November 29, 2011

Gravity wave characteristics from OH airglow imager over Maui

Alan Z Liu

Xian Lu

Zhenhua Li

Gary R Swenson, *University of Illinois at Urbana-Champaign*

Steven Franke, *University of Illinois at Urbana-Champaign*

Gravity wave characteristics from OH airglow imager over Maui

Zhenhua Li,¹ Alan Z. Liu,² Xian Lu,¹ Gary R. Swenson,³ and Steven J. Franke³

Received 27 February 2011; revised 12 September 2011; accepted 12 September 2011; published 29 November 2011.

[1] The gravity wave (GW) characteristics over Maui from 2002 to 2007 are deduced from OH airglow imager. The dominant horizontal wavelength falls between 15 and 30 km. Intrinsic phase speeds are mainly in the range of 50–100 m s⁻¹. Observed phase speeds are mainly in the range of 20–70 m s⁻¹. Intrinsic wave periods are mostly below 30 min with the majority clusters around 5–15 min. For intrinsic wave period larger than 5 min, as wave period increases, the number of wave events decreases. Observed wave periods fall in the range of 8–20 min. The vertical wavelength distribution peaks between 20 and 30 km and falls rapidly as wavelength decreases. GWs propagating against (along) background wind are Doppler shifted to a higher (lower) frequency, larger (smaller) vertical wavelength, and decreased (enhanced) chance of observation. During all seasons Doppler shifting is part of the explanation as to why the majority of observed GWs propagate against the background wind. The preferentially poleward or equatorward propagation of GWs during summer or winter observed by airglow imagers at Maui can be explained by the effects of critical layer filtering, ducting, and Doppler shifting. In March, when critical layer filtering effects are not important, Doppler shifting and ducting play dominant roles in determining preferred zonal propagation direction. In December, ducting and wave source location predominantly determine wave propagation direction. Momentum flux deduced from OH imager is also highly anticorrelated with background wind on both seasonal and diurnal time frames. Momentum flux is highly modulated by tidal wind. Waves with shorter vertical wavelengths have much smaller power in GW spectra. OH imager may capture the majority of momentum flux of high-frequency GW spectra through the airglow layer.

Citation: Li, Z., A. Z. Liu, X. Lu, G. R. Swenson, and S. J. Franke (2011), Gravity wave characteristics from OH airglow imager over Maui, *J. Geophys. Res.*, 116, D22115, doi:10.1029/2011JD015870.

1. Introduction

[2] Gravity waves (GWs) play important roles in the middle atmosphere dynamics. GWs are mostly excited in the lower atmosphere and as they propagate upward they can break due to instabilities either due to the growth of wave amplitudes or encountering of critical layer where their phase speeds approach that of mean wind. When GWs break, the momentum they carry is deposited into the mean flow. Especially in the mesopause region, GW drag drives a meridional circulation that reverses the meridional temperature gradient forced by solar heating at solstices. Though the major roles GWs play in the mesospheric dynamics have long been discovered, due to their broad range of temporal and spatial scales, resolving the full spectrum of GW in general circulation model is not practical due to the com-

putational cost of extremely high resolution. Thus, their effects must be accounted for by parameterization in the numerical models as a tunable momentum forcing term. To simulate the middle atmosphere realistically, the quantification of GW momentum flux, propagation characteristics are needed for constraining GW parameterization schemes based on observation.

[3] Short-period GWs (wave period shorter than 1 h) make a major contribution to the GW momentum budgets at mesopause altitude [Fritts and Vincent, 1987]. Though this part of GW spectra is difficult to be observed by other instruments, they are observable through imaging of airglow layers in the Mesopause-Lower-Thermosphere (MLT) region. The OH airglow layer on average resides at 87 km altitude with thickness of about 10 km. Airglow imager detects the perturbation of intensity of airglow layer induced by GWs, in this case OH. OH airglow imager is capable of detecting high-frequency GWs with periods shorter than 1 h and vertical wavelengths larger than the mean depth of OH airglow layer (~10 km). GWs often appear in the OH airglow images as quasi-monochromatic (QM) waves due to wave dispersion and filtering process in the lower atmosphere, which makes them relatively easy to be identified through spectral analysis. Typical horizontal wavelengths of GWs observed by airglow imager in the MLT are between ~20 and ~100 km. Phase

¹Department of Atmospheric Sciences, University of Illinois at Urbana-Champaign, Urbana, Illinois, USA.

²Department of Physical Sciences, Embry-Riddle Aeronautical University, Daytona Beach, Florida, USA.

³Department of Electrical and Computer Engineering, University of Illinois at Urbana-Champaign, Urbana, Illinois, USA.

speeds are from $\sim 30 \text{ m s}^{-1}$ to $\sim 100 \text{ m s}^{-1}$. Intrinsic wave periods range from 5 min to several dozen minutes with the majority is around 10 min [Taylor *et al.*, 1997; Swenson *et al.*, 2000; Hecht *et al.*, 2001; Ejiri *et al.*, 2003; Medeiros *et al.*, 2003; Nakamura *et al.*, 1999, 2003; Tang *et al.*, 2005a; Pautet *et al.*, 2005; Nielsen *et al.*, 2009].

[4] GWs observed by imagers over midlatitude sites often show dominant eastward and poleward propagation during summer [Taylor *et al.*, 1993; Nakamura *et al.*, 1999; Walterscheid *et al.*, 1999; Hecht *et al.*, 2001; Ejiri *et al.*, 2003; Tang *et al.*, 2005a; Dou *et al.*, 2010]. Preferred equatorward propagation GWs during winter are also observed over many midlatitude sites [Hecht *et al.*, 2007; Ejiri *et al.*, 2003; Tang *et al.*, 2005a; Dou *et al.*, 2010] though less dominant as summer poleward propagation bias. Previous wave studies propose that the dominant zonal propagation directions are caused by critical layer filtering. Taylor *et al.* [1993] showed that the waves observed over Nederland, Colorado (40.0°N 105.6°W) from May to June 1988 were predominantly northward and eastward. Assuming tropospheric source and using CIRA-1986 zonal winds, they computed the “block diagram” which are polar plots indicate direction and speed ranges that GWs would be blocked by critical layer filtering of mean wind. The eastward dominance of wave propagation is explained by critical layer filtering by mean westward zonal wind below MLT. Stockwell and Lowe [2001a] analyzed OH images taken by a three-station network scanning radiometers of during the summer of 1998 and found the majority of the waves propagate to the northeast direction. In their accompanying paper [Stockwell and Lowe, 2001b], the authors further examined the mechanism affecting the directionality of wave propagation and concluded that critical layer filtering by the background zonal wind explains the dominant eastward propagation in the summer and westward propagation in the winter, whereas the tidal components of the meridional wind account for the northward tendency of wave propagation (and the corresponding relative absence of southward propagating waves). Medeiros *et al.* [2003] analyzed airglow images over Cachoeira Paulista (23°S, 45°W) from October 1998 to September 1999 and found GWs mainly propagate southeast during summer and northwest during the winter. Applying critical layer filtering theory to the observation as in the article by Taylor *et al.* [1993], they found remarkable agreement between seasonal variation of observed wave propagation direction and that of blocking by stratospheric winds.

[5] Other researches found wave ducting condition and source location together may also cause observed dominant propagation direction. Walterscheid *et al.* [1999] reported that during summer (winter) QM GWs are predominantly propagating poleward (equatorward) over Adelaide (35°S), Australia. They interpreted the QM waves observed as waves trapped in the lower thermospheric duct or between the ground and the layer of evanescence above the duct. The dominant meridional propagation direction arises from changing of remote summer or winter source locations for these ducted waves. Hecht *et al.* [2001] analyzed 15 months of airglow imaging observation of GWs over Urbana, Illinois (40°N, 80°W) and explained the preferred northward propagation around summer solstice by wave ducting by

both thermal structure and mean wind shear. Suzuki *et al.* [2004] found waves propagate both poleward and equatorward, while in summer almost all waves propagate poleward at Darwin, Australia (12.4°S, 131.0°E), and concluded it is caused by the same mechanism proposed by Walterscheid *et al.* [1999]. Ejiri *et al.* [2003] studied GWs over two midlatitude sites. The meridional propagation direction of GWs show not only seasonal variation but also latitudinal dependence. In summer, the meridional propagation are mainly poleward for both sites, whereas in winter they are only equatorward at Shigaraki (34.9°N) and both poleward and equatorward at Rikubetsu (43.5°N). Both ducting condition and critical layer filtering can be responsible for observed meridional wave propagation preference in the two sites.

[6] Other than critical layer filtering and wave ducting, GW sources variation alone are also an important factor to determine wave propagation direction on OH airglow images. Using more than 1 year of imager observation over Indonesia, Nakamura *et al.* [2003] concluded the seasonal variation of GWs propagation direction observed there were more likely related to tropospheric source variation noticing the mean zonal and meridional wind were both weak near the Equator. Medeiros *et al.* [2005] found the majority of the waves observed over several sites in Brazil near the Atlantic from close to the equator to 23°S propagated from the continent to the ocean. They suggested the active convection over the continent account for the dominant wave propagation direction from the land to the ocean.

