preferred zonal propagation direction will be derived in terms of zonal-mean, geographical distribution and temporal variations. Interpretation of the results is discussed in detail in Sect. 4, where we emphasize on two major gravity wave sources – topography and convection, and interannual variabilities. Wave enhancements in the equatorial lower stratosphere and the winter pole at 10 hPa will be discussed in a great length, and a numerical experiment is performed to support the speculation on the observed GWs being low-frequency waves. Concluding remarks can be found in Sect. 5.

For reader's convenience, we hereafter refer to "internal GWs" as high-frequency ($\varpi \gg f$), long vertical wavelength ($\lambda_z > 10$ km) and short horizontal wavelength waves ($\lambda_h \sim 100$ km or less), and use "inertia GWs" to represent low-frequency ($\varpi \sim f$), short vertical wavelength ($\lambda_z < 10$ km) and long horizontal wavelength ($\lambda_h \sim 1000$ km) waves, where ϖ and f denote wave frequency and the Coriolis parameter. One should keep in mind that GWs in the stratified atmosphere are essentially all internal waves.

2 AIRS instrument and method

AIRS is an infrared spectrometer and sounder that contains 2378 channels in 3.74–4.61 µm, 6.20–8.22 µm and 8.8–15.4 µm wavebands. It makes a cross-track scan every 8/3 second that includes 90 footprints on the ground, 4 independent views over the cold space, and 3 views into 3 different calibrators (Aumann and Miller, 1994). The scan swath is ~1600 km wide, reaching ±48.95° from nadir. The angle difference between two adjacent footprints, therefore, is about 1.1°, which corresponds to ~13 km footprint size at nadir. The satellite orbits over 24 h are divided into 240 granules, each of which contains 135 scans (6 min duration).

We use AIRS CO_2 15 µm radiance emission bands with wavenumbers ranging from 665 cm^{-1} to 693 cm^{-1} in this study. 50 channels are selected with 11 distinguished

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weighting functions (WFs) peaking between 2 hPa and 100 hPa. The channel numbers are listed in Table A1 in the Appendix A. Ideally, only channels at the crests or troughs of the radiance spectra, or paired channels at two side-wings of peaks or troughs should be used since they are mostly stable with small detector drifts on the spectra, which mainly occur due to the expansion/contraction of the glass filter depending on whether it faces the Sun or in darkness (i.e., ascending/descending of the orbits). However, we are losing information over most of the middle to upper stratosphere by only selecting these channels. With this consideration, several channels on one side wing of peaks/troughs (bold numbers in Table A1) are also selected. Results from ascending and descending orbits have been compared carefully, and no significant differences have been found whatsoever, so this problem shouldn't cause major problems or failures of using the wing channels in our research.

Since AIRS scans are always perpendicular to the orbit track, the scan line can therefore be treated approximately as along the west-east direction for most of the time, except at high latitudes where the scan is most meridional. We hence exclude the data beyond ±80° latitudes.

The variance of the radiance (σ^2) has three components:

$$\sigma^2 = \sigma^2_{\mathrm{GW}} + \sigma^2_{\mathrm{noise}} + \epsilon^2 \tag{1}$$

Normally other variance (ϵ^2) is much smaller than the first terms on the right-hand of Eq. (1), and hence is omitted here. Evidence shows that AIRS may observe some turbulence signals or small vertical wavelength inertia GWs, which will be discussed later on in Sect. 4.3. However, these additional sources of variance are highly uncertain, and we would rather include them in σ^2_{GW} than separating them out. Since the instrument noise is on the same order with the magnitude of the GWs, it needs to be carefully evaluated and subtracted consequently. The instrument noise ($\sigma^2_{\mathrm{noise}}$) is estimated in an analagous way to Wu and Waters (1996). Instead of using pre-launch information, we feel it is more accurate to estimate the random component of the noise from the real data with the assumption that $\sigma^2_{\mathrm{noise}}$ only depends on the frequency channel, and

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should not vary with location, time and viewing angle. To extract the radiance noise, we first apply a 3-pt running smooth window to each independent scan and compute the variance as the squared difference between the original and smoothed data (referred as “3-pt result” hereafter). Then, the minimum variance from the monthly averaged $2^{\circ} \times 2^{\circ}$ map for each month is determined. Because the 3-pt variance is affected least by GWs, the minimum variance is assumed to be very close to the measurement noise. These minimum variances are found always coming from the summer hemisphere high-latitudes. Finally, the mean is computed for all the variances that are below the median of the timeseries, and we attribute this mean as the instrument noise σ^2_{noise}. The 3-pt variance is used for the noise estimation since it gives the smallest cut-off wavelength and hence contains the most high-frequency information. Nevertheless, the 3-pt result still contains a fair amount of GW contributions in that the latitudinal and geographical distributions of the 3-pt result look similar to 7-pt result even without noise subtraction (not shown). The noise estimation method used here is confirmed to be reasonable and accurate by the fact that variance reaches minimum at summer hemisphere high latitudes, and is quite uniformly and stably distributed there throughout the time. Moreover, there are no statistically significant differences among different viewing angles there.

We furthermore compared the noises estimated by the above method with the NEdT (Noise Equivalent Delta Temperature) listed in AIRS property reports (http://airs.jpl.nasa.gov/data_products/algorithms/), and as one can easily tell from Table A1 in the Appendix A, they agree very well. A linear increasing trend has been found in the estimated noise at various levels, which is on the same order of the trend found in the AIRS documentation. Since this trend is far from significant (regression coefficients on the order of $10^{-5}\,\text{K}^2\,\text{month}^{-1}$), which is probably due to the run-off of the sensors, we do not remove the trend in this study.

As suggested by Wu and Waters (1997) and Wu and Eckermann (2008), the minimum detectable GW variance (σ^2_{GW}) is $\sigma^2_{\text{min}} = \sqrt{2(M-2)/N}\sigma^2_{\text{noise}}$, where M is the number of points used in this filter for variance estimation ($M = 7$ in this case). Since σ^2_{noise}

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is determined from a monthly mean $2^\circ \times 2^\circ$ map, a zonal average with N points in this grid size can further reduce the detection threshold. The outmost $M/2$ points are used to get an averaged variance value, which also needs to be considered. The minimum detectable values at various pressure levels are listed in Table A1 in the Appendix A.

We cannot directly calculate the GW horizontal propagation direction from radiance measurements of a single scan. However, the preferred zonal propagation direction can be indirectly inferred from the difference of GW variances between two different FOVs. This is illustrated in a schematic picture Fig. 1, where a typical mountain GW packet is generated downstream of the mountain under a westerly wind. Wave phases, though stationary in the Earth frame, are propagating downward and westward relative to the westerly mean flow, while energy is propagating upward and westward in this case. In the west field-of-view (FOV) of AIRS, the wave amplitudes are largely smeared out, while the east FOV would yield a much greater reading. Therefore, negative (positive) value of ($GW_w - GW_e$) reflects a predominant westward (eastward) zonal propagation. Since multiple factors affect the magnitude of the difference, such as the viewing angle relative to the phase front, the original GW amplitude, etc., the absolute value of the difference does not have a definitive physical meaning. As long as it is statistically significant, the sign of the difference is the variable that really matters. This approach has been formerly used on Aura MLS observations to study GW meridional propagation direction, and has turned to be a very useful and effective method (Wu and Eckermann, 2008).

Before presenting the results, one should note that the majority of the GW signals in AIRS comes from high-frequency GWs. The visibility can be computed as a function of along-track wavelength (λ_x) and vertical wavelength (λ_z) by convolving the weighting function (WF) with a plane wave of unity amplitude (1 K) (McLandress et al., 2000). In the along-track direction, the peak of the sensitivity is mainly determined by the length of the truncation window, as suggested by Fig. 2. The left half of three curves, which is similar for the 3-pt, 7-pt and 15-pt results, is controlled by the width of the AIRS FOV. All three rise sharply once λ_x becomes larger than the footprint size, which

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is ~ 13 km at nadir. The length of the smoothing window decides the peak as well as the tail of the filter. In the vertical direction, AIRS can hardly detect GWs with λ_z shorter than the full width at half maximum (FWHM) of the WFs, which is ~ 12 km. However, there are some exceptions, which will be discussed later in Sect. 4.3. The sensitivity increases monotonically with the increase of the λ_z before reaching the peak at $\lambda_z \approx \lambda_{x\max}$, where $\lambda_{x\max}$ denotes the most sensitive along-track wavelength. This can be better understood from revisiting Fig. 1, where the largest amplitude would be reached when the GW fronts are parallel to the viewing angle. For a 7-pt window, the peak sensitivity occurs when both λ_x and λ_z equal to ~ 100 km. Alexander and Barnet (2007) studied multiple mountain GW events over the tip of Andes and Antarctic Peninsula, and found out the most dominant λ_x is around 100–150 km. So the 7-pt window can effectively capture these GWs. With a wider window, we are at the cost of possible inclusion of other waves (e.g., Kelvin wave), and we may lose the dominant GWs as suggested by Alexander and Barnet (2007). Nevertheless, the major conclusion's robustness does not depend much on window size. We hence use 7-pt results in this paper.

3 Climatology of gravity wave variances and the preferred zonal propagation direction

The monthly mean latitudinal, geographical and temporal variations will be presented in this section. Year 2005 is used to indicate the mean GW climatology, which represents a condition with no strong sudden stratospheric warming nor ENSO events.

3.1 Latitudinal distributions

Figures 3 and 4 plot the monthly averaged zonal mean 7-pt radiance variance as a function of latitude and height in January and July of 2005, respectively (with estimated instrumental noise deducted), overplotted with UK Met Office monthly mean zonal winds.

