Gravity Waves-induced Airglow Temperature Variations, Phase Relationships, and Krassovsky Ratio for OH(8,3) Airglow, O₂(0,1) Atmospheric Band, and O(¹S) Greenline in the MLT Region

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Abstract

The gravity wave-induced secular variations and fluctuations of airglow intensity-weighted temperatures for OH(8,3) nightglow, O₂(0,1) atmospheric band, and O(¹S) greenline in the MLT region were simulated to study the wave effects' on airglow temperatures. They were investigated with a time-dependent OH Chemistry-Dynamics (OHCD) model and a Multiple Airglow Chemistry-Dynamics (MACD) model with a small-scale linear gravity wave packet. The largest wave-induced airglow temperature secular variations are found to be ~ 0.8% in the O₂(0,1) temperature and the largest wave-induced airglow temperature fluctuations are found to be ~ 0.25 % in the O(¹S) temperature. We also investigated the phase relationships between the airglow intensities and temperatures. Our results show that the airglow intensities lead the airglow temperatures most of the time. Also, airglow located at a higher altitude leads the airglow located at a lower altitude, indicating a downward phase progression, which is consistent with the phase of the wave packet we used. The Krassovsky ratio for these airglow emissions were calculated and found to be decreasing with increasing altitude. The amplitude of the Krassovsky ratio for OH(8,3) airglow, O₂(0,1) atmospheric band, and O(¹S) greenline were ~10, 6, and 4, respectively.

Keywords: gravity waves, OH nightglow, atmospheric band, greenline, secular variation, airglow temperature, phase relationship, Krassovsky ratio

1. Introduction

We present in this paper the numerical results of gravity wave-induced airglow secular variations and fluctuations of OH(8,3) airglow, O₂(0,1) atmospheric band, and O(¹S) greenline temperature in the Mesosphere/Lower Thermosphere (MLT) region. The results of gravity wave-induced airglow secular variations and fluctuations of these airglow intensities have been published in Huang and George [2014] (HG2014, hereafter). Since both studies use the same numerical models for the investigations, the reader is referred to HG2014 for the methodology and for more detailed model description. Here we will only briefly describe the salient features of the models and essential results from HG2014 when necessary. Both studies use a two-dimensional, time dependent, nonlinear OH Chemistry-Dynamics (OHCD) model and a Multiple-Airglow Chemistry-Dynamics (MACD) model to investigate the wave-induced secular variations and fluctuations of airglow intensities and airglow-intensity weighted temperatures. The important findings from HG2014 are that a small-scale linear gravity wave packet can induce a substantial secular increase in the airglow intensities. The largest secular increase of airglow intensity is found in O(¹S) greenline (~33%), followed by O₂(0,1) atmospheric band (~30%), and then OH(8,3) airglow (~22%). As for the wave-induced airglow intensity fluctuations, it is found that OH(8,3) airglow exhibits the largest fluctuation amplitude (~2.4%) followed by O₂(0,1) atmospheric band and O(¹S) greenline that have a comparable maximum fluctuation amplitude (~1%). The other important finding from HG2014 is that the airglow intensities and volume emission rates are very

sensitive to the branching ratios of the three body recombination reaction (see their Table 1, reactions 1-3) and that strategic observations are suggested in order to better assess the branching ratio values.

There have been several studies examining the phase relationship between airglow and its temperature using observational data [Chung, et al., 2006; Taori and Taylor, 2006; Takahashi et al., 2011]. The results of their studies do not always converge. It seems that the phase relationship between airglow and its temperature is rather complicated and that it depends on the chemistry and properties of waves. In this study, in addition to reporting the results of wave-induced secular variations and fluctuations of airglow temperatures, we also present the phase difference between airglow intensities, between airglow temperatures, and between airglow intensity and its weighted temperature for the three airglow emissions under study. Multiple airglow intensity variations located at different altitudes can provide information of the gravity waves that have caused the variations. The theoretical studies by Liu and Swenson [2003] and Vargas et al. [2007] show that the relative phase information of different airglow layers separated in altitude can be used to estimate the vertical wavelength of gravity waves propagating through the layers. The state of the gravity waves, whether they are propagating, stationary, or dissipating, can also be inferred from the phase differences of airglow intensities. Therefore, our simulation results of the phase relationships of multiple airglow intensities and temperatures should be of interest to the Aeronomy community.