[7] Another factor that might contribute to wave propagation direction preferences is Doppler shifting by the local background wind field. Modification of vertical wavelengths of GWs by mean wind can cause directional variation in wave occurrence frequency due to observational filtering and/or wave dissipation. This effect on observed wave propagation preference on airglow imager has not been investigated.

[8] The objective of this paper is to take advantage of the long-term archive of OH imager observations and meteor radar wind measurements at Maui to determine the GW characteristics and momentum flux in the MLT over Maui. On the basis of the analysis on wave characteristics, we further explore the relationship between the seasonal variation of wave propagation direction, momentum flux and background wind.

2. Data and Methodology

[9] University of Illinois at Urbana-Champaign (UIUC) OH airglow imager was situated in Maui (20.7°N, 156.3°W) from January 2002 to August 2007. Because of its high elevation of 3058 m, the site was seldom covered by cloud. This all-sky imager is similar to the airglow imaging system described by Haque and Swenson [1999]. It incorporates an all-sky lens with an Apogee 1024 × 1024 CCD camera to take images of the all sky field. A broadband filter (750.0–930.0 nm with a notch at 865.0 nm) is used to measure the mesospheric hydroxyl (OH) airglow emission centered at a mean altitude of 87 km. The exposure time is 60 sec and images are taken every 2 min. The images are binned to 512 × 512 pixels to increase the signal-to-noise ratio. Wind observations in the mesopause region at Maui were obtained hourly by UIUC Meteor Radar from May 2002 to June 2007 [Franke *et al.*, 2005]. The meteor radar detects Doppler

shifting of backscattered radio waves by meteor trails and determines hourly mean wind field at 1 km height interval from 80 km to 100 km through least squares fitting of the radial velocities of meteors.

[10] Using the method developed by *Tang et al.* [2005b], QM GWs are identified using two-dimensional spectral analysis. Wave characteristics and momentum flux are estimated for clear nights when both OH imager and meteor radar wind observation are available. The method of wave amplitude estimation used in this article was developed based on the analytical model relating OH volume emission rate measured by airglow images to the relative atmospheric density perturbation described by *Swenson and Gardner* [1998]. After retrieving horizontal wavelength and intrinsic frequency, the momentum flux estimation is based on the model described by *Swenson and Liu* [1998], which connects GW momentum flux to monochromatic GW spectral parameters and associated intensity perturbation.

[11] As described in detail by *Tang et al.* [2005b], OH imager data and simultaneous wind measurement by meteor radar were processed to identify high-frequency QM GWs and to estimate the momentum flux associated with them. First, the raw OH images were preprocessed by flat fielding, star suppression, and detrending to correct the coordinate distortion from the fish-eye lens, remove stars, and eliminate large horizontal-scale variations, respectively. The spectral analysis was conducted on the difference between two consecutive images, time difference (TD) images, to enhance the short-term variation of airglow intensity and damps the nonvariant background. To calculate the intrinsic GW parameters from TD images, each group of three consecutive images are corrected for Doppler shifting by horizontal winds. Horizontal winds for the OH airglow layer are calculated from meteor radar wind using a height weighting function that is comparable to the weighting function governing OH airglow emission. The height weight function is defined as the form $\exp(-(z - z_0)^2/\delta^2)$, where $z_0 = 85.9$ km and $\delta = 5.5$ km. The pixels in the first (third) image were shifted toward (against) the background wind direction to account for the linear distance translation during the time lapse from the center image.

[12] The method is appropriate on clear nights without major contamination. First, quality control is done to eliminate observations that are contaminated by moonlight and clouds by excluding images with large brightness (moon) and high contrast (cloud). However, even during clear nights, Milky Way moving across the sky produces many spurious waves. The star removal algorithm based on edge detecting technique cannot remove Milky Way because it is a patch of bright region on the OH images without sharp edges. An effective way to strongly attenuate Milky Way on TD images is to shift OH airglow images by the same distance as Milky Way advances during the interval between images such that Milky Way on the two images almost overlap. Before the correction, the strongest signal in the TD images are from the Milky Way; after the correction, the Milky Way are strongly damped and GW signals are more prominent. While the two-dimensional spectra of the TD images following Milky Way keep the GW spectra, they effectively remove those introduced by Milky Way. By comparing the spectra from TD images following star field and those calculated with Doppler-shifted TD images, the

true GW spectra were identified by choosing the common peaks. Comparing to *Tang et al.*'s [2005b] method, we only add an additional step to verify the dominant spectral peaks are not from Milky Way. When GW amplitude is weak, Milky Way signal may still contaminate our results. The results are further verified by examining TD movies. Suspicious waves propagate in the direction perpendicular to Milky Way alignments are not counted in the wave statistics. Waves with relative intensity perturbation smaller than 1% are also excluded in the wave characteristic statistics and momentum flux calculation to further avoid contamination.

3. Wave Characteristics

[13] Airglow imager observation relies on the dominance of intensity in the spectral range of airglow by the airglow emission. Favorable observation conditions are defined as nights with no moon or cloud cover. There are 513 nights of observation from 2002 to 2007 with at least 1 h favorable observation time. The number of nights with favorable observation condition for each month are summarized in Figure 1. For the majority of the months, we have more than 30 nights with favorable condition except for January and December. Especially in December, the total length of observation time is even shorter than what the number of favorable nights indicates since only a few nights have favorable conditions throughout the nights. Therefore the climatology of GW statistics for December is the least reliable due to its small sample size. Each sequence of three images with time interval of 2 min are analyzed as a group. Waves persisted during the 6 min time interval was considered a wave event. Analysis of these images reveals 9991 waves with persistence longer than 6 min. Wave event lasts longer than 6 min would be counted as several waves in our analysis with slightly varied wave characteristics. Several wave events in our analysis may only be counted for one event in other studies. The wave occurrence frequency is defined as the number of persistent waves with relative intensity perturbation larger than 1% over the number of observation interval (6 min). The time and month distribution of wave occurrence frequency is shown in Figure 2. Local time is 10 h behind universal time.

[14] When comparing our results with other studies, several differences should be noted. First, there are wave-like structures traveling with the same speed of background wind which have intrinsic phase speed close to zero. These non-GW structures, turbulence or small-scale inhomogeneity in the airglow layer traveling with background wind, are largely damped by applying correction on Doppler shifting by background wind. Using Doppler-shifted TD images in the analysis, our study would count less waves comparing to those do not remove these structures. Second, a lot of studies involve subjectively identifying wavefronts, whereas in our analysis QM GWs are extracted using spectral analysis. There are more waves from April to July and December with occurrence frequencies more than 50%. Fewer waves are observed from January to March with occurrence frequency lower than 30% most of time. Comparing to some studies [e.g., *Ejiri et al.*, 2003; *Dou et al.*, 2010], our results show a stronger seasonal variation. The abundance of GWs during late spring and early summer are not associated to convection occurrence frequency and

Figure 1. Number of nights with favorable observation condition longer than 1 h.

precipitation. Figure 3 shows the average convective pixel count as function a month from TRMM (The Tropical Rainfall Measuring Mission) satellite for both a $6^\circ \times 6^\circ$ domain and a $20^\circ \times 20^\circ$ domain centered at Maui. The larger domain is for the consideration of ducted GWs, which can travel much longer distance horizontally. The peak of wave occurrence season actually corresponds to a time of less convection and convective precipitation in both local region and the larger domain centered at Maui. Further investigation on the seasonal variation of observed wave occurrence frequency will be conducted in future research.

[15] In Figure 4, the horizontal wavelength histogram shows that horizontal wavelength (L_h) ranges from 10 km to 125 km and peaks in 15–30 km. This is consistent with previous studies [Taylor *et al.*, 1993; Swenson *et al.*, 2000; Hecht *et al.*, 2001; Nakamura *et al.*, 1999, 2003; Ejiri *et al.*, 2003; Hecht *et al.*, 2004; Dou *et al.*, 2010]. The abundance of waves in the range of 15–30 km has some implications on wave source or generation mechanism. Several modeling

studies have reported the peak horizontal wavelengths in the momentum flux spectra of GWs generated by convection are in this range [e.g., Lane and Moncrieff, 2008]. Though, typical horizontal wavelengths are found to be 20 to 30 km from various studies, the characteristic horizontal wavelength at some sites are longer, ranging from ~ 50 to ~ 80 km [Suzuki *et al.*, 2004, 2007]. Short horizontal wavelengths also correspond to short intrinsic wave period, 94% of waves with horizontal wavelength < 20 km have intrinsic wave period shorter than 10 min. The peak around 15 to 18 km is mostly composed of waves with intrinsic periods of 4–5 min. Assuming climatology of the static stability in the

Figure 2. GW occurrence frequency as function as month and universal time. Local time is 10 h behind universal time.

Figure 3. Time series of monthly convective pixel count for a $0.5^\circ \times 0.5^\circ$ region averaged over the local ($6^\circ \times 6^\circ$) and the larger domain ($20^\circ \times 20^\circ$) for each calendar month from 2002 to 2007. The latter larger domain is for the consideration of ducted waves.

Figure 4. Histogram of horizontal wavelength shows the clustering around 15–30 km. The bin width is 2 km.

OH airglow layer, most of them are evanescent. GWs with short horizontal wavelength are also more affected by background wind comparing to large horizontal wavelength waves since the modification of intrinsic phase speed is proportional to horizontal wave number.