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The variances can be explained by GWs, since they in general grow exponentially with height at mid-high latitudes. In the middle to upper stratosphere (above 20 hPa), the two largest variance centers colocate with the westerly and easterly jet cores in the upper stratosphere, and the one in the winter hemisphere is apparently stronger. This indicates that large amplitude GWs (or frequently generated GWs) are from sources beneath the two jet centers. The overall feature is similar to previous GW studies using Aqua AMSU-A, UARS MLS and Aura MLS (Wu et al., 2006) with comparable strengths in the winter hemisphere, but weaker amplitudes in the summer hemisphere. However, the wave enhancement at the 10 hPa near the winter pole is a new feature revealed by AIRS data. An interpretation of this feature is inertia GWs, which will be discussed in Sect. 4.3. The GWs are almost tranquil at the summer polar region in the AIRS observation. Even without the deduction of estimated noise, the variance there is still significantly smaller than other satellite observations. AIRS observations seem to suggest that high-frequency internal GWs have difficulty propagating or being generated at the mid-high latitudes of the summer hemisphere under weak winds. The presence of the zero wind line in the summer hemisphere is another key reason whereby orographic GWs are mostly removed.

In the lower stratosphere, AIRS GW variance is generally larger in the tropics and decreases with latitude. Similar results have been found in some measurements that represent the behavior of the inertia GWs (e.g., Wang and Geller, 2003; Ratnam et al., 2004; Preusse et al., 2006) and middle frequency GWs (e.g., Wu and Eckermann, 2008). AMSU-A, which has the same scan angle as AIRS, and hence which is supposed to be also sensitive to internal GWs, observes features opposite to AIRS at the equatorial lower stratosphere where the GW variance reaches minimum (Wu et al., 2006). This interesting feature will be discussed in details later on in Sect. 4.3.

As stated in Sect. 2, one can indirectly infer the GW preferred zonal propagation direction from the difference of the GW variances between the two outmost views. As in the right columns of Figs. 3 and 4, the zonal mean zonal propagation directions, on a monthly scale, always tend to be in the opposite direction to the mean zonal

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winds, which is in general westward (eastward) in the winter (summer) hemisphere. This is true especially for large amplitude GWs at high altitudes, as will be shown in the geographical patterns in the next section. Mountain forcing is one of the most prominent GW sources in the winter hemisphere. The high-frequency mountain waves generated in the mid-latitude mountain areas can propagate exclusively in the relatively strong westerly wind to make a zero ground-based phase speed. The overall pattern in Fig. 3c does not strictly follow the mean zonal wind contours, as GWs always try to drag the total wind, and hence the meridional components of both the wind and the GW propagation need to be taken into account. Note that there is no significant difference between two outmost views at the 10 hPa winter pole and tropical lower stratosphere, which means that the same amount of eastward propagating and westward propagating GWs exists at those levels. Simulation results from Sect. 4.3 confirm this inference.

Very similar features can also be obtained with the 3-pt results, though the amplitudes are much smaller (not shown). Considering the fact that the 3-pt running smooth window is a high-pass filter, we can conclude that the majority of GWs observed by AIRS are really high-frequency, short horizontal wavelength internal GWs.

3.2 Geographical distributions

Monthly mean geographical maps are plotted in Figs. 5 and 6 at various levels to represent typical situations of January and July. Only the variance that is larger than $2\sigma_{min}$ is colored, where σ_{min} is the minimum detectable variance. Zonal mean wind (black solid contours) and wind speed at 700 hPa that exceeds 10 m s^{-1} (brown hatched areas) are overplotted on the variance maps. Such orographic GWs have been seen through other satellite instruments, such as Aura MLS (Wu and Eckermann, 2008) and AMSU-A (Wu et al., 2006). The GW amplitudes in AIRS, however, are the strongest among all even after taking into account the averaging effect. This indicates that mountain GWs likely possess greater amount of energy toward the high-frequency part of the frequency spectrum (relative to the mean wind), and AIRS is particularly sensitive to them due to its high resolution and the steep viewing angle. In both hemispheres,

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orographic GWs propagate poleward as they travel upward, in agreement with what has been found from Aura MLS observations (Wu and Eckermann, 2008). The orographic GWs are closely associated with the upper level polar night jet, and hence the location of the enhancements vary from year to year with the shift of the polar night jet. Comparing the 10 hPa map with the 2.5 hPa map, significant increase of GW activities is found in Alaska in January, 2005 (Fig. 5) and West of Antarctic Peninsula in July, 2005 (Fig. 6). They are responsible for the peak at the 10 hPa pole in the zonal mean maps Figs. 3 and 4. They are highly variable along the longitudes, and not necessarily related with the jets. Interestingly, the Brunt-Väisälä frequency (N^2) also peaks at the same locations for these two particular months (not shown). We will discuss about this feature later on in Sect. 4.3.

In the subtropics and tropics, large GW activities are found in the upper stratosphere over the deep convective regions. The deep convective regions are identified from the ice water content (IWC) from Aura MLS (Wu and Eckermann, 2008). In particular, they are Western Pcific warm pool region, Amazon rainforest region, and Central Africa rainforest region for NH winters, and Southeastern US monsoon region and India-South China monsoon region for SH winters. Comparing with Aura MLS, the magnitude of these convectively generated GWs seen from AIRS are smaller. This is consistent with the convective source spectrum suggested by Beres et al. (2004), where source momentum fluxes decrease rapidly at high-frequency part of the spectrum. The AIRS waves tend to propagate eastward relative to the mean wind, but the differences between the two outmost views have values that are marginally above the minimum detectable ones, which means both eastward and westward propagating GWs exist in the upper stratosphere, and the westward propagating ones are smaller than the eastward propagating ones with respect to the integrated magnitude. Convection is known to generate a broad spectrum of GWs that propagate in both directions at the top of the deep convection (e.g., Alexander et al., 1997; Beres et al., 2004). Afterwards, the majority of the low-frequency westward propagating GWs are wiped out by the easterlies in the tropics and subtropics, but some of the high-frequency waves may

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survive because they generally have higher phase speeds. These convectively excited high-frequency GWs do not grow into a comparable strength with mountain GWs until they reach the upper stratosphere. In the lower stratosphere, a uniform belt of GW enhancements is found in the tropical region, with some highlights in the deep convective regions, which creates the tropical maximum in the lower stratosphere in the zonal mean maps. This belt has been seen by previous studies using various GW observational instruments (e.g., radiosonde, GPS, etc.). It apparently is not closely associated with the convective sources. It is very interesting that AIRS can see it but AMSU-A, as another nadir viewing sounder, doesn't see it at all. The belt will be revisited later on in Sect. 4.3, where we claim the belt is caused by the propagation of inertia GWs there.

In the high latitude summer hemisphere, GW signals are barely detectable. The 3-pt results without subtracting noises yield a quite uniform distribution of "background-like" variance, which is independent of location and wind. In the NH summer (Fig. 6), it is easy to explain the NH serenity since the wind at 700 hPa is too weak to generate large orographic GWs (brown hatched regions barely exist in NH in Fig. 6). In the SH summer, we do see some enhancements over the Andes in the lower stratosphere (Fig. 5i), but the waves hardly grow upward probably because of the critical level filtering effect at the zero-wind line.

Large amplitude mountain GWs vary their location year to year (not shown). The $10\,\mathrm{m\,s^{-1}}$ wind speed at 700 hPa roughly provides a good threshold for generating noticable topographic GWs in the stratosphere. Only at places where westerly wind is consistently strong from low troposphere to upper stratosphere can those GWs grow into significant strengths. Locations of high occurrence of convectively generated GWs are also closely associated with the movement of tropical deep convective zones.

3.3 Temporal variations

An apparent annual cycle with maximum in the winter hemisphere over mid-high latitudes is found throughout the region of interest in both the variance and difference timeseries. We pick the levels of 2.5, 10, 40 and 80 hPa for the timeseries in Fig. 7,

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together with the monthly zonal wind derived from UK Met Office dataset. Since the annual cycle is highly repeatable, multi-year AIRS variances are lumped together in Fig. 7 as well as the UKMO winds. The annual cycle found in AIRS is consistent with other observations obtained by MLS (Wu and Eckermann, 2008), radiosonde (Wang and Geller, 2003), GPS (Tsuda et al., 2000), etc. Eckermann (1995) attribute the annual cycle in both lidar and rocket sounding observations to seasonal variations in the density stratification of the atmosphere (i.e., N^2). Besides these major factors, source properties can also contribute to the variations in that stronger westerly winds during hemispheric winters tend to create larger mountain GWs, which propagate further downstream under preferable conditions.

The Southern Hemisphere differs from the Northern Hemisphere in terms of the location of the maximum variance. In the Northern Hemisphere, wave variance nearly always peaks at the time when jet is the strongest (December–Feburary), as do the differences. Interestingly, in the Southern Hemisphere, the variance peak occurs ahead of the peak of the westerly jet below 10 hPa, but the difference between the two outmost viewpoints follows the variation of the jet closely, both of which propagate northward in time (May through October). Below 10 hPa, GW amplitudes are largest in July, but the mean wind maximum comes in one month later. A close look at the vertical configuration of the zonal wind reveals that tropospheric westerlies in the SH high latitudes actually peak in July (not shown). Another plausible explanation is based on the fact that the coastline of Antarctica is approximately parallel to the latitude lines. Therefore, meridional winds are more important than zonal winds there in generating large amplitude waves. Actually, the meridional wind contours follow the change of GW variance more closely than zonal wind contours (not shown). The Andes seem to be vital at high levels (above 10 hPa) at the intensification stage of the jet speed, while Antarctic Peninsula is more important in lower levels and seems to be a major factor causing the northward movement of the center of polar night jet. This phenomenon is also evident in Fig. 6. Note that the polar night jet in SH winter (e.g., Fig. 3) curves poleward (equatorward) below (above) 10 hPa. Antarctic Peninsula and the Andes perhaps play

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different roles at different levels in the formation of this shape. In the NH winter, there is no such northward curvature in either the jet center or the center of wave variance maximums. This result again suggests the importance of mountain GWD in shaping and regulating the polar night jets.