The Krassovsky ratio is another quantity that the experimenters and modelers in the Aeronomy community oftentimes calculate from their observations or their modeling results [Krassovsky, 1972; Walterscheid et al., 1994; Makhlouf et al., 1995; Reisin and Scheer, 2001; Takahashi et al., 2011]. It is a quantity that relates relative airglow intensity fluctuation to relative airglow temperature fluctuation. The phase of the Krassovsky ratio can provide information on the state of the waves that are either evanescent, standing, or travelling. In this paper, we calculated and presented the amplitude and phase of the Krassovsky ratio for the three airglow emissions.

In this paper, we briefly describe the OHCD and the MACD model in Section 2. Different airglow temperature definitions are listed in Section 3. Our numerical results and Krassovsky ratio calculations are presented in Section 4. Discussion is in Section 5 and Conclusions are in Section 6.

2. The OHCD and the MACD Model

The OHCD model consists of a Spectral Full Wave Model (SFWM) and an OH airglow chemistry model [Huang and Hickey, 2008]. The MACD model consists of the same SFWM and a multiple-airglow chemistry model for O₂ atmospheric bands ((0,0) and (0,1)) and O(¹S) greenline [Huang and George, 2014]. The SFWM simulates a linear, transient, dissipative gravity wave packet [Hickey et al., 2000]. In this study, we use a small-scale gravity wave packet (peak amplitude less than 5% at the peak altitude at 104 km) that consists of waves of different horizontal wavelengths and periods. The forcing gravity wave has a horizontal wavelength of 30 km and a period of 20 min. The vertical

wavelength of the forcing wave packet is ~ 6 km. Major gas species' number densities and temperature are output from the MSIS-90 model [Hedin, 1991]. They can also be found in Figure 9 of Huang and Hickey [2007]. Initial number densities of minor species reacting in the airglow chemistries at a latitude of 18 degrees north in March are taken from Garcia and Solomon [1985] and supplied by Garcia [1990, private communication].

3. Different Airglow Temperature Definitions

There are three most commonly used temperatures inferred from airglow emissions in the Aeronomy community. They are intensity-weighted temperature (called brightness-weighted temperature in Makhlof et al. [1995]), rotational temperature, and Doppler temperature. Modelers have traditionally used the intensity-weighted temperature in their modeling work [Hines and Tarasick, 1987; Walterscheid et al., 1987; Schubert et al., 1991; Huang and Hickey, 2008]. The airglow intensity-weighted temperature is defined

as $\langle T_I \rangle = \frac{\langle IT \rangle}{\langle I \rangle}$, where the angled brackets denote column integration of the enclosed

variable, and where I is airglow volume emission rate and T is atmospheric temperature.

The rotational temperature and Doppler temperature are derived from airglow intensity measurements depending on the type of instrument used. For example, Michelson interferometer is often used to measure intensities of rotational lines [Thuillier and Herse, 1991; Shepherd et al., 1993] and Fabry-Perot interferometer is used to measure the Doppler shift and broadening of an emission line [Hernandez, 1980; Hernandez and Smith, 1984]. The rotational temperature is derived from the ratio of measured rotational line intensities in a band of an emission. The Doppler temperature is derived from the spectral width of a single emission line. Makhlof et al. [1995] gave a detailed account

on these airglow temperature definitions and derivations within their analytical framework.

In our work, we simulate the wave-induced intensity-weighted temperatures with our numerical models. We will defer the discussion of the implications of using this temperature for the Krassovsky ratio calculations as opposed to using the other two temperatures to the Discussion section.

4. Results

The simulation results of wave-induced airglow intensity-weighted temperatures as a function of time are shown in Figure 1. The dash-dotted curve is for OH(8,3) temperature (short for OH(8,3) airglow intensity-weighted temperature, and so forth for the other two airglow temperatures), the dashed curve is for O₂(0,1) atmospheric band temperature, and the solid curve is for O(¹S) greenline temperature. Distinctive wave oscillations can be clearly seen in the temperatures from the time at 220 min to 410 min. However, there is no steep increase in the temperatures as is seen in the airglow intensities, e.g., Figure 2 of HG2014. The OH(8,3) temperature is initially at 196.2 K and shows a slight increase with time and reaches a maximum value of 197.7 K at the time of 335 min, but then it starts to decline slightly with an ending temperature at 196.1 K. The initial O₂(0,1) atmospheric band temperature is at 182 K, and the temperature increases and reaches a maximum value of 183.7 K at 350 min, and then it slightly decreases to an ending value of 183.2 K. As for the O(¹S) greenline temperature, the initial value is 180 K. It shows a miniscule increase with a maximum value of 180.7 K at

the time around 330-350 min, and then the final O(¹S) greenline temperature is at 180.15 K. The net wave-induced temperature change is -0.1 K (-0.05%) for OH(8,3) temperature, +1.2 K (+0.66%) for O₂(0,1) atmospheric band temperature, and +0.15 K (+0.08%) for O(¹S) greenline temperature. Bear in mind that we used a small-scale linear gravity wave packet, and thus a small temperature change is expected. The small temperature change induced by the wave packet indicates that the wave packet's main effect on the airglow temperatures is mostly shown through the wave pattern in the airglow temperatures but not in magnitude.