[16] Intrinsic phase speeds are inferred from wave phase progression between two consecutive Doppler-shifted TD images. Figure 5 shows intrinsic phase speed (C) peak around 70 m s^{-1} with majority of the waves in the range of 50 to 90 m s^{-1} . Intrinsic phase speed varies as GW propagates through the mean flow. Thus it is important to examine how observed phase speed (the phase speed relative to the ground observer) distributes. Observed phase speed (C_g) is calculated by adding the background wind's projection on the wave propagation direction to the intrinsic phase speed. Figure 6 presents the histogram of observed phase speed, which shows a broader distribution with

Figure 5. Intrinsic phase speed histogram shows that the majority of GWs fall in the range between 50 and 100 m s^{-1} . The bin width is 4 m s^{-1} .

Figure 6. Observed phase speed histogram shows that the majority of GWs fall in the range between 20 and 70 m s^{-1} . The bin width is 4 m s^{-1} .

majority of waves in the range of 30 to 70 m s^{-1} . The fact that observed phase speeds are generally less than intrinsic phase speeds indicates the majority of GWs propagate against background wind. The nearly absence of waves with observed phase speeds close to zero is consistent of critical layer filtering by lower level winds which often switch sign in both zonal and meridional directions between the tropopause and the OH airglow layer. The observational limitation of airglow imager does not prohibit the observation of zero phase speed waves if they can reach the airglow layer.

[17] Intrinsic wave periods (T) are derived from horizontal wavelengths and intrinsic phase speeds. In Figure 7, intrinsic wave period histogram shows short-period waves dominate the distribution. Especially waves with intrinsic period shorter than 10 min account for 77% of the total. Intrinsic wave period distribution strongly skews toward short period

Figure 7. Intrinsic wave period histogram shows that short-period waves shorter than 10 min dominate the distribution.

Figure 8. Observed wave period histogram shows a distribution shifted toward longer period compared to intrinsic periods. Most GWs have observed periods larger than 8 min. The bin width is 2 min.

and tails off very fast as wave period increases. Most of the waves have wave periods shorter than 30 min. High-frequency GWs tend to have larger vertical wavelengths because they suffer little cancellation effect and are favored in imager observation. The majority (>80%) of the waves observed are not evanescent in the OH airglow layer. In fact, their ground-based (observed) wave periods are large enough that they cannot be ducted solely by the thermal structure in the MLT. Observed wave periods are inferred using observed phase speed and horizontal wavelength. Figure 8 shows the histogram of observed wave period (T_g). The distribution of observed wave period peaks at a larger value than intrinsic wave period which again indicates that background wind Doppler shifts GWs toward higher intrinsic frequency, larger vertical wavelengths in general. This implies that the majority of the GWs observed tend to propagate against the background wind. This is also consistent with the fact that the OH airglow intensity perturbation is more sensitive to larger vertical wavelength waves due to their small cancellation factors. *Nielsen et al.* [2009] showed that

$$m^2 = \frac{N^2 - \omega^2}{\omega^2 - f^2} k^2 - \frac{1}{4H^2}. \quad (1)$$

[18] With horizontal wavelength and wave period, vertical wavelengths (L_z) of GWs can be inferred according to the GW dispersion relationship (equation (1)). In equation (1), m is vertical wave number, N is Brunt-Väisälä or buoyancy frequency, ω is intrinsic GW frequency, f is inertial frequency, and $k = \frac{2\pi}{L_h}$ is horizontal wave number. When the vertical wave number m is real, a wave is vertically propagating, otherwise it is evanescent. N^2 is temperature dependent with typical value for the MLT is about $4 \times 10^{-4} \text{ s}^{-2}$. Because there is no concurrent measurement of temperature, N^2 is obtained by averaging over airglow layer (10 km centered at 87 km) using NRLMSISE-00 empirical model [*Picone et al.*, 2002] according to the local time and day of year at the time of a

wave event. Out of the total of 9991 waves, 1821 or 18% are evanescent. The calculation of m^2 depends on N^2 and background wind. Using climatology N^2 and hourly mean wind introduce a large uncertainty in the estimate of m^2 and may change its sign for waves with intrinsic wave period close to buoyancy period. Thus the percentage of evanescent waves may be different if temperature profiles in the MLT depart significantly from climatology. For the remaining waves, their vertical wavelength distribution peaks between 15 and 30 km and tail off quickly as wavelength increases. As shown by *Swenson and Gardner* [1998], waves with vertical wavelength shorter than the thickness of airglow layer (10 km for OH) should be greatly attenuated on OH imager. However, the observation shows a population of waves (about 14%) with inferred vertical wavelength shorter than 15 km, 4.3% of waves with inferred vertical wavelength shorter than 10 km. This may be due to the variation in the thickness of airglow layer in the real atmosphere or the amplitude of these waves are large enough to show up on imager despite their large cancellation factors. The other possibility is that the wave like structure are generated in situ [*Ejiri et al.*, 2002]. Figure 9 shows the histogram of vertical wavelength. The peak of vertical wavelength distribution centers around 20 km to 30 km. The decrease of waves with smaller vertical wavelength can be explained by stronger dissipation and insensitivity by airglow imager. The decrease of waves with larger vertical wavelength may be due to the smaller power in GW source spectrum for very high phase speed waves. Only the waves with vertical wavelengths larger than 15 km are included in the calculation of momentum flux because waves with vertical wavelength shorter than 15 km suffer large attenuation and introduce large uncertainty in the calculation of momentum flux due to their large cancellation factors [*Swenson and Liu*, 1998].

4. Propagation Direction and Momentum Flux

[19] The wave characteristics such as horizontal wavelength, phase speed, wave period, vertical wavelength do not change significantly with season. The most variable wave characteristics is the GW propagation direction. As

Figure 9. Histogram of vertical wavelength. The bin width is 2 km.

Figure 10. Wave propagation direction (blue) and wind direction (red) histograms for each month. The wind direction is the wind in the OH airglow layer at the time of a wave event.

shown in Figure 10, monthly histogram of GW propagation direction (blue) and corresponding background wind direction (red) vary significantly with season. In January, most waves propagate westward with background wind toward east. From February to April, the majority of waves propagate toward east with background wind mainly toward west. From May to July, GWs propagate to the north with background wind turn to the south. In August and September, the dominant wave propagation direction turn to east with background wind turn to west. In October, background wind shows no dominant direction and many waves propagate to the southeast. In November and December, dominant wave propagation direction is toward south. The poleward propagation of GWs during summer and equatorward propagation of GWs during winter in the midlatitudes have been noticed by previous studies as mentioned in section 1:

$$F_M = \frac{k}{m} \frac{g^2}{N^2} \left\langle \left(\frac{T'}{T} \right)^2 \right\rangle = \frac{kg^2}{mN^2 C_f^2} \left\langle \left(\frac{I'}{I} \right)^2 \right\rangle (m^2 s^{-2}). \quad (2)$$

[20] The method to calculate momentum flux is documented in detail by *Tang et al.* [2003]. With the assumption that GWs observed in the MLT propagate upward, GW momentum flux can be inferred from wave induced airglow intensity perturbation. Momentum fluxes are calculated using wave parameters deduced from the OH images as described by equation (2). F_M is momentum flux, k horizontal wave number, m vertical wave number, g acceleration due to gravity, $\frac{I'}{I}$ relative intensity perturbation. C_f is cancellation factor which relates temperature perturbation $\frac{T'}{T}$ to relative intensity perturbation $\frac{I'}{I}$. Smaller cancellation factor corresponds to stronger cancellation in airglow intensity perturbation observed on the ground. The direction of momentum flux is determined by the wave propagation direction relative to the mean flow. The magnitude of GW momentum flux is determined by the wave induced relative density perturbation, which in turn is related to the wave induced relative OH airglow intensity perturbation by the cancellation factor as a function of vertical wavelength [Swenson and Gardner, 1998]. For waves with vertical wavelengths shorter than 15 km, because their cancellations in the airglow layer are strong (small C_f s), small errors in airglow intensity perturbation divided by small C_f s would introduce large errors in inferred momentum flux. Thus waves with vertical wavelength shorter than 15 km are excluded in the calculation of momentum flux. These waves are only a small portion of our observation.

[21] The monthly mean GW momentum fluxes are plotted in Figure 11 with corresponding monthly mean background wind in the airglow layer. The magnitude of momentum fluxes is on the order of $10 \text{ m}^2 \text{ s}^{-2}$. The meridional component of momentum flux is about the same magnitude of zonal component. GW momentum fluxes at mesopause altitude are affected by both GW sources in the lower atmosphere and critical layer filtering by the mean flow between the sources and the MLT. Stronger meridional GW momentum fluxes are consistent with more convective activities in the north-south of Maui and stronger wave filtering in zonal direction in the stratosphere and lower mesosphere. The monthly mean zonal and meridional GW momentum flux tend to direct to the opposite direction of the mean flow. The monthly mean zonal flux shows more month to month variation while the monthly mean meridional flux shows more prominent annual oscillation. There is also more month to month variation in the zonal wind than in the meridional wind. The seasonal variation of zonal momentum flux shows high correlation to the variation of zonal mean wind in the layer in the OH airglow layer. The correlation coefficient between the monthly mean zonal wind at 87 km and monthly mean zonal flux is -0.79 , while the correlation coefficient for the meridional components is -0.61 . During summer the mean meridional circulation in the mesosphere is from summer pole to winter pole, the meridional momentum flux is directed toward the summer pole. During winter, the direction of meridional circulation and momentum flux are both the opposite. The monthly mean zonal wind shows a clear semiannual cycle with maximum westerly in spring equinox and maximum easterly in June and July. The seasonal variation of GW zonal momentum flux is noisier than that of GW meridional flux, which is similar to the noisier seasonal variation of mean zonal wind in the MLT.