Strong interannual variability exists in NH high latitudes (not shown). Polar night jet splits during the winters of 2004, 2006 and 2009 at the polar region, and GW variance reduces a significant amount simultaneously. These are the years of strong sudden stratospheric warming (SSW) events (Lee et al., 2009). Lee et al. (2011) observed enhanced planetary wave activities during the SSW events from Aura MLS observations, and they believe the planetary waves serve as an efficient filter to prevent upward propagation of GWs. More explanations can be found in Dunkerton and Butchart (1984).

The annual cycle also dominates the subtropical stratosphere, but completely opposite phases are found between the upper stratosphere and lower stratosphere. In the upper stratosphere (first row of Fig. 7), GW variance as well as difference peak during hemispheric summers. Zonal mean maps (Figs. 5 and 6) show that the convectively generated GWs grow significantly in the upper stratosphere, which collocate with the easterly jet centers as well. Therefore, this summer peak could be largely explained by the convective source properties. In the lower stratosphere, winter maximum dominates, and this should be again attributed to the variations of density stratification similar to that of the mid-high latitudes (Eckermann, 1995).

The tropical lower stratosphere has an annual cycle with NH winter maximum, and this largely agrees with the fact that deep convections in the SH summers are in general stronger (Liu et al., 2007). QBO signals are relatively weak but still significant compared with the annual cycle, showing very interesting features that will be discussed in the next section. In the middle to upper stratosphere, a prevailing semi-annual cycle is found (not shown), which we believe is a combined effect of background winds (semi-annual oscillations there) and the source (convection crossing the equator twice per year at equinoxes).

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4 Discussions

Some new features from AIRS GW observations are worth further discussions. In this section, we will provide in-depth analyses on GW sources, interannual variations, and enhancements from inertia GWs.

4.1 Orographic and convective GW sources

Topographic GWs have been extensively studied during the last decades, and relatively mature parameterization schemes of the mountain GWDs have already been employed widely in state-of-art GCMs. These schemes usually take into account factors such as mountain height, surface wind speed, atmospheric stability and the alignment of the mountain ridges (McFarlane, 1987; Bachmeister, 1993). The non-stationary convective GWs received less attention compared to the orographic sources until recent years (Song and Chun, 2005; Beres et al., 2005). As to the parameterization of these GWs, some treat them as "moving mountains", and some think they create much broader spectra (e.g., Song and Chun, 2005). Both of the orographic and convective GWD parameterizations have free parameters to represent the wave intermittency, which is, by far, the poorest constrained parameter. Other "tunable" parameters include horizontal wavelength, source momentum spectrum, etc. AIRS, as one powerful GW observation instrument, can certainly improve our global understanding on the "observational constraints" of the parameterization schemes. For example, from the geographical distribution that has been discussed in Sect. 3.2, we can tell that mountain GWs tend to have larger energy variance toward the high-frequency part of the energy spectrum, while it seems to be the other way around for the convective GWs. Besides, some interesting features are found in AIRS, which are quite different from what's been observed before by other instruments.

To look into more details, we pick two orographic GW cases and one convective GW case that are evident in January 2005, which occur at the Rockies, the warmpool region, and Iceland. The monthly mean FOV-dependent GW variance within the white grid box is plotted at 2.5 hPa and 80 hPa for both January, 2005 and 2008 in Fig. 8. At

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first glance, the linear regression line confirms our idea that orographic GWs in general propagate westward while tropical convective GWs propagate eastward, both relatively to the local mean zonal wind. Ideally, if a monochromatic wave exists, we should see a smooth curve with a single peak or no peak depending on whether the upward propagation angle is smaller or larger than 48°. However, multiple peaks stand out in almost all curves in Fig.8d–p, indicating a spectrum of GWs are generated down below. In the lower stratosphere (Fig. 8k–p), convectively excited GWs may have sporadic spikes (Fig. 8i,o) while orographic GWs can generate a rather broad spectrum (Fig. 8n). The peaks of the variance in certain cases are close to the nadir (e.g., Fig. 8o,p), which means those Doppler-shifted GWs propagate almost vertically. By comparing the 2.5 hPa FOV curves with the ones at 80 hPa, we may conclude that background wind plays a leading role in shaping the GW spectrum during the upward propagation. However, three preferred angles at FOV No. 25, No. 0 and No. 20 persist from lower to upper altitudes over the Rockies during January 2005, and the FOV No. 0 and No. 20 also have peaks of GW variance at 2.5 hPa during January 2008, even though no big wave event occurred over the Rockies in January 2008. Similar phenomena happen over Iceland (Fig. 8f,j), and the Andes (not shown). These preferred directions should be associated with the detailed structures of the topography, and background wind is probably another candidate that prefers to select certain angles.

GWs seen by AIRS provide at least two new discoveries about the orographic and convective GWs. For orographic GWs, vertical wavelength is already large at the wave generation level compared with convective GWs, and certain angles are favored for generating large amplitude GWs, which vary among different mountain ranges and vertical wind structures as well. The ratio between peaks and troughs as well as maximum amplitude occurring FOV angles can be used to infer wave intermittency, which is an important wave parameter in GCMs that lacks observational constraints. It is not clear why certain angles are preferred. The reason why large amplitude GWs can be seen in the nadir view is another interesting point and beyond the scope of this study. They remain as potential topics for future investigations.

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Convectively generated GWs, as suggested here in AIRS observations, have much smaller wave scales in terms of both vertical and horizontal wavelengths. Therefore, it is expected to have fair amount of momentum flux in the high-frequency end of the momentum spectrum, which also varies quite a lot based on sporadic spikes present in AIRS. The current assumptions of source spectrum (e.g., Song and Chun, 2005; Beres et al., 2004) for convective GWs is, at least, not accurate enough to represent all situations. Note that these curves are all derived from monthly means. Spikes in both mountain and convection cases may simply come from different events. Single case studies need to be carried out in the future with sophisticated model runs to give more definitive conclusions.

4.2 Interannual variations at the equatorial region

The quasi-biennual oscillation (QBO) acts as an efficient GW filter in the tropical lower-middle stratosphere, and the interations between GWs and background winds are, in turn, an essential ingredient in the QBO. Different GWs contribute differently to driving the QBO. Kawatani et al. (2010a,b) were able to simulate a realistic QBO with a high-resolution climate model without parameterizing GWD. Their GCM can resolve part of the inertia GW spectrum (limited by the model's vertical grid resolution), and the simulated QBO has a shorter period than that of the real atmosphere, but with realistic strength and structure. Their work suggests the importance of inertia GWs in generating the QBO. Some other model studies suggest that the momentum deposition associated with the filtered GWs accounts for at least half of the total momentum flux in driving the QBO (Giorgetta et al., 2002; Kawatani et al., 2010a). QBO signals have also been seen in various GW observations, such as radiosonde (Vincent and Alexander, 2000; Wang and Geller, 2003), GPS (Torre et al., 2006) and Aura MLS (Wu and Eckermann, 2008). These instruments are ideal for seeing low to middle frequency GWs.

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AIRS GWs contain a strong annual cycle in the tropical lower stratosphere, as shown in Fig. 7. On top of the annual cycle, a clear QBO signal stands out. After removing the annual cycle, the linear trend, and sub-seasonal oscillations, we present the timeseries of the GW variance/difference anomalies at the equator in Fig. 9. Apparent GW variance enhancements occur along with the descent of the QBO easterly phase (defined as when the monthly zonal wind is easterly), and a reduction of GW activity happens right before and lasts through the whole period of the descending of the westerly phase (Fig. 9a). Year 2003–2004 is somewhat different from other years, when the enhancement lasts for a very short time, and a strong negative anomaly continues throughout the descending of the QBO easterly shear. The GW differences also show preferred zonal propagation directions during different QBO phases. Although it is more ambiguous in Fig. 9b, in general the GWs that can survive the wind perfer to propagate eastward (westward) in the presence of QBO easterly (westerly) shears. Again, year 2003–2004 is an exception.

The phenomena of enhanced (reduced) GW variance in the QBO easterly (westerly) phase can be explained as follows. Deep organized convection is the most dominant source at the tropics that is responsible for GWs we see in AIRS, as discussed in the last section. Since most of the deep convective events in the equatorial region move eastward (Wang and Rui, 1990; e.g., convections embedded in Madden-Julian Oscillations during boreal winters and Intra-Seasonal Oscillations during boreal summers), they tend to generate eastward propagating GWs. In the QBO easterly phase, the eastward propagating waves can survive, while they are mostly filtered out in the QBO westerly phase. This leads to the observed appearance that more GWs survive in QBO easterly phase. This result is consistent with what has been observed by Aura MLS (Wu and Eckermann, 2006), where suppressions (enhancements) of wave activities are observed in the QBO westerly (easterly) phase in the equatorial lower stratosphere. It suggests that the GWs observed by AIRS may play a more important role for the descent of QBO westerly phase than that of the easterly phase as more AIRS GWs are removed and hence deposit their momentum fluxes in the QBO westerly

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phase. However, the AIRS variance explained by the QBO is about 10 times smaller than that explained by the annual cycle, which means the GW variance is only slightly modified by the QBO rather than playing a dominant role on the formation/propagation of QBO phases. This is expected since GWs seen by AIRS are mostly high frequency waves that are usually with fast vertical group velocity. They usually break in the upper stratosphere and above, and significantly modulate the mean wind there (Alexander and Fritts, 2003).