We did a running average on the airglow temperatures to separate the wave-induced fluctuations from the wave-induced secular variations of the airglow temperatures. We show the wave-induced fluctuations of the airglow temperatures as a function of time in Figure 2. The dash-dotted curve denotes the OH(8,3) temperature fluctuation, the dashed curve denotes the O₂(0,1) atmospheric band temperature fluctuation, and the solid curve denotes the O(¹S) greenline temperature fluctuation. From the figure, we can see that the O(¹S) greenline temperature fluctuation exhibits the largest maximum peak-to-peak amplitude of 0.48% occurring at around 308 min. The OH(8,3) temperature fluctuation has the second largest maximum peak-to-peak amplitude of 0.45% occurring at around 280 min, and the O₂(0,1) atmospheric band temperature fluctuation has the smallest maximum peak-to-peak amplitude of 0.32% occurring at 305 min. The OH(8,3) temperature fluctuation reaches its maximum first while the O(¹S) greenline temperature fluctuation reaches its maximum last but has the largest amplitude of the three. This is different from the airglow intensity fluctuation, i.e., the OH(8,3) intensity fluctuation

reaches its maximum first and has the largest fluctuation amplitude of the three airglow intensity fluctuations [HG2014].

The time evolution of wave-induced secular variations of the airglow temperatures are shown in Figure 3. The dash-dotted curve denotes the secular variation of OH(8,3) temperature, the dashed curve denotes the secular variation of O₂(0,1) atmospheric band temperature, and the solid curve denotes the secular variation of O(¹S) greenline temperature. The 30-km wave packet induces an insignificant secular change in all three airglow temperatures. The secular variation of OH(8,3) temperature initially is in the increasing trend but after the time at 340 min when it reaches the largest increase (~0.58%), it starts to decline. The net secular change is around -0.05% of its initial value at the end of the simulation time. On the other hand, the secular variations of O(¹S) greenline temperature and O₂(0,1) atmospheric band temperature also show an increase with time, and then they go down slightly for about 20 min after then they go slightly up again. After the time at 400 min the magnitudes of the secular variations become approximately constant. The O(¹S) greenline temperature shows a net 0.18% secular increase and the O₂(0,1) atmospheric band temperature shows a net 0.75% secular increase. The largest secular change in the O(¹S) greenline temperature is seen at 370 min with a magnitude of 0.2%, whereas for the O₂(0,1) atmospheric band temperature it occurs ten minutes earlier with a magnitude of 0.75%. By examining Figures 2 and 3, it is interesting to note that the maximum secular change occurs approximately 3 wave cycles (~60 min) after the temperature fluctuation reaches its maximum value for all three airglow temperatures. However, such behavior is not seen in the airglow intensity

simulation results in HG2014. In addition, temperature fluctuations are larger than temperature secular variations for OH(8,3) and O(¹S) greenline temperatures. However, the secular variations of airglow intensities are in general much larger than the airglow intensity fluctuations [HG2014].

The phase relationships of multiple airglow intensities or airglow temperatures located at different altitudes can oftentimes reveal whether a gravity wave's amplitude is increasing, decreasing, or constant [Reisin and Scheer, 2001; Liu and Swenson, 2003; Taori and Taylor, 2006; Vargas et al., 2007]. Here we first show the results of phase difference of any two of the airglow intensities as a function of time in Figure 4. The vertical scale is the phase difference divided by 2π , ranging from -1 to $+1$. The curve with solid circles denotes the phase difference between O(¹S) greenline and OH(8,3) airglow intensity, the curve with solid triangles denotes the phase difference between O₂(0,1) atmospheric band and OH(8,3) airglow intensity, and the curve with solid squares denotes the phase difference between O(¹S) greenline and O₂(0,1) atmospheric band intensity. To find the phase difference, we use the relationship $\frac{\Delta t}{T} = \frac{\Delta \phi}{2\pi}$, where Δt is the time difference of airglow intensity fluctuations of any two neighboring airglow intensity peaks, T is the forcing wave period, 20 min in our case, and $\Delta \phi$ is the phase difference in radians. We identify the times when the airglow intensity fluctuations reach their peaks (see Figure 4 in HG2014), and then we calculate the time difference of adjacent peaks using the times of the airglow located at the lower altitude as the reference values, i.e., for the O1S-OH phase differences, we subtract the peak times of OH(8,3) airglow intensity fluctuation from the peak times of O(¹S) greenline intensity fluctuation. The airglow intensity