Figure 11. Time series of monthly mean (a) zonal and (b) meridional momentum fluxes and zonal and meridional winds.

[22] Figure 12 shows the mean zonal and meridional momentum flux and zonal and meridional wind distribution as a function of universal time (UT) and month. Local time (LT) of Maui is 10 h behind UT. For the meridional component, in the summer the momentum flux are mainly northward and in the winter the momentum flux are mainly southward. For the zonal component, the momentum flux shows a semiannual variation more clearly with largely eastward momentum flux in the solstices and westward momentum flux around the equinoxes. Momentum flux seems to follow the pattern of background wind not only on the seasonal variation but also local time change. This does not necessarily require GWs to vary in characteristics such as amplitude and direction in the same manner as momentum flux direction. For example a stationary GW ($C_g = 0$) will generate momentum flux variation highly anticorrelated

to the background wind because the momentum flux measured will always be opposite to background wind direction as the sign of momentum flux is the same as $C - U$. The fact momentum flux generally points opposite to background wind supports the notion that GW acts as a damping mechanism for tides.

5. Causes of Preferred Wave Propagation Direction

[23] As mentioned in the introduction, preferred GW propagation directions in different seasons have been explained by several mechanisms: critical layer filtering, wave ducting, and source variation. All the above mechanisms are discussed in this section. In addition to these mechanisms, Doppler shifting by background wind would

Figure 12. Mean zonal and meridional momentum flux distribution along month and UT with corresponding wind component contours. The unit for wind speed is m s^{-1} .

also be investigated because it strongly affects wave intrinsic phase speeds and vertical wavelengths, which strongly affect wave dissipation and visibility on airglow imager.

5.1. Critical Layer Filtering

[24] Waves generated in the lower atmosphere suffer dissipation when they propagate upward. When they encounter critical layer, where their phase speeds equal the background wind, they break and deposit momentum to the mean flow. GW transmission has been investigated by *Stockwell and Lowe* [2001b] considering the dissipation of gravity waves through the lower atmosphere. Wave transmission is calculated according to equation (3) following [*Stockwell and Lowe*, 2001b; *Plumb and McEwan*, 1978]. The eddy viscosity is in the form of equation (4), which is based on work by *Matsuno* [1982]:

$$\tau(c, \theta, z_1, z_2) = \exp \left[- \int_{z_1}^{z_2} \frac{2\mu(z')N^3}{k_h[c - U(z') \cos \theta]^4} dz' \right] \quad (3)$$

$$\mu(z) = 400 \times 10^{\frac{z-110\text{km}}{40\text{km}}} \text{ m}^2 \text{ s}^{-1}. \quad (4)$$

Here, the source spectrum of gravity waves is specified for each UT and month with horizontal wavelength $L_h = 15:3:42$ km, observed wave period $T = 6:2:24$ min, and 24 azimuthal angles. The wavelengths and wave periods are chosen to be representative of the majority of gravity waves observed over Maui. The background atmosphere in the troposphere and stratosphere are specified using European Center for Medium-Range Weather Forecasts (ECMWF) reanalysis with monthly mean condition from January to December averaged from 2002 to 2007. Above 50 km, the temperature profiles are derived from NRLMSISE-00 empirical model. The wind profiles in the MLT region (80–100 km) are meteor radar observation for each UT and month averaged over 2002 to 2007. Where the meteor radar

Figure 13. Preferred wave propagation direction indicated by wave transmission percentage as a function of azimuthal angle.

Figure 14. Buoyancy period averaged over an 8 h period centered at midnight as a function of month and height at Maui.

and ECMWF reanalysis wind data are not available, an empirical model, Horizontal Wind Model-07 (HWM-07) [Drob *et al.*, 2008], are used to generate monthly mean hourly wind profile.

[25] Figure 13 shows the wave transmission as a function of azimuthal angle between 20 km and the bottom of OH airglow layer (80 km). In January and from June to September critical layer filtering effect seems to be important. In January the model predicts westward propagation, which is consistent with Figure 10. In June the prediction for critical layer filtering is mainly northward which shifts to mainly eastward by September, which are also seen in Figure 10. There is a northward and westward preferred propagation during summer, but only preferred westward propagation during winter. This does not match what is observed on OH airglow imager during winter. Also there is not much zonal preference in wave transmission in spring and fall, but there is an eastward bias on OH airglow observation. Therefore, critical layer filtering is important for several months but not the dominant mechanism causing the preferred propagation direction throughout the year at Maui.

5.2. Ducting

[26] Figure 14 shows the nightly mean buoyancy period as a function of month and altitude. The monthly mean temperature profiles in the middle atmosphere above Maui show that without background wind the thermal structure can afford the propagation of the majority of the gravity waves observed over Maui, most of which have observed wave periods larger than 8 min as shown in Figure 8. Yu and Hickey [2007] investigated the thermal ducting by the temperature profiles in the middle atmosphere using numerical modeling. The region between two low static stability layer (lower mesosphere and lower thermosphere) forms a thermal duct for waves with wave period of about 6–7 min as shown by Figure 14. The intrinsic periods of our observation also has a peak in its distribution at this range. GW breaking in the MLT region can generate secondary

waves with high frequencies [Franke and Robinson, 1999; Snively and Pasko, 2003]. This mechanism can seed ducted gravity waves with appropriate frequencies in the MLT region and propagate horizontally for an extended period in the OH airglow layer. This is beyond the scope of this research and will not be discussed further. Both zonal and meridional wind vary significantly on diurnal and seasonal time frames below and above OH airglow layer. Wave ducting considering only thermal structure is no longer valid for waves propagate directly from the lower atmosphere.

[27] To investigate wave ducting at Maui, m^2 of a representative set of wave parameters are examined with diurnal and seasonal variation of wind and temperature field of the whole atmosphere. Horizontal wavelength is chosen to be 30 km, which is the peak of horizontal wavelength distribution. Observed wave period is set to be 10 min. Shorter wave periods (6 and 8 min) and longer wave periods (12 and 16 min) are also tested. Shorter-period waves are easier to be ducted whereas longer-period waves suffer more dissipation. The background atmosphere for the investigation is monthly mean field with hourly change during nighttime. Temperature fields are set up using 2002–2007 6 year mean SABER (Sounding of the Atmosphere using Broadband Emission Radiometry) satellite observation from 20 km to 120 km. Above 120 km, the temperature profiles are calculated from NRLMSISE-00 model [Picone *et al.*, 2002]. Wind data below 80 km and above 100 km are from HWM-07 [Drob *et al.*, 2008]. Between 80 and 100 km, 6 years (2002–2007) averaged of monthly wind field from UIUC meteor radar are used.

[28] Figures 15, 16, 17, and 18 show the m^2 as function of UT and altitude for a wave with horizontal wavelength $L_h = 30$ km, observed wave period $T = 10$ min in December, March, June, and September, respectively. Where m^2 are negative, the gravity wave is evanescent. The green color shading indicates the optimal vertical wavelength range for wave observation on airglow imager. The blue, purple, and black color shading indicates strength of the evanescence layer. Yellow and red color shading indicates where vertical wavelengths are small and strong wave dissipation occurs. For each figure, m^2 for eastward (E), northward (N), westward (W), and southward (S) propagating waves are plotted.

[29] As shown by Figures 15–18, there is always an evanescence layer above 120 km for eastward propagating waves. This is due to the small N^2 above 120 km and generally westward wind there. Westward propagating waves are Doppler shifted to a lower intrinsic frequency and can still be freely propagating there. Due to the strong zonal wind in the stratosphere, there is strong wave filtering and evanescence for zonally propagating waves. In summer, strong easterly in the upper stratosphere–lower mesosphere flow filters westward propagating waves and reflects or attenuates eastward propagating waves. In winter, strong westerly flow in the upper stratosphere–lower mesosphere filters eastward propagating waves and reflects or attenuates westward propagating waves. Thus, the strong evanescence layer above 120 km only forms a ducting layer for eastward propagating waves around equinoxes when zonal flows in the upper stratosphere, lower mesosphere are weak. This is consistent with the fact that more waves are seen on airglow imager in March, April, September, and October. Eastward biases are also seen during these months as shown in

$m^{-2}(10^{-9}m^{-2})$ for GW ($T = 10$ min, $L_h = 30$ km)

Figure 15. The height-time cross section of m^2 for waves propagating toward four directions (east, north, west, and south) with 10 min observed wave period and 30 km horizontal wavelength in December. Dark blue indicates strong evanescence layer. Yellow and orange shading denotes strong dissipation. For positive m^2 corresponding vertical wavelength is label below its color bar.