Some of the previous studies using radiosonde and GPS also found suppression of GWs in the QBO westerly phase (e.g., Sato and Dunkerton, 1997; Wang and Geller, 2003; Torre et al., 2006), but they suggest that GWs are significantly enhanced in the QBO's descending westerly shear zone. Since Kelvin waves are unavoidably blended in the measurements from radiosonde and GPS, they are a major contributor, and help to bring the westerly phase down because their momentum flux is eastward. MLS scans along-track (almost meridionally over most of the latitudes), and Kelvin waves are truncated by a technique with short horizontal wavelength cut-off. 7-pt running smooth window we applied on AIRS measurements produces a cut-off wavelength at ~ 105 km, which can also effectively filter out Kelvin waves. Another possible reason lies in the fact that radiosonde and GPS are likely observing different parts of the GW spectrum from AIRS. Vincent and Alexander (2000) found increased GW activities during QBO easterly phase through radiosonde data over a tropical island, which is a bit different from other radiosonde observations but is consistent with what we observe in AIRS. The reason for the strong negative anomaly during year 2003–2004 remains unclear. It might be associated with other interannual variabilities such as El Nino Sothern Oscillation (ENSO).

Theoretical work predicts both westward and eastward propagating GWs should both contribute to the QBO formation (Baldwin et al., 2001). This is further confirmed by simulation results from high-resolution GCMs (e.g., Sato et al., 1999; Watakani et al., 2010a,b). However, the GWs with eastward phase speed are hardly observed through various GW measurement techniques (e.g., Sato et al., 1997; Wheeler et al., 2000) or

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it is difficult to separate these GWs from massive Kelvin wave signals as stated in the last paragraph. AIRS observes the tilt of wave phases in the zonal-height plane, from which the zonal direction of the intrinsic phase propagation can be inferred, assuming that the group propagation is upward. The inferred preferred zonal wave propagation direction from AIRS at the equator show both eastward and westward propagation signals with about equal probability to occur (Fig. 9b). In particular, since GWs with phase speed of the same sign with the zonal wind tend to meet the "critical level" more easily, westward (eastward) propagating GWs are more abundant in QBO westerly (easterly) shear. This is the first time the eastward propagating GWs are seen in satellite GW observations. It indicates the powerfulness and uniqueness of satellite instruments in obtaining the GW information.

4.3 Wave enhancement in the equatorial lower stratosphere and near the winter pole at 10 hPa

Mountain GWs are excessively strong in the AIRS observations, which overwhelm other GW components. In the lower stratosphere, however, GWs are stronger in the equatorial region. These GWs are not closely associated with deep convetion, as the variance forms a uniform belt (Fig. 5h,i, Fig. 6h,i, and bottom row of Fig. 7). By revisiting Fig. 3, we can find that the amplitude of these GWs maximizes at the tropopause ($\sim$ 100 hPa), decreases dramatically right above, and increases again above 80 hPa. These features are in good agreement with previous studies using radiosonde (e.g., Wang and Geller, 2003), GPS (e.g., Ratnam et al., 2004), Aura MLS (Wu and Eckermann, 2008), CRISTA and SABER (Preusse et al., 2006), and high-resolution GCM simulations (e.g., Sato et al., 1999), but not shown in UARS MLS (Wu and Waters, 1996) nor Aura AMSU-A (Wu, 2004). Meanwhile, there is also a peak at the winter pole at 10 hPa, as shown in Figs. 3 and 4. This is the first observation of such a 10 hPa winter-pole-enhancement. Since this feature is not uniformly distributed among the high-latitude hemisphere winters (Figs. 5 and 6), and the locations vary from year to year (not shown), it is believed not purely caused by underestimation of the instrument noise, but rather a real atmospheric phenomenon.

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A careful evaluation of FOV-dependent variance profile ("FOV curve" hereafter) suggests that the 10 hPa variance is something other than high-frequency internal GWs and instrumental noise. The zonal mean FOV curves derived at 60° N, 0° and 60° S for January, 2005 at 80 hPa are plotted in Fig. 10. At 60° N, as we would expect to see from topographically generated internal GWs in Northern Hemisphere winter, the variance is the largest at the eastmost view. At 60° S, the FOV curves are basically flat and small, close to the estimated noise level. At the equator, the FOV curve bends downward, with maximum value occurring at the nadir view. This is also true at the two flanks of the polar night jet at middle to upper stratosphere (not shown).

Turbulence can cause "bell" shape FOV curves seen in Fig. 10b due to the fact that the power index for 2-D turbulence is more negative at the off-nadir view than the nadir view and hence the area integral is smaller for the off-nadir view (Gruninger et al.,1998). Since shear instabilities easily occur at the two flanks of the polar night jet, turbulence can be particularly strong there. However, we cannot explain why turbulence is particularly strong in the equatorial lower stratosphere and the winter pole at 10 hPa with this theory. Besides, according to Aumann and Miller (1994), the most off-nadir view has a FOV swath area 4 times larger than that of the nadir view, which could result in a deeply downward curved FOV-dependent variance profile. Even with noise being already subtracted, the FOV curve in Fig. 10b has a difference between off-nadir and nadir views of only 7%, which is way too small to be purely interpreted as the turbulence.

Another explanation of the downward curvature of FOV curves in Fig. 10b invokes low-frequency inertia GWs, and we will show that AIRS can in fact observe them. Alexander et al. (2002) proposed the idea that slow GWs have larger probabilities to be observed than fast waves. Since the vertical group velocity is approximately proportional to square of the wave intrinsic frequency, AIRS has a larger chance to observe inertia GWs in the lower stratosphere, and the overall enhancement of wave activity observed in the tropics is largely due to the reason that these inertia GWs propagate slantwise and spread out from the source consequently. Since the GW enhancement

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in the equatorial lower stratosphere observed by AIRS is similar to GWs revealed by radiosonde and GPS as well as simulation results in high-resolution GCMs (e.g., Fig. 8c in Sato et al., 1999), we have more confidence to claim that AIRS observes inertia GW signals.

Here we conduct a sensitivity simulation to show how AIRS variance has a good sensitivity to inertia GWs in the lower stratosphere. AIRS's vertical WFs have a thickness of about 12 km, which leads to a strong response to long vertical wavelength waves (Alexander and Barnet, 2007). The WF is practically a delta function with respect to these long vertical wavelength GWs and therefore yield GW variance at the level where the WF peaks. For waves with vertical wavelength less than 12 km, the convolution of AIRS WF with waves of a constant amplitude will yield a very small variance. However, if the wave amplitude varies with height, the GWs at the edge of the WF cannot be completely cancelled, and will yield some response in AIRS variance. Figure 11a gives an ideal WF at nadir view at 80 hPa to represent mean WFs of multiple AIRS channels we select which all peak at 80 hPa (Appendix A). The WF at the outmost off-nadir view covers an area 4 times broader than that of the nadir WF (see Appendix B for details). Figure 11b shows the GW that is a result of the original imported GW convolved with the WF. The original GW has an amplitude varying with N^2, and the maximum amplitude A_{max} is 5 K (see Appendix B for details).

The final goal of this experiment is to try to find the best wave parameters that give the convolved variance at the nadir and the variance difference between the nadir and outmost views consistent with the observations (black line in Fig. 10b). In order to compute the variance, we make a horizontal phase shift of $n\pi$ forward of the convolved wave, as shown in shades of Fig. 11b. This procedure is to account for the 7-pt smoothing window we applied to AIRS data. The difference between the two convolved waves are then calculated and compared with the observation (Fig. 10b) at the nadir view. With a fixed λ_x of 700 km, 7-pt smoothing window length ($\sim$ 105 km) corresponds to 0.15π phase shift. We can therefore plot out the expected AIRS variance as a function of vertical wavelength λ_z in Fig. 12a. The same procedure can be applied to the

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extreme off-nadir point to obtain a reading of wave variance there, and the difference between the nadir and outmost views can also be computed as shown in Fig. 12b. The two curves meet with the observed values both at $\lambda_z = 6$ km. We tried other λ_x values and it turns out $\lambda_x = 700$ km, $\lambda_z = 5$ km and $A_{max} = 5$ K gives the best agreements simultaneously. Typical horizontal and vertical wavelengths for equatorial inertia GWs are around 1000 km and 5 km, respectively (Wang and Geller, 2003), and the peak amplitude is around 5 K (Sato et al., 1999), which is in good agreement with our simulation result. The inertia GW variance can also explain no obvious propagation direction in the difference between eastmost and westmost views, as the radiance response involves no propagation direction information whatsoever in this case. The layer near 80 hPa (Fig. 11a) is less affected by the QBO wind than the layers above and hence we still do not see a dominant QBO signal even these waves are believed to be dominantly inertia GWs. The simulation with a realistic AIRS WF supports the idea that the enhanced GW variance at 10 hPa winter pole is also low-frequency inertia-gravity waves. See Appendix B for details about this experiment.

The inertia GWs at the equatorial lower stratosphere cannot propagate off-equator further to a high altitude since their paths are too slantwise, and they easily meet the lower limit of frequency, which is the inertia frequency f, as f increases with latitude (Sato et al., 1999). Hence the tropical belt of GW enhancement disappears at high altitudes (20 hPa and above), and only those high-frequency internal GWs that are closely related with convective sources can survive. Choi et al. (2011) applied 3-D AIRS WFs to their parameterized convective GWs, and found out that merely any GWs can be seen in the equatorial lower stratosphere since the parameterized GWs are too small in terms of the vertical wavelength and horizontal wavelength as well. Propagating upward, these small GWs are Doppler-shifted toward longer vertical wavelength, and become detectable by AIRS, which agree very well with what we see at middle to upper stratosphere.