$\text{m}^{-2}(10^{-9}\text{m}^{-2})$ for GW ($T = 10$ min, $L_h=30$ km)

Figure 16. Same as Figure 15 except for March.

Figure 10. Also it should be noted that convective sources during these months do not show stronger distribution to the west of Maui than to the east (not shown). Ducting alone

may play a role in enhanced wave occurrence frequency and preferred eastward propagation.

[30] For meridionally propagating waves, there are no critical layers or evanescence layers in the stratosphere. The

$\text{m}^{-2}(10^{-9}\text{m}^{-2})$ for GW ($T = 10$ min, $L_h = 30$ km)

Figure 17. Same as Figure 15 except for June.

evanescence layers and critical layers for these waves are mostly above 100 km due to meridional components of diurnal tide. Thus, the ducting condition for meridionally propagating waves only lasts for several hours. During

winter, southward propagating 10 min waves are ducted between 10 and 14 UT. During summer, northward propagating 10 min waves are ducted between 6 and 11 UT.

$\text{m}^{-2}(10^{-9}\text{m}^{-2})$ for GW ($T = 10$ min, $L_h=30$ km)

Figure 18. Same as Figure 15 except for September.

Though the dependence on local time does not show up in observed wave propagation direction preference.

[31] If waves with shorter periods (e.g., 6 min) are considered, the ducting layer for northward and southward

propagating waves almost always exist throughout the year. If the ducting mechanism is the dominant mechanism to determine wave propagation preference, one would expect the dominant wave source location also has a seasonal

Figure 19. Monthly mean convective pixel count as a function of latitude and month from TRMM satellite observation averaged over 2002–2007.

dependence as indicated by wave propagation. Figure 19 shows the distribution of convection from TRMM satellite observation as a function of latitude and month. During summer there is more convection in the southern domain of Maui, which can contribute to northward propagation preference in summer. During winter, however, convective sources exist both north and south of Maui with slightly more convection in the north and closer to the site. Thus, the observed southward propagation preference during winter can be partly explained by source location.

[32] The thermal structure in the above calculation are monthly nightly mean, while in the real atmosphere, the temperature profiles vary significantly due to strong tides, planetary and large-scale gravity waves. Mesospheric Inversion layers are also present on a sporadic time scales. Our estimation of the ducting condition may change if observed temperature profiles are used. However, one important fact is that the waves extracted from the observation have wave periods long enough (>6 min) which are affordable by the MLT even when diurnal tidal perturbation to the static stability profiles are considered. The other dominant factor to determine which direction are favored for ducting is the wind distribution. In the altitudes of 80 to 100 km, monthly hourly mean meteor radar wind profiles are used for this investigation. Most of the variations of the ducting condition are caused by wind variability throughout the MLT. However, due to the uncertainty in the temperature profiles the existence and location of evanescent layers might change significantly. Thus, even if the ducting condition in the above calculations are not favored for a certain time and month, this might not be the case if consideration is given to the large variability of temperature and wind profiles in the MLT.

[33] In summary, ducting mechanism are consistent with high wave occurrence frequency and eastward propagation preference in March, April, September, and October. Ducting for northward waves in summer and southward waves in winter are also present for several hours during the night, but they are not persistent and waves suffer strong dissipation above MLT due to tidal wind. Convective sources distribution is consistent with summer time northward propagation preference. During winter, stronger convections in the north of Maui are consistent with southward wave propagation preference though the contrast between northward and southward propagation should be less than during summer.

5.3. Doppler Shifting

[34] Both Figure 11 and Figure 12 show the momentum flux is highly anticorrelated with the background wind. The difference between intrinsic phase speed and observed phase speed distribution (Figures 5 and 6) indicates a large portion of GWs observed by OH imager propagate against background wind. The above observations are further verified explicitly by the observed wave propagation direction relative to in situ wind in the OH airglow layer. Figure 20 shows the azimuthal angle between each GW propagation direction and wind direction at the time of observation. The prominent feature is the highly clustered distribution of waves around 180° . About 44% of GWs are in the 60° sector centered on 180° . Ninety percent of GWs are in the opposite hemisphere in the polar diagram of the local background wind directions. This feature is consistent throughout the year and across different wavelengths and directions. This means the background wind's impact on GW propagation direction is very strong and consistent rather than a coincidence.

[35] Background wind affects GW intrinsic frequency through Doppler shifting as described in equation (5). ω_g is the wave frequency relative to the ground. According to GW dispersion relation as in equation (1), lower or higher intrinsic frequency corresponds to smaller or larger vertical wavelength. Thus, vertical wavelength is also modified by background wind

$$\omega = \omega_g - k \cdot U. \quad (5)$$

[36] If these GWs propagate in the same direction of mean wind, they are Doppler shifted to small intrinsic phase speed and small vertical wavelength, therefore unlikely to show up in the airglow imager. The intrinsic frequencies change drastically by the mean wind when GW phase speeds and background wind speeds are comparable, which is the case in the MLT. GWs with high observed phase speed are expected to be less affected by the background wind since their intrinsic phase speeds can still be large enough when they propagate in the direction of background wind. As shown in Figure 21, for waves with observed phase speed larger than 50 m s^{-1} , the direction difference between GWs and wind direction stills clusters around 180° , though the contrast between waves against and along background wind is smaller. Also noticed is the much larger observed phase speeds when waves are in the same direction of mean wind comparing to waves that are against mean wind. This is consistent with the theory that the observed waves propagating against background wind is related to observation limit of airglow imager and physical damping. Thus

Figure 20. Difference in GW propagation direction and background wind for all observed waves. Most waves propagate against the background wind. The numbers on the circle indicate percentage of total waves.

the monthly variation of wave propagation direction over Maui is more a manifestation of modulation by in situ background wind than critical layer filtering and source variations. The results shown here strongly indicate the observed dominant northward GW propagation during summer and southward GW propagation during winter over Maui are largely caused by the modulation of background wind in the airglow layer. More eastward GWs are also consistent with generally westward wind in the OH airglow layer during the observation time frame throughout the year. Of course other factors also contribute to the wave propagation direction. For example critical layer filtering in the stratosphere and lower mesosphere may contribute to eastward bias in wave propagation direction from April to September.

[37] Now we further examine how the background wind affects observed wave propagation statistics through Doppler shifting. The observation of GWs on OH airglow imager is affected by two main factors caused by Doppler shifting of the background wind. As indicated by GW dispersion relation, GWs propagate against the background wind are shifted to longer vertical wavelengths, shorter intrinsic wave periods and those propagate in the direction of the background wind are shifted to shorter vertical wavelengths and longer intrinsic wave periods. Longer vertical wavelength waves

not only suffer little dissipation but also introduce larger airglow relative intensity perturbation on airglow imager because they suffer little cancellation effect. Shorter vertical wavelengths waves, on the other hand, are strongly attenuated on the airglow imager due to their cancellation in the airglow layer in addition to their stronger physical dissipation due to their small vertical scales.

[38] To illustrate the effect of cancellation effect of airglow imager on causing observed GWs propagating against background wind, we can consider a simple case. For the typical static stability ($N^2 = 4 \times 10^{-4} \text{ s}^{-2}$) in the MLT, horizontal wavelength ($L_h = 25 \text{ km}$) and observed wave period ($T_g = 10 \text{ min}$), waves propagate along typical background wind ($U = 30 \text{ m s}^{-1}$) or perpendicular to the mean wind, their intrinsic phase speed C would be 11.7 m s^{-1} and 41.7 m s^{-1} . These correspond to intrinsic wave period T of 35.6 min and 10 min, vertical wavelength L_z of 3.7 km and 15.6 km, respectively. It is clear that for this typical wave when it propagates along the mean wind it will not show up on the OH imager; when it propagate perpendicular to the mean wind, its intensity perturbation will be slightly attenuated. On the contrary, when a wave with the same characteristic propagates against the mean wind, the intrinsic phase speed C becomes 71.7 m s^{-1} , intrinsic wave period T becomes 5.8 min, and vertical wavelength L_z becomes

Direction Difference for GWs with observed phase speed > 50 m/s

Figure 21. Difference in propagation direction and background wind for GW with observed phase speed larger than 50 m s^{-1} . There is still clustering around 180° . The numbers on the circle indicate percentage of total waves.

57.7 km. Figure 22 shows the dependence of vertical wavelength and cancellation factor as a function of direction angle between wave propagation direction and background wind direction for a typical GW ($L_h = 25 \text{ km}$, $T_g = 10 \text{ min}$). The black curve is the vertical wavelength as a function of azimuthal angle, the red curve is the corresponding C_f (ratio between $I/\bar{I}$ and $T/\bar{T}$). The ratio between C_f s in the direction against background wind and along background wind is as large as 871. The waves with the typical horizontal wavelength and observed wave period propagating in the direction of background wind are basically invisible. Therefore the observation limit of airglow imager on vertical wavelength not only excludes waves with short vertical wavelength but also strongly disfavor waves propagate in the same direction as the background wind. Thus observed wave propagation against background wind can be explained by the Doppler shifting by local mean wind. Due to the Doppler shifting by background wind, the instrumental filtering of OH imager and other instrument with observation limitation on vertical wavelength not only reflects on wave frequency, vertical wavelength, but also wave propagation direction. This effect is especially important for the MLT region because both diurnal and seasonal variation of zonal and meridional wind are strong comparing to stratosphere, which would significantly affect GW observation by Doppler shifting effect.

[39] Figure 23 shows the distribution of cancellation factor and vertical wavelength versus wave phase speed and propagation direction relative to background wind for a wave with typical horizontal wavelength of 30 km. The change of cancellation factor and vertical wavelength as a function of azimuthal angle is more prominent for slow to moderate phase speed which composes the majority of waves observed. Except for extremely high observed phase speed waves, by and large the ratio is substantially large that it suffices to say waves with same amplitudes are strongly favored when they propagate against background wind. The average ratio between cancellation factor of 180° and 0° is 10.04 for waves with observed phase speed from 10 to 90 m s^{-1} with background wind 30 m s^{-1} . Different models give substantially different value of C_f for the same atmospheric condition [Hickey and Yu, 2005]. Even the same model produces different C_f with different atmospheric conditions and wave dissipation assumption. However, the ratio of C_f between large vertical wavelength waves and small vertical wavelength waves are large for all these models and assumptions. Though the exact value of C_f need further verification, the validity of C_f as an important factor for preferred wave propagation direction still holds.

[40] On the other hand, wave dissipation also strongly depends on the vertical scale of waves, which also contributes to the apparent propagation of GWs against background wind. To compare its impact with that of cancellation

Figure 22. Cancellation factor (red) and vertical wavelength (black, in units of 10 km) as a function of propagation direction relative to background wind of 30 m s^{-1} . It resembles the distribution of observed waves along propagation direction relative to background wind direction.

factors, we calculated the transmission coefficient between 80 and 90 km for a GW with typical horizontal wavelength of 30 km using the method as described in section 5.1. The transmission coefficient indicates how much wave energy remains after damping due to dissipation. The results are summarized in Figure 24 for different observed phase speed and direction with typical MLT wind speed of 30 m s^{-1} . Comparing to cancellation factor, the dissipation effect depends more on intrinsic phase speed rather than angle difference between wave propagation and background wind. The ratio between wave transmission for waves against the background wind and along the background wind is also several times smaller comparing to cancellation factors. For waves with observed phase speed between 10 and 90 m s^{-1} , the average ratio of transmission between 180° and 0° is 1.7 comparing to 10 due to cancellation factor. For GWs with relatively large observed phase speed (larger than 50 m s^{-1}), their intrinsic phase speeds are not small enough to introduce strong damping for those traveling in the direction of background wind. These fast waves account for about half GWs observed. If there is no instrumental filtering, we would not observe strong preferential propagating against background wind for waves 10 or 20 m s^{-1} faster than average wind speed in the MLT (30 to 40 m s^{-1}). This is not consistent with our observation where waves with relatively large observed phase speed (larger than 50 m s^{-1}) still show prominent preference of against background wind propagation, which indicates the cancellation factor is the domi-

nant mechanism to modulate observed wave propagation direction.

[41] Thus the apparent dominance of waves propagating against background wind on imager associated with Doppler shifting is mostly caused by observational filtering of airglow imager rather than physical damping. If we can observe the short horizontal wavelength waves perfectly, due to the damping effect, short horizontal wavelength GWs are still expected to propagate against background wind though to a lesser degree than what we observed on OH airglow imager.

[42] Doppler shifting of intrinsic frequency is largely proportional to horizontal wave number, which explains why airglow observation of small-scale waves are strongly affected by background wind. Other instruments with limit on vertical resolution observing the same range of GW spectra would be expected to be affected by this effect too. Though the observed GW parameter and spectra are different, previous studies using radars have found Doppler-shifting effects may be large in MLT and frequency spectra of vertical velocity exhibit significant variability and dependence on mean wind [Fritts *et al.*, 1990; Fritts and Wang, 1991; VanZandt *et al.*, 1991]. Several studies using vertical velocity frequency spectra from radar observation have shown a tendency for high-frequency GWs to propagate preferentially against the local mean wind [Fritts and Wang, 1991; VanZandt *et al.*, 1991].

Figure 23. (a) Cancellation factor and (b) vertical wavelength as a function of wave phase speed and propagation direction relative to background wind of 30 m s⁻¹ for GWs with horizontal wavelength 30 km.

[43] The GW spectra observed by OH airglow imager are only a portion of the total GW spectra both in frequency and wavelength. Because waves propagate against the mean wind are strongly favored by airglow imager and waves propagate along the mean wind are damped on the airglow images, there is an instrumental bias. Therefore, even for the part of spectra (high-frequency or short horizontal wavelength waves) observed by airglow imager, we must be cautious to interpret the wave characteristics and momentum flux. Thus the simultaneous observation of wind at MLT is very important for interpretation of GW characteristics and momentum flux inferred from airglow imager. For example, when interpreting dominant GW propagation direction, we should examine how much is due to Doppler shifting of in situ background wind.

[44] In the MLT meridional circulation around solstices is from summer pole to winter pole. Airglow imager observation favor the GWs propagating against background wind, therefore GWs propagation shows strong concentration

toward summer pole. This may be a cause for the summer poleward propagation GWs observed by imagers for many sites in the midlatitudes. There are also other causes of dominant direction of GW propagation such as wave source, ducting condition, and critical layer filtering. When these factors are considered, the propagation preference against local wind may not be so prominent as in the case of Maui. In the case of Maui, it is situated in the central Pacific distant from tropical convection zone ITCZ (Intertropical Convergence Zone) in the south and midlatitude storm track in the north. The absence of strong meridional anisotropy in GW sources as in the observed meridional wave propagation preference is largely determined by background wind. Because observation of airglow is limited to nighttime, therefore in addition to seasonal variation of mean wind, the amplitude and phase of tidal wind also affect the background wind and observed wave propagation preference. For instance, at Maui (20°N), the meridional component of diurnal tide is large and usually southward in summer as is the mean meridional circulation. Thus during summer, there is a strong southward flow in the MLT over Maui at night. Around winter solstice, the mean meridional wind and meridional tidal component acting against each other and the mean MLT wind at night are largely zonal. Especially in December, the dominant southward propagation is likely due to source location rather than Doppler-shifting effects. Though it should be noted this month is also when we have the least observation. At higher latitudes, semidiurnal tides are stronger and the meridional component of wind can change sign during nighttime observation period without the presence of low-frequency waves such as inertial GWs and planetary waves. Latitudinal dependence of meridional wind may explain the latitudinal difference in wave propagation direction between higher latitudes and lower latitudes such as that noticed by *Ejiri et al.* [2003], where higher-latitude site (43.5°N) has waves propagating in both poleward and equatorward during winter.

Figure 24. GW transmission through OH airglow layer as a function of propagation direction relative to background wind of 30 m s⁻¹ for GWs with horizontal wavelength 30 km.

Figure 25. Observed power spectrum from OH imager compared to a curve $B m^{-3}$ and $B \cdot C_f m^{-3}$. On the high m region, the observed spectra is much lower than the curve predicted by saturation theory and close to the curve included cancellation factor.

[45] Despite the instrumental filtering, airglow imager can still capture the momentum flux from high-frequency GWs in the airglow layer. GWs that are shifted to short vertical wavelength, short intrinsic period suffer strong dissipation in the MLT, and deposit their momentum in the layer. Also previous studies have found that GW kinetic energy in the atmosphere are related to vertical wave number by $N^2/(6m^3)$ [Smith et al., 1987; Tsuda et al., 1989] for $m \gg m^*$, where m^* is characteristic wave number and corresponds to vertical wavelength of ~ 10 – 30 km [Fritts and Alexander, 2003] in the mesosphere. The spectral slope estimates at high m ($m > m^*$) also ranges from -2.5 to -3 according to various observation from radars [Fritts and Alexander, 2003]. Thus waves with short vertical wavelength have much smaller power in GW spectra. In fact, the observed vertical spectra of high-frequency GWs from imager follow a much steeper slope than Bm^{-3} as shown in Figure 25. B is chosen to match the magnitude of observed power near m^* , which is ~ 10 km vertical wavelength identified as the peak of GW kinetic energy spectra. The observed spectra between 10 and 6 km resembles a curve in the form of $Bm^{-3} C_f$. The steeper vertical spectral slope is caused by cancellation effect of airglow imager. Though airglow imager biases against short vertical wavelength waves, these waves also have much smaller power. Considering the above facts, OH imager might capture the majority of momentum fluxes of high-frequency GW spectrum after all. What are not observed are short vertical wavelength waves, which likely propagate in the same direction of background wind, suffer strong dissipation in the airglow layer and do not carry much momentum upward. Therefore momentum fluxes from high-frequency GWs are highly modulated by the background wind with momentum flux direction mostly against local background wind. Because the phase of diurnal tidal wind changes sign in about 10 to 15 km, the waves carrying momentum flux opposite to the local wind in the airglow layer would likely encounter wind in the same direction of wave propagation in the lower thermosphere. A detailed

investigation on GWs and tide interaction will be conducted in future research.