A peak at 10 hPa at the winter pole appears in the N^2 field in Fig. 1 of the Appendix B. Therefore, as in the tropical lower stratosphere, inertia GWs should grow again at those levels, but with a short penetration. Convolving with the weighting function, and the GWs result in another enhancement in AIRS variance. Due to the sensitivity limitations, radiosonde and GPS sensors cannot observe those GWs at this level. UARS MLS and Aura AMSU-A cannot see these features probably because of the coarse horizontal resolutions (Wu and Eckermann, 2008). Aura MLS also observes similar equatorial enhancement at lower levels, but not at the 10 hPa winter pole, which remains to be investigated in the future (Wu and Eckermann, 2008).

After all, the AIRS GW variance is believed to be composed of large amplitude internal GWs that are closely related with the local sources (e.g., orography, deep convections), and small amplitude "background-like" inertia GWs that are determined by the shape of the WFs and the shape of the original amplitude spectra. It is shown that AIRS can observe the inertia GWs even though their vertical wavelengths are less than AIRS WF thickness. This interpretation might be applicable to explaining signals discovered in other instrument measurements. The two groups of GWs are potentially separable by regressing the variance to N^2. Turbulence is another possibility, but the strength and mechanism remain as a question for future investigation.

5 Conclusions

In this study, GW induced temperature variance is succesfully extracted from the brightness temperature variance observed by Aqua AIRS by carefully removing instrumental noises at various pressure levels. A total of 50 AIRS channels is used with WFs peaking at pressure levels between 2 hPa and 100 hPa, covering GWs in the entire stratosphere. Because of the nadir viewing geometry and small footprint size, AIRS radiance variance is mostly sensitive to high-frequency internal GWs with comparable vertical and horizontal wavelengths with wave fronts in parallel to the scanning angle. Large smearing occurs when the wave front is across the AIRS scan, leading to possibility of inferring the preferred GW propagation direction using variance difference between two outmost views. Considering the fact that the scan is perpendicular to the orbit

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paths and AIRS is in a polar orbit, the inferred wave propagation direction is mostly in the zonal direction. Climatology of the GW variance and zonal preferred propagation direction are documented in this study.

Comparisons have been carried out throughout the paper between AIRS GW variance and other measurements including satellite and non-satellite instruments (e.g., Aura MLS, Aqua AMSU-A, GPS, radiosonde, GCM simulations, etc.). The GW variance seen by AIRS in general agrees with previous studies in terms of the distributions and variations of maxima. The AIRS GW peaks are generally found to be above high meridionally-oriented mountain ranges, as well as near tropical deep convection regions. The GW amplitudes grow exponentially with height due to the decrease in air density. Different from others, the GWs observed in AIRS are highly localized, since they are dominated by high-frequency components. This provides better information on GW sources from direct retrievals of GWs, as shown in Alexander and Barnet (2007). The locations of GW peaks vary significantly from year to year. The 700 hPa wind is used to evaluate the GW generation, showing that stratospheric GWs tend to grow into great amplitudes at the places with wind speed greater than $10\,\mathrm{m\,s^{-1}}$ collocated with the upper-level jet maxima. In the tropics, the deep convection controls the GW variance peaks.

The inferred preferred zonal propagation directions of GWs seen in AIRS are opposite to the mean zonal wind directions. Topographic GWs in the winter high-latitudes tend to propagate westward relative to the mean flow, while convectively excited GWs in the tropics and subtropics tend to propagate eastward relative to the mean winds. The preferred wave propagation becomes clearer with increasing altitude, suggesting more wave filtering and increasing GWD on the mean flow. The GW variance contours do not exactly follow the wind contours, suggesting less filtering of the high-frequency GWs. Since the high-frequency GWs tend to have larger horizontal phase speeds, there are fewer that meet the critical levels in the lower atmosphere compared to lower-frequency inertia GWs. Therefore, they become more important in the mesosphere and above.

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An analogous approach of determining wave propagation direction has also been carried out on Aura MLS observations (Wu and Eckermann, 2008), where wave preferred meridional propagation direction is inferred in a similar way considering the fact that MLS scans along the orbit. One can therefore obtain a more complete picture of the GW horizontal propagation direction from the combined AIRS and MLS results. Recent studies show that convective GWD parameterizations are sensitive to the wave propagation direction, which is set to be a free parameter in the model (Song and Chun, 2005, 2008). Our results from AIRS will certainly contribute to constraining GWD parameterization schemes in GCMs.

The new findings with AIRS GW variance differ from other GW observations in three aspects:

Firstly, AIRS 90-footprint FOV-depedent GW variance above mountain ranges and deep convection occurs with preferred angles. Because of the AIRS steep viewing angle, this finding indicates that really high-frequency GWs with large amplitudes are present, and mainly excited from orographic and convective sources. It requires further investigation on why certain angles are preferred, which is a parameter missing in all GWD parameterization schemes.

Secondly, the QBO signals observed in AIRS are different from other measurements that mainly observe inertia GWs and Kelvin waves. With relatively fast vertical group velocity, these high-frequency internal GWs seen in AIRS are more difficult to be filtered out by QBO winds, and hence they are less important in the formation and descending of QBO phases. As more of a direct response to the wind filtering effect, AIRS GWs tend to propagate in opposite directions to QBO winds. Since the AIRS GWs are slightly suppressed during the QBO westerly phases, the AIRS GWs might be more important to QBO westerly phase than to the easterly phase. The QBO has not been well simulated so far in GCMs. On one hand, the coarse vertical resolution results in deficiency in resolving some inertia GWs. On the other hand, the poorly observationally constrained GWD parameterization schemes cannot precisely represent the characteristics of high-frequency GWs. AIRS not only provides abundant information

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on the latter on to improve our understanding of high-frequency GWs, but also show potential to observe low-frequency inertia GWs.

Lastly, the most important and encouraging finding from this work is that AIRS can observe GWs with vertical wavelength smaller than the thickness of the WFs. By assuming inertia GWs' amplitudes varying only with background stabilities, the variance from the inertia GWs is measurable by AIRS. AIRS appears to be highly selective on high-frequency GWs, and therefore we literally can separate the GW variance between the inertia-gravity and high-frequency components. Moreover, this new capability will help us to further understand GW characteristics as observed by other sensors.

This paper provides the first climatological report of GW characteristics from nadir viewing sounders. Since AIRS is sensitive to high-frequency GWs, the results from this study also provide the first global survey of the characteristics and properties of high-frequency GWs. Some large-amplitude GW events as well as FOV-dependent variance can be studied in further detail with models. Improving the GWD parameterizations with the AIRS GW observations is another important task to explore in future study.

Appendix A

Channel numbers and the estimated noise level

The channel numbers we used in this paper are listed below in Table A1. Together we also give the estimated noise from "3-pt running average window" method and the provided instrument NEdT on 30 August 2002. The minimum detectable GW variances for zonal means and monthly maps are listed in the last two columns. Bold numbers in the 2nd column are channels that locate at the wings of the radiance spectrum and have no paired channels. These channels are particularly sensitive to small drift in the wavelength while others are stable.

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Appendix B

Some details about the simulation

In the simulation mentioned in Sect. 4.3, we assume the imported wave amplitude varies with N. The reason is as follows. Eckermann (1995) derived temperature variance as:

$$\overline{\hat{T}'^2} = \frac{N^2}{g^2}\left[1-(f/\omega)^2\right]\left(\overline{u'^2}+\overline{v'^2}\right) \tag{B1}$$

where g, u', and v' are the gravitional acceleration, zonal wind perturbation, and meridional wind perturbation caused by GWs, respectively. $\overline{\hat{T}'^2} = \overline{T'^2}/T_0^2$, where T' is the absolute temperature perturbation, while T_0 is the background temperature. The overbar denote the column average.

Think of it physically. Larger N^2 corresponds to more stable atmosphere, which means the potential temperature increases more rapidly with height. Therefore, same oscillations in height would cause larger temperature perturbations. By assuming that waves neither break nor meet saturations, the vertical group velocity varies inversely with N^2, so that the wave activity coming up from below accumulates in the lowermost tropical stratosphere even if the waves are initially conservative there.

The zonal mean values of N^2 at January, 2005 are shown in Fig. 1. The input 2-D wave amplitude A is assumed to be proportional to N at the equator, which is approximated as:

$$A(x,z) = \begin{cases} A_0\cos\left(\frac{2\pi z}{\lambda_z}+\frac{2\pi x}{\lambda_x}\right)e^{\frac{-(z-z_{60})}{\lambda_z}}, & \text{if } z \geq z_{60} \\ A_0\cos\left(\frac{2\pi z}{\lambda_z}+\frac{2\pi x}{\lambda_x}\right)e^{\frac{-|z-z_{60}|}{4\lambda_z}}, & \text{if } z < z_{60} \end{cases} \tag{B2}$$

where $z_{60} \sim 21.5$ km roughly corresponds to the geometry height at 60 hPa. x is the along-track distance, and z is the height.