[46] Figure 20 shows not only few waves propagate toward the mean wind but also few waves propagate perpendicular to the mean wind. This can be explained if the wave amplitude observed by airglow imager and other propagation factors are considered. The distribution of GW relative intensity perturbations shows most waves are small amplitude waves. Most waves' amplitude in terms of relative intensity perturbation ($I'/\bar{I}$) are below 2%. These waves would not show up in the imager (threshold of $I'/\bar{I}$ is 1%) if they are not propagating against the background wind. There are 1648 waves with relative intensity perturbation larger than 4% in the sector 120° and 240° from mean wind direction. They can show up on imager if they are propagating perpendicular to the background wind considering the CF ratio in different direction for waves with observed phase speed between 40 and 80 m s^{-1} . The number of wave events in the two 30 degree sectors (75° to 105° and 255° to 285°) centered on directions perpendicular to the background wind direction are 615. It's about 40% of the number of large amplitude waves in the sector against background wind. The mismatch of the numbers indicate that other factors such as damping, ducting condition for specific direction are also important. Damping in the airglow layer introduces about a factor about 1.5 between waves against mean flow and those perpendicular to the mean flow for waves with typical parameters (horizontal wavelength 20–40 km, observed phase speed 40– 80 m s^{-1}). Wind speed around stratopause and above are large to affect the propagation of high-frequency gravity waves. The wind shear between stratopause and MLT also often shows a rotation larger than 90° except in April, September, and December (not shown). On the one hand, waves propagate in the same direction of upper stratosphere–lower mesosphere wind suffer strong dissipation and filtering. On the other hand, high-frequency waves propagate in the opposite direction of the wind in the upper stratosphere and lower mesosphere encounter a strong evanescence layer, suffer strong attenuation. For waves propagating against mean wind in the upper stratosphere–lower mesosphere, only slow and low-frequency waves can reach the OH airglow layer. They are not visible on the imager or have very small amplitudes due to physical damping. The filtering and evanescence layer in the upper stratosphere–lower mesosphere might help reduce the number of waves perpendicular to the mean wind in the MLT. In addition, a large portion (30%) of the large amplitudes ($>4\%$) waves in the former mentioned against mean wind sector are in the 60° sector centered on the east direction, which is the direction favored for ducting waves, especially in the months when wind rotation with altitude is weak.

[47] During April and September, when the winds in the OH airglow layer are generally easterly in the airglow layer, the ducting mechanism would increase the percentage of waves propagating in the opposite direction of mean wind. In December, compared to other months, significantly more waves propagate in the direction perpendicular to the mean wind. Figure 10 also shows that during December the wind vector rotation between the stratosphere and MLT is relatively small. As mentioned in the ducting discussion, in December the preference for southward wave propagation is

strengthened due to both favorable ducting condition for meridional propagation and stronger convection to the north of Maui. The dominant mechanism in December might be source location and ducting rather than Doppler shifting, which helps to explain that significantly more waves are in the direction perpendicular to the mean wind in the airglow layer in December.

[48] Though Doppler shifting can explain most of the feature in the observation, there are still several discrepancies as mentioned above. To fully explain the observed feature in wave propagation preference, all three factors must be considered: critical layer filtering, ducting, and Doppler shifting.

6. Summary and Conclusion

[49] The GW characteristics over Maui from 2002 to 2007 are deduced from OH airglow imager. The dominant horizontal wavelength falls between 15 and 30 km. Intrinsic phase speeds are mainly in the range of 50 m s^{-1} to 100 m s^{-1} . Observed phase speeds are mainly in the range of 20 m s^{-1} to 70 m s^{-1} . Intrinsic wave periods are mostly below 30 min with most of the waves clustered around 5–10 min. For intrinsic wave period larger than 5 min, as wave period increases the number of wave events decreases. Observed wave periods fall in the range of 8–20 min. The difference between intrinsic phase speed and observed phase speed, intrinsic wave period and observed wave period indicate that a large portion of GWs propagate against the background wind. Vertical wavelength distribution peaks between 20 and 30 km and decreases fast as wavelength increases.

[50] Observed GW propagation directions are largely related to the background wind in the airglow layer. Background wind Doppler shifts GWs propagating against (along) background wind to higher (lower) frequency and larger (smaller) vertical wavelength. The observed GWs tend to propagate against background wind except in December. The apparent against background wind propagation is consistent with contrast in cancellation factor and wave dissipation for waves propagate in different direction in most months.

[51] While previous studies investigated the roles of critical layer filtering, GW source locations and wave ducting as the cause of the wave propagation directions seen in airglow images there has been to date little consideration of the observational bias of this technique due to Doppler shifting. This study shows that wave cancellation effects can cause the wave amplitude to become below the threshold of the camera system causing a bias in those images for waves that propagate in directions away from the direction of the background wind. Our data show that in certain months contributions from other effects, such as (1) critical layer filtering and viscous dissipation (in January and April to September) and (2) ducting and source location (November to December), determines the directionality of the wave propagation. However, Doppler shifting effects are also of significance and need to be taken into account to determine the actual distribution of the observed wave propagation direction.

[52] Momentum fluxes deduced from OH imager are also highly anticorrelated with background winds. Although

imager observation disfavors waves propagate along background wind, waves with short vertical wavelength also have much smaller power in GW spectra. Despite the bias toward the direction against background wind, OH imager may capture the majority of momentum flux of high-frequency GW spectrum after all. Momentum flux from OH airglow imager generally points against the background wind. Momentum flux from imager shows strong anticorrelation with local mean flow on both seasonal and diurnal time frame.

[53] **Acknowledgments.** This work is supported by NSF grants ATM-1049451 and ATM-0737656. We would like to acknowledge the TRMM Online Visualization and Analysis System (TOVAS) for TRMM data. We would like to thank the anonymous reviewers for the critical reviews that helped to improve this paper.

References

- Dou, X., T. Li, Y. Tang, J. Yue, T. Nakamura, X. Xue, B. P. Williams, and C.-Y. She (2010), Variability of gravity wave occurrence frequency and propagation direction in the upper mesosphere observed by the OH imager in northern Colorado, *J. Atmos. Sol. Terr. Phys.*, **72**(5–6), 457–462, doi:10.1016/j.jastp.2010.01.002.
- Drob, D. P., et al. (2008), An empirical model of the Earth's horizontal wind fields: HWM07, *J. Geophys. Res.*, **113**, A12304, doi:10.1029/2008JA013668.
- Ejiri, M., K. Shiokawa, T. Ogawa, M. Kubota, T. Nakamura, and T. Tsuda (2002), Dual-site imaging observations of small-scale wave structures through OH and OI nightglow emissions, *Geophys. Res. Lett.*, **29**(10), 1445, doi:10.1029/2001GL014257.
- Ejiri, M., K. Shiokawa, T. Ogawa, K. Igarashi, T. Nakamura, and T. Tsuda (2003), Statistical study of short-period gravity waves in OH and OI nightglow images at two separated sites, *J. Geophys. Res.*, **108**(D21), 4679, doi:10.1029/2002JD002795.
- Franke, P. M., and W. A. Robinson (1999), Nonlinear behavior in the propagation of atmospheric gravity waves, *J. Atmos. Sci.*, **56**, 3010–3027, doi:10.1175/1520-0469(1999)056<3010:NBTP>2.0.CO;2.
- Franke, S., X. Chu, A. Liu, and W. Hocking (2005), Comparison of meteor radar and Na Doppler lidar measurements of winds in the mesopause region above Maui, Hawaii, *J. Geophys. Res.*, **110**, D09S02, doi:10.1029/2003JD004486.
- Fritts, D. C., and M. J. Alexander (2003), Gravity wave dynamics and effects in the middle atmosphere, *Rev. Geophys.*, **41**(1), 1003, doi:10.1029/2001RG000106.
- Fritts, D. C., and R. A. Vincent (1987), Mesospheric momentum flux studies at Adelaide, Australia: Observations and a gravity wave-tidal interaction-model, *J. Atmos. Sci.*, **44**(3), 605–619, doi:10.1175/1520-0469(1987)044<0605:MMFSA>2.0.CO;2.
- Fritts, D. C., and D.-Y. Wang (1991), Doppler-shifting effects on frequency spectra of gravity waves observed near the summer mesopause at high latitude, *J. Atmos. Sci.*, **48**(13), 1535–1544, doi:10.1175/1520-0469(1991)048<1535:DSEOF>2.0.CO;2.
- Fritts, D. C., U.-P. Hoppe, and B. Inhester (1990), A study of the vertical motion field near the high-latitude summer mesopause during MAC/SINE, *J. Atmos. Terr. Phys.*, **52**, 927–938, doi:10.1016/0021-9169(90)90025-1.
- Haque, R., and G. R. Swenson (1999), Extraction of motion parameters of gravity-wave structures from all-sky OH image sequences, *Appl. Opt.*, **38**, 4433–4442, doi:10.1364/AO.38.004433.
- Hecht, J. H., R. L. Walterscheid, M. P. Hickey, and S. J. Franke (2001), Climatology and modeling of quasi-monochromatic atmospheric gravity waves observed over Urbana, Illinois, *J. Geophys. Res.*, **106**, 5181–5195, doi:10.1029/2000JD900722.
- Hecht, J. H., S. Kovalam, P. T. May, G. Mills, R. A. Vincent, R. L. Walterscheid, and J. Woithe (2004), Airglow imager observations of atmospheric gravity waves at Alice Springs and Adelaide, Australia during the Darwin Area Wave Experiment (DAWEX), *J. Geophys. Res.*, **109**, D20S05, doi:10.1029/2004JD004697.
- Hecht, J. H., A. Z. Liu, R. L. Walterscheid, S. J. Franke, R. J. Rudy, M. J. Taylor, and P.-D. Pautet (2007), Characteristics of short-period wavelike features near 87 km altitude from airglow and lidar observations over Maui, *J. Geophys. Res.*, **112**, D16101, doi:10.1029/2006JD008148.
- Hickey, M. P., and Y. Yu (2005), A full-wave investigation of the use of a “cancellation factor” in gravity wave–OH airglow interaction studies, *J. Geophys. Res.*, **110**, A01301, doi:10.1029/2003JA010372.

- Lane, T. P., and M. W. Moncrieff (2008), Stratospheric gravity waves generated by multiscale tropical convection, *J. Atmos. Sci.*, **65**(8), 2598–2614, doi:10.1175/2007JAS2601.1.
- Matsuno, T. (1982), A quasi one-dimensional model of the middle atmosphere circulation interacting with internal gravity waves, *J. Meteorol. Soc. Jpn.*, **60**, 215–226.
- Medeiros, A. F., M. J. Taylor, H. Takahashi, P. P. Batista, and D. Gobbi (2003), An investigation of gravity wave activity in the low-latitude upper mesosphere: Propagation direction and wind filtering, *J. Geophys. Res.*, **108**(D14), 4411, doi:10.1029/2002JD002593.
- Medeiros, A. F., H. Takahashi, R. A. Buriti, K. M. Pinheiro, and D. Gobbi (2005), Atmospheric gravity wave propagation direction observed by airglow imaging in the South American sector, *J. Atmos. Sol. Terr. Phys.*, **67**, 1767–1773, doi:10.1016/j.jastp.2005.03.015.
- Nakamura, T., A. Higashikawa, T. Tsuda, and Y. Matsushita (1999), Seasonal variations of gravity wave structures in OH airglow with a CCD imager at Shigaraki, *Earth Planets Space*, **51**, 897–906.
- Nakamura, T., T. Aono, T. Tsuda, A. Admiranto, E. Achmad, and Suranto (2003), Mesospheric gravity waves over a tropical convective region observed by OH airglow imaging in Indonesia, *Geophys. Res. Lett.*, **30**(17), 1882, doi:10.1029/2003GL017619.
- Nielsen, K., M. J. Taylor, R. E. Hibbins, and M. J. Jarvis (2009), Climatology of short-period mesospheric gravity waves over Halley, Antarctica (76°S, 27°W), *J. Atmos. Sol. Terr. Phys.*, **71**, 991–1000, doi:10.1016/j.jastp.2009.04.005.
- Pautet, P., M. Taylor, A. Liu, and G. Swenson (2005), Climatology of short-period gravity waves observed over northern Australia during the Darwin Area Wave Experiment (DAWEX) and their dominant source regions, *J. Geophys. Res.*, **110**, D03S90, doi:10.1029/2004JD004954.
- Picone, J. M., A. E. Hedin, D. P. Drob, and A. C. Aikin (2002), NRLMSISE-00 empirical model of the atmosphere: Statistical comparisons and scientific issues, *J. Geophys. Res.*, **107**(A12), 1468, doi:10.1029/2002JA009430.
- Plumb, R. A., and A. D. McEwan (1978), The instability of a forced standing wave in a viscous stratified fluid: A laboratory analogue of the quasi-biennial oscillation, *J. Atmos. Sci.*, **35**, 1827–1839, doi:10.1175/1520-0469(1978)035<1827:TIOAFS>2.0.CO;2.
- Smith, S., D. C. Fritts, and T. VanZandt (1987), Evidence for a saturated spectrum of atmospheric gravity-waves, *J. Atmos. Sci.*, **44**, 1404–1410.
- Snively, J. B., and V. P. Pasko (2003), Breaking of thunderstorm-generated gravity waves as a source of short-period ducted waves at mesopause altitudes, *Geophys. Res. Lett.*, **30**(24), 2254, doi:10.1029/2003GL018436.
- Stockwell, R. G., and R. P. Lowe (2001a), Airglow imaging of gravity waves: 1. Results from a small network of OH nightglow scanning imagers, *J. Geophys. Res.*, **106**, 17,185–17,203, doi:10.1029/2001JD900035.
- Stockwell, R. G., and R. P. Lowe (2001b), Airglow imaging of gravity waves: 2. Critical layer filtering, *J. Geophys. Res.*, **106**, 17,205–17,220, doi:10.1029/2001JD900036.
- Suzuki, S., K. Shiokawa, Y. Otsuka, T. Ogawa, and P. Wilkinson (2004), Statistical characteristics of gravity waves observed by an all-sky imager at Darwin, Australia, *J. Geophys. Res.*, **109**, D20S07, doi:10.1029/2003JD004336.
- Suzuki, S., K. Shiokawa, Y. Otsuka, T. Ogawa, M. Kubota, M. Tsutsumi, T. Nakamura, and D. C. Fritts (2007), Gravity wave momentum flux in the upper mesosphere derived from OH airglow imaging measurements, *Earth Planets Space*, **59**, 421–428.
- Swenson, G., and C. Gardner (1998), Analytical models for the responses of the mesospheric OH* and Na layers to atmospheric gravity waves, *J. Geophys. Res.*, **103**(D6), 6271–6294, doi:10.1029/97JD02985.
- Swenson, G., and A. Liu (1998), A model for calculating acoustic gravity wave energy and momentum flux in the mesosphere from OH airglow, *Geophys. Res. Lett.*, **25**(4), 477–480, doi:10.1029/98GL00132.
- Swenson, G., M. Alexander, and R. Haque (2000), Dispersion imposed limits on atmospheric gravity waves in the mesosphere: Observations from OH airglow, *Geophys. Res. Lett.*, **27**(6), 875–878, doi:10.1029/1999GL010738.
- Tang, J., F. Kamalabadi, A. Liu, and G. Swenson (2003), Extraction of momentum flux of monochromatic gravity waves using spectroscopic imaging, in *Geoscience and Remote Sensing Symposium Proceedings, IEEE Catalog Ser.*, vol. 6, pp. 3961–3963, IEEE Press, Piscataway, N. J.
- Tang, J., G. Swenson, A. Liu, and F. Kamalabadi (2005a), Observational investigations of gravity wave momentum flux with spectroscopic imaging, *J. Geophys. Res.*, **110**, D09S09, doi:10.1029/2004JD004778.
- Tang, J., F. Kamalabadi, S. Franke, A. Liu, and G. Swenson (2005b), Estimation of gravity wave momentum flux with spectroscopic imaging, *IEEE Trans. Geosci. Remote Sens.*, **43**(1), 103–109, doi:10.1109/TGRS.2004.836268.
- Taylor, M., E. Ryan, T. Tuan, and R. Edwards (1993), Evidence of preferential directions for gravity wave propagation due to wind filtering in the middle atmosphere, *J. Geophys. Res.*, **98**(A4), 6047–6057, doi:10.1029/92JA02604.
- Taylor, M. J., W. R. Pendleton, S. Clark, H. Takahashi, D. Gobbi, and R. A. Goldberg (1997), Image measurements of short-period gravity waves at equatorial latitudes, *J. Geophys. Res.*, **102**, 26,283–26,299, doi:10.1029/96JD03515.
- Tsuda, T., T. Inoue, S. Kato, S. Fukao, D. C. Fritts, and T. E. VanZandt (1989), MST radar observations of a saturated gravity wave spectrum, *J. Atmos. Sci.*, **46**, 2440–2447, doi:10.1175/1520-0469(1989)046<2440:MROOAS>2.0.CO;2.
- VanZandt, T. E., G. D. Nastrom, and J. L. Green (1991), Frequency spectra of vertical velocity from flatland VHF radar data, *J. Geophys. Res.*, **96**, 2845–2855, doi:10.1029/90JD02220.
- Walterscheid, R., J. Hecht, R. Vincent, I. Reid, J. Woithe, and M. Hickey (1999), Analysis and interpretation of airglow and radar observations of quasi-monochromatic gravity waves in the upper mesosphere and lower thermosphere over Adelaide, Australia (35°S, 138°E), *J. Atmos. Sol. Terr. Phys.*, **61**(6), 461–478, doi:10.1016/S1364-6826(99)00002-4.
- Yu, Y., and M. P. Hickey (2007), Numerical modeling of a gravity wave packet ducted by the thermal structure of the atmosphere, *J. Geophys. Res.*, **112**, A06308, doi:10.1029/2006JA012092.

S. Franke and G. Swenson, Department of Electrical and Computer Engineering, University of Illinois at Urbana-Champaign, 1308 W. Main St., Urbana, IL 61801, USA.

Z. Li and X. Lu, Department of Atmospheric Sciences, University of Illinois at Urbana-Champaign, 105 S. Gregory St., Urbana, IL 61801, USA. (zli2@atmos.illinois.edu)

A. Z. Liu, Department of Physical Sciences, Embry-Riddle Aeronautical University, Daytona Beach, FL 32114, USA.