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The weighting function $\mathrm{WF}(x,y,z)$ for AIRS can be calculated directly from radiative transfer models (e.g., Eckermann et al., 2007). Here we use a 2-D function combined with two Gaussian type functions to approximate the actual weighting function. $wf(x,z)$ at the nadir at 80 hPa is defined as:

$$wf(x,z)=\begin{cases} e^{\left(\frac{-(z-z_{80})}{\sqrt{2}\times 6.1}\right)^2} e^{\left(\frac{x}{\sqrt{2}\times 6.75}\right)^2}, & \text{if } z \geq z_{80} \\ e^{\left(\frac{-(z-z_{80})}{\sqrt{2}\times 1.5}\right)^2} e^{\left(\frac{x}{\sqrt{2}\times 6.75}\right)^2}, & \text{if } z < z_{80} \end{cases} \tag{B3}$$

and

$$\mathrm{WF}(x,z)=\frac{wf(x,z)}{\iint wf(x,z)\mathrm{d}x\mathrm{d}z} \tag{B4}$$

An example of this weighting function has been included in Fig. 11a. At the outmost off-nadir view, the width of this weighting function on along-track direction is broadened by a factor of 5.

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Table A1. A list of channel numbers, noise, NEdT, and minimum detectable GW variance at each pressure level. Bold numbers in the 2nd column are channels that are unpaired. See context for details.

Pressure (hPa)	Channel numbers	Noise (K^2)	NEdT (K^2)	Min. detectable GW var. ($\times 10^{-3}$ K^2) Zonal mean	Map
2	74	0.149	0.165	3.78	26.64
2.5	75	0.147	0.166	3.72	26.22
3	76	0.143	0.161	3.63	25.55
4	77	0.145	0.160	3.66	25.80
7	**78**	0.153	0.162	3.88	27.34
10	**79**	0.182	0.172	4.62	32.53
20	81, 82	0.084	0.078	2.14	15.05
30	**102, 108, 114, 120**, 125, 126	0.039	0.029	0.98	6.88
40	64, 88, 90, **94, 100**, 106, 118	0.033	0.028	0.83	5.86
60	66, 68, 70, 86, 87, 91, 93, **97**, 130	0.026	0.018	0.66	4.68
80	92, 98, 104, 105, 110, 111, **116**, 117, 122, 123, 128, 129, 134, 140	0.020	0.011	0.50	3.54
100	132, 133, 138, 139, 149, 152	0.026	0.014	0.67	4.73

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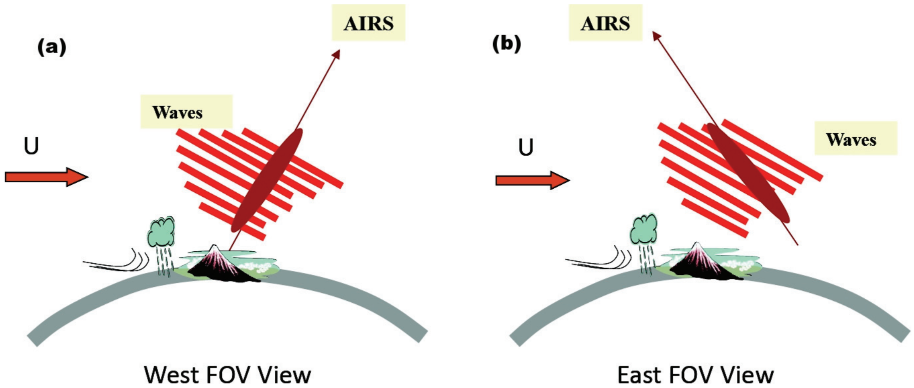

Fig. 1. Cartoons showing the wave smearing effect from different views. **(a)** Westmost off-nadir view, where in this particular case wave signals are largely smeared out; **(b)** Eastmost off-nadir view, where the strongest variance should be observed. Red parallel rectangles represent the phase lines of the GW, while bold red arrows denote the mean wind direction.

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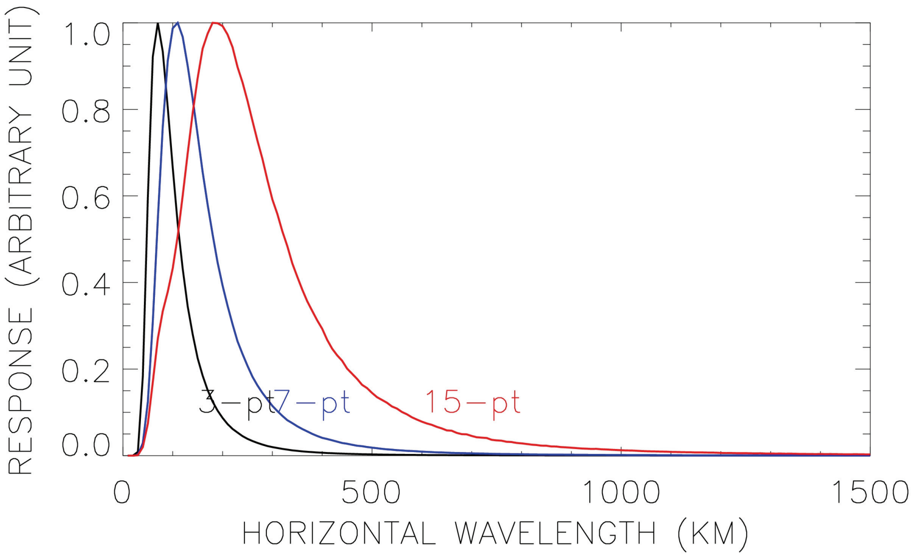

Fig. 2. AIRS visibility as a function of the along-track wavelength with 3-pt (black), 7-pt (blue) and 15-pt (red) running smooth window applied.

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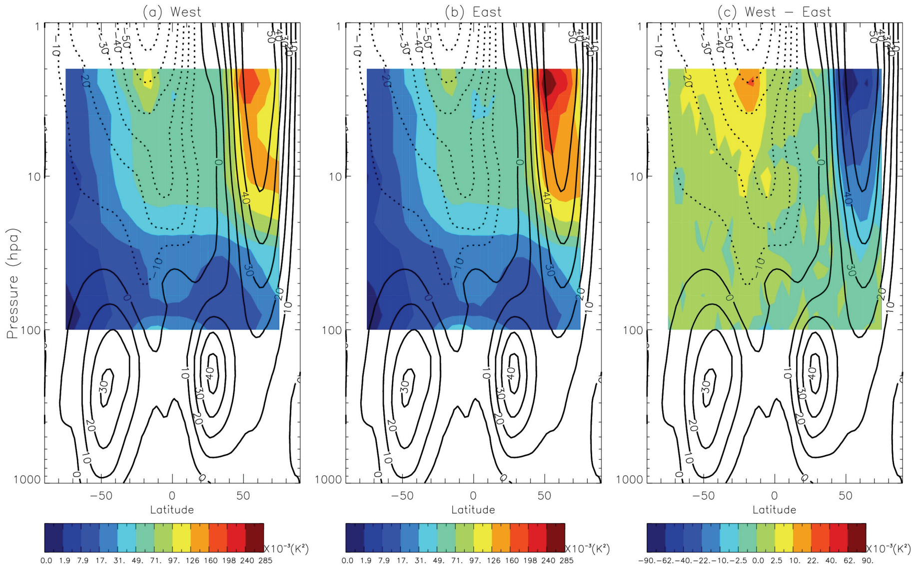

Fig. 3. Zonal mean of the GW variance at the westmost view **(a)**, eastmost view **(b)** and the difference between the westmost and eastmost views **(c)** as a function of latitude and height for January, 2005. Monthly mean zonal winds obtained from UK Met Office are contoured in solid (westerly) and dotted (easterly) lines with an interval of $10\,\mathrm{m\,s^{-1}}$. The color scale is linear by taking a square-root of the values. In **(c)**, positive (negative) values correspond to preferred eastward (westward) propagation direction.

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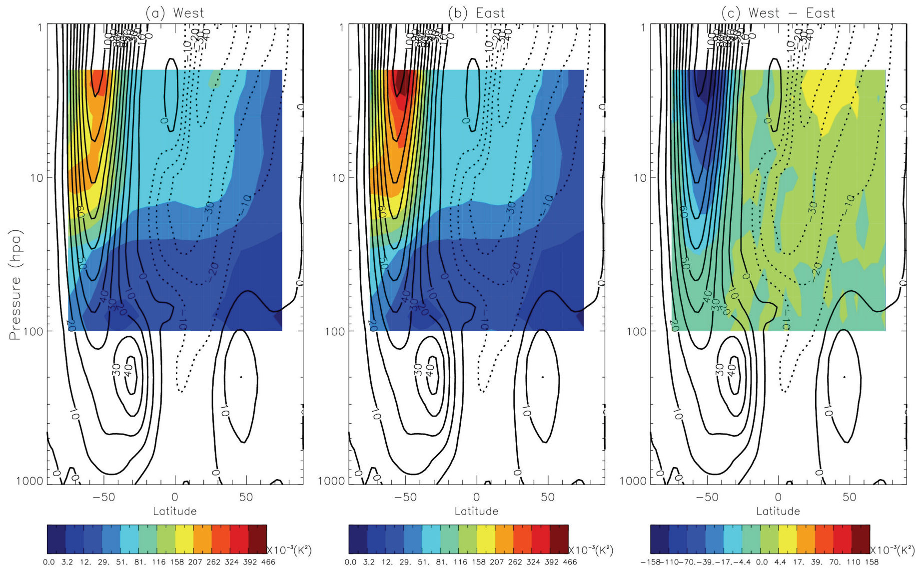

Fig. 4. Same with Fig. 3, except for July 2005.

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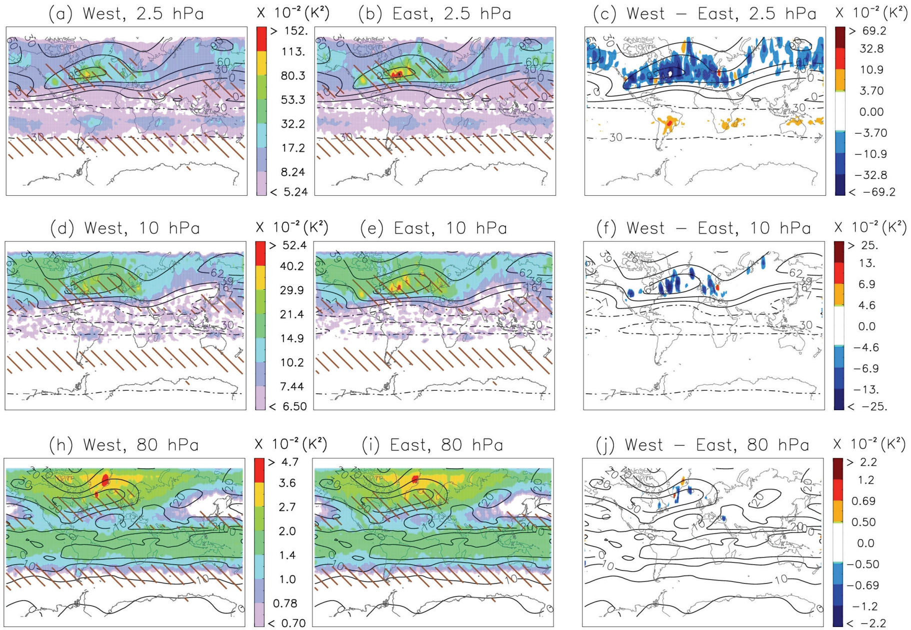

Fig. 5. Geographical distribution of GW variance on a $2^{\circ} \times 2^{\circ}$ grid at westmost view (left column), eastmost view (middle column) and the difference (right column) for January 2005 at 2.5, 10, and 80 hPa from top to bottom. 3-pt smoothing has been further applied to make the signals stand out. In the left two columns, values smaller than 2σ are uncolored, where σ is the minimum detectable variance. In the rightmost column, absolute values smaller than $\sqrt{2}\sigma$ are whitened, where $\sqrt{2}\sigma$ is the smallest detectable difference. The UK Met Office monthly mean zonal winds at corresponding levels are contoured in black solid (westward) and dotted (eastward) lines. Total wind velocity greater than $10\,\mathrm{m\,s^{-1}}$ areas at 700 hPa are hatched with brown lines.

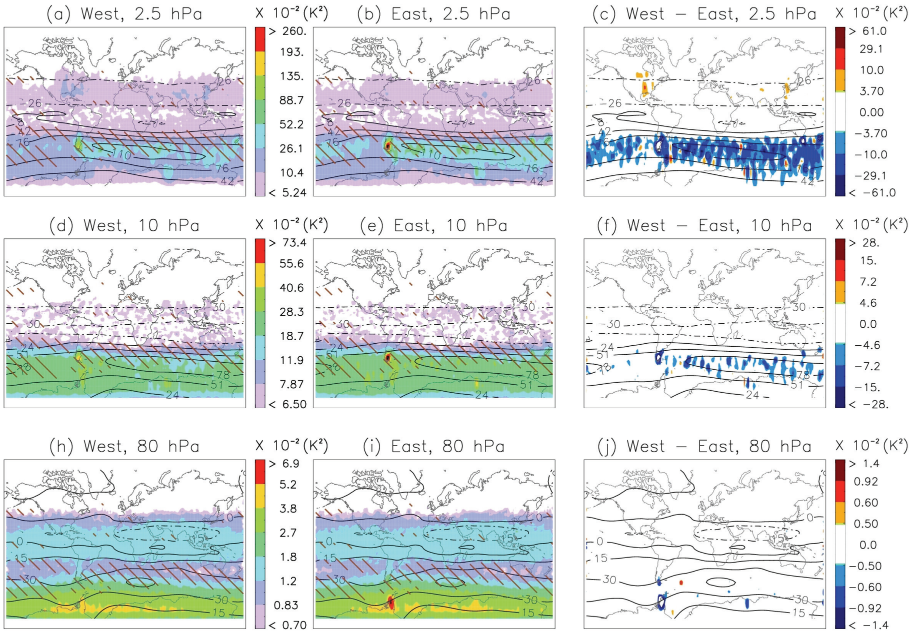

Fig. 6. Same as Fig. 5, except for July 2005.

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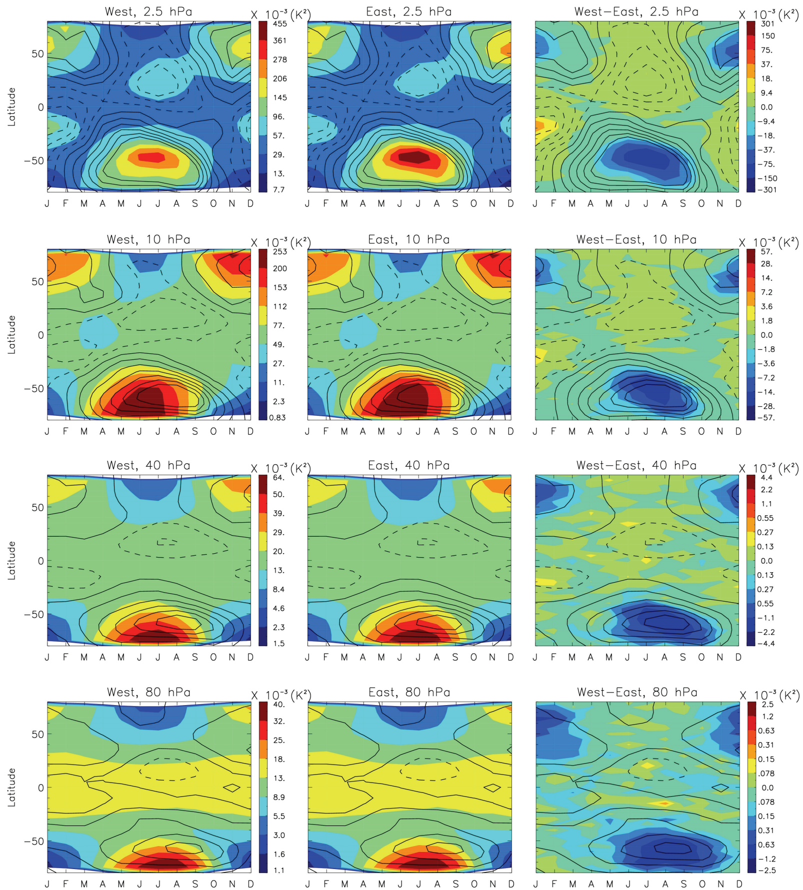

Fig. 7. Multi-year averaged timeseries of GW variance for different months at westmost view (first column), eastmost view (second column) and the difference (third column) at 2.5, 10, 40 and 80 hPa from top to bottom. The UK Met Office zonal winds are overplotted in solid (dashed) lines representing westerlies (easterlies) with contour interval of 10 m s^{-1}.

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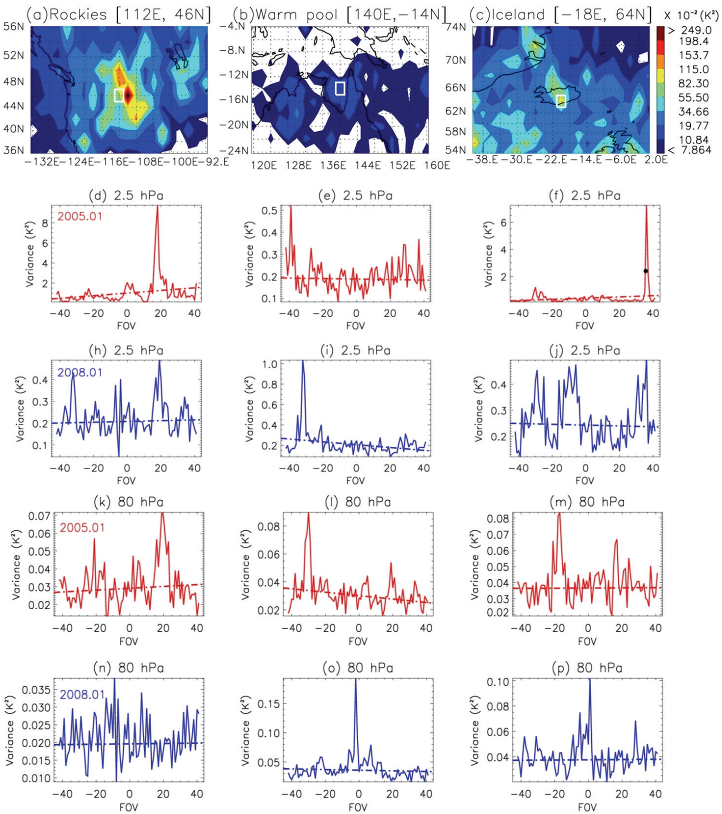

Fig. 8. (a–c) Geographical map of GW variance at eastmost view at three different locations for January 2005 at 2.5 hPa. The GW variances within the white box of **(a–c)** as a function of FOV numbers are plotted in solid for each location at **(d–f)** January 2005, 2.5 hPa; **(h–j)** January 2008, 2.5 hPa; **(k–m)** January 2005, 80 hPa; and **(n–p)** January 2008, 80 hPa. Negative (positive) FOV numbers correspond to west (east) views, and 0 corresponds to nadir view. The linear fittings of each FOV curve are overplotted in dash-dotted lines.

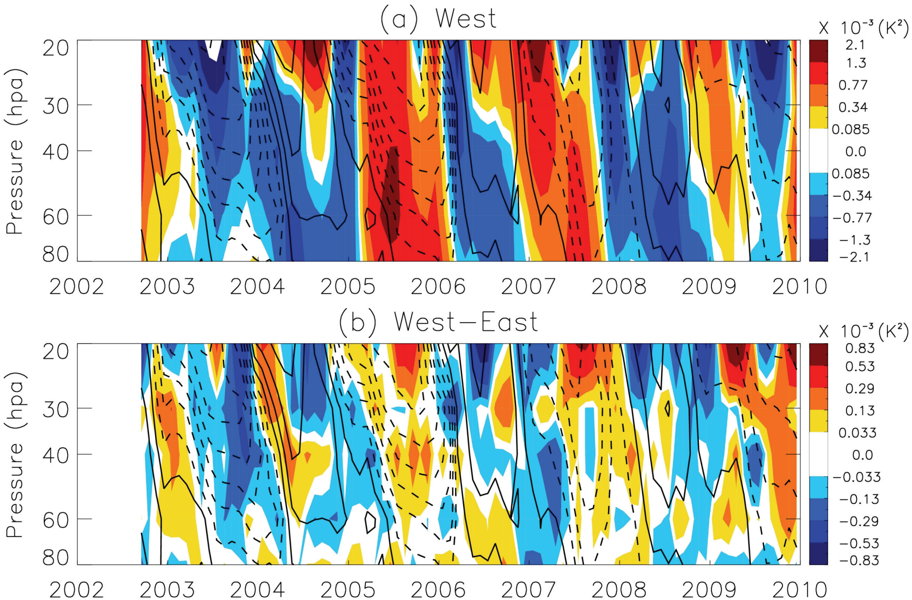

Fig. 9. Monthly mean timeseries (annual cycle and linear trend removed) of the zonally averaged GW variance taken at the westmost view **(a)** and the variance differences between the westmost and the eastmost view **(b)**. The vertical cross-section is taken at the equator from 80 hPa to 20 hPa. 3-month running smooth window is applied to the data to remove the sub-seasonal cycle. UKMO monthly mean zonal winds are overplotted in lines with contour intervals of $5\,\mathrm{m\,s^{-1}}$ and solid (dotted) lines indicating eastward (westward) winds. Variance within $\pm 0.085 \times 10^{-3}\,\mathrm{K}^2$ and difference within $\pm 0.033 \times 10^{-3}\,\mathrm{K}^2$ are uncolored.

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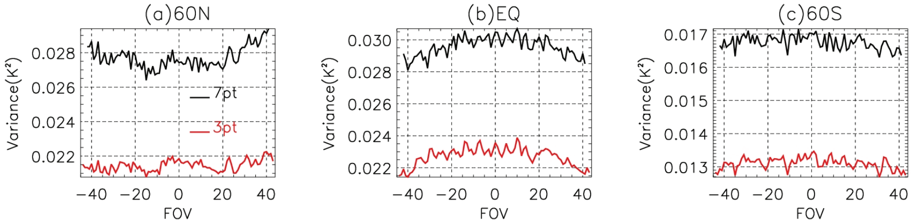

Fig. 10. Zonal mean FOV curves at 80 hPa for 3-pt (red) and 7-pt (black) results at January 2005 at 60° N **(a)**, 0° **(b)** and 60° S **(c)**. The horizontal axes are the number of scanning points along a FOV scan, where −(+)45 corresponds to westmost(eastmost) view, while 0 is at nadir.

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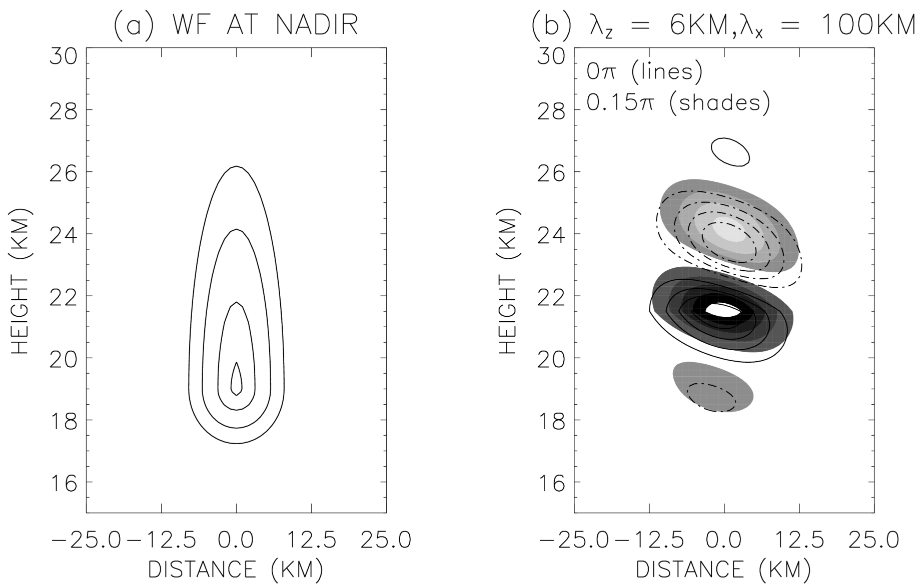

Fig. 11. **(a)** Weighting function over nadir at 80 hPa as a function of height and along-track distance for the numerical simulation. The contours are 0.5, 0.7, 0.9 and 0.99 of the peak response. The outmost off-nadir weighting function is 4 times broader than that of the nadir. **(b)** One example of the original imported wave convolved with the nadir weighting function (lines) and resultant wave with 0.15π phase shift (shades). The wave has horizontal wavelength (λ_x) of 100 km and vertical wavelength (λ_z) of 6 km. Negative values are contoured with dash-dot lines (light colors), and positive values are denoted with solid lines (dark colors). The contour interval is 0.2 of the maximum and minimum amplitudes, respectively.

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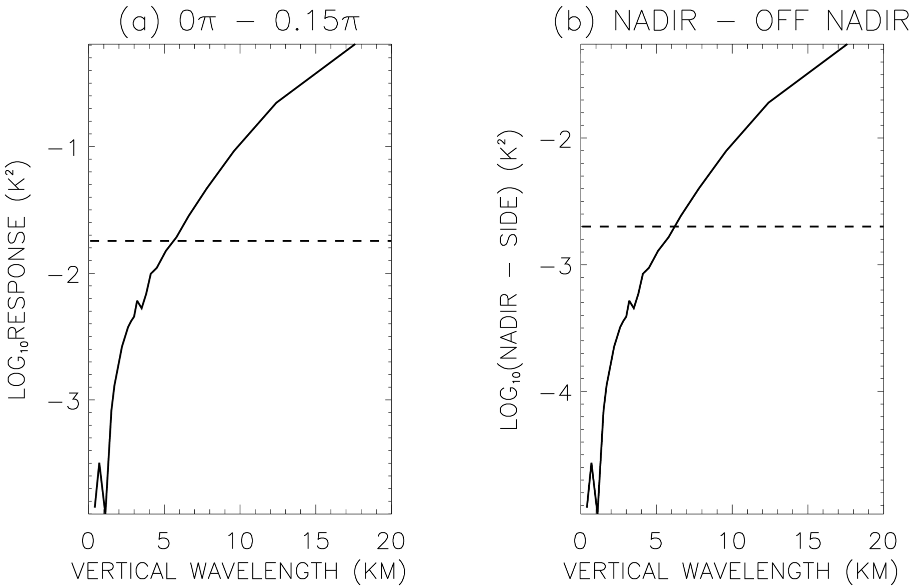

Fig. 12. Variance differences between unshifted wave and the wave shifted by 0.15π for λ_x = 700 km calculation **(a)**, and the variance difference between nadir FOV and the outmost FOV **(b)** as functions of λ_z. The AIRS observed values (7-pt smoothing window applied) are denoted by dash lines. See text for details.

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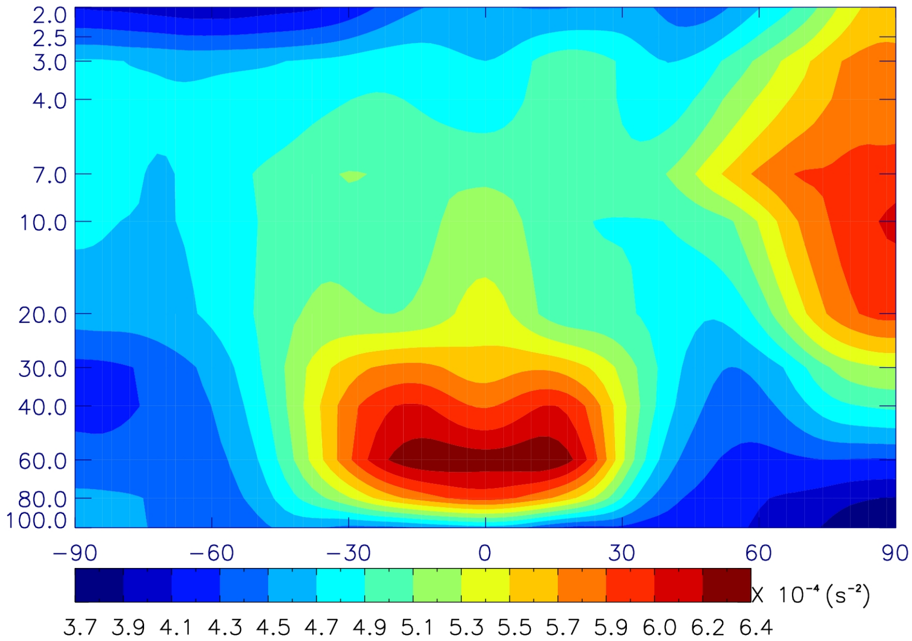

Fig. B1. Zonal mean of the Brunt-Väisälä frequency (N^2) as a function of latitude and height for January 2005, derived from ERA Interim dataset.

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