fluctuations' peak times can be clearly identified between the time from 180 min to 420 min, and so we only show the phase differences in that time range. From the figure, we can see that $O(^1S)$ greenline leads both the $O_2(0,1)$ atmospheric band and $OH(8,3)$ airglow. $O_2(0,1)$ atmospheric band initially lags but soon leads $OH(8,3)$ airglow. All three curves also show that the magnitudes of the phase differences increase with time and do not maintain constant. This is in large part due to the wave packet's amplitude changing with time and altitude coupled with the airglow chemistry. From the figure, it can be seen that the change of $O1S-OH$ phase difference is the greatest of the three pairs ($\sim -1.3\pi$) if we compare the phase difference at the start and end points. The change of $O2B1-OH$ phase difference and that of $O1S-O2B1$ phase difference are roughly the same ($\sim -0.64\pi$).

The phase differences of any two of the airglow temperatures as a function of time are presented in Figure 5. The procedure for obtaining the phase differences is the same as the one described in the preceding paragraph. The peak times are identified from the airglow temperature fluctuation data from 180 min to 420 min. From Figure 5, we can see that the $OH(8,3)$ temperature initially leads both the $O_2(0,1)$ atmospheric band temperature and $O(^1S)$ greenline temperature, but then the situation quickly reverses where the $O(^1S)$ greenline temperature leads the $OH(8,3)$ temperature at around 200 min and the $O_2(0,1)$ atmospheric band temperature leads at around 230 min. The $O(^1S)$ greenline temperature leads the $O_2(0,1)$ atmospheric band temperature throughout the entire simulation time. The phase relationships showing which temperature is leading or lagging are apparent in the wave trains in Figure 2. Again, we can see in Figure 5 that

the phase differences of the three pairs do not remain constant. Comparing the phase differences at the start and end points, we see that the change of the O1S-OH temperature phase differences is the largest ($\sim -1.6\pi$), followed by the O2B1-OH pair ($\sim -1.1\pi$), and then by the O1S-O2B1 pair ($\sim -0.4\pi$). Comparing Figures 4 and 5, we can see that the general behavior of the phase differences for airglow intensity fluctuations and airglow temperature fluctuations is similar, i.e., the phase differences get larger (more negative) with time and the phase difference of the O1S-O2B1 pair has the smallest change. The latter is not surprising since the vertical separation of the peak altitudes of these two emissions is the smallest ~ 2 km.

Figure 6 shows the phase difference between airglow intensity fluctuation and airglow temperature fluctuation for the three airglow emissions as a function of time. The curve with solid circles is for OH(8,3) airglow, the curve with triangles is for O₂(0,1) atmospheric band, and the curve with solid squares is for O(¹S) greenline. Again, the vertical axis is set to be the phase difference over 2π for easy inspection. To find the phase differences, we first subtract the peak times of airglow temperature fluctuation from those of airglow intensity fluctuation, and then the time differences are divided by the forcing wave period of 20 min. As the curves show, the OH(8,3) temperature initially leads the OH(8,3) airglow intensity, but soon at around 210 min the intensity leads the temperature. As for O₂(0,1) atmospheric band and O(¹S) greenline, both intensities lead their temperatures. The phase difference between airglow intensity and its temperature is not constant. If we compare the start and end points of the curves, we find a net change of -0.4π for OH(8,3) airglow, $+0.12\pi$ for O₂(0,1) atmospheric band, and -0.15π for O(¹S)

greenline, indicating that the phase difference between intensity and its temperature increases with time (more negative) for OH(8,3) airglow and O(¹S) greenline. Our results of the phase differences presented in Figures 4-6 suggest that the phase relationship of these quantities is complex. The variation of the phase relationship is primarily due to the dissipative upward propagating wave packet modulating the airglow species reacting in the airglow chemistry in different altitudes. In addition, the cancellation effect of the integrated airglow quantities within the airglow layers for waves of short vertical wavelength also affects the phase relationships.

We also calculated the Krassovsky ratio using our simulation results of airglow intensities and airglow temperatures. The Krassovsky ratio, η , is a measure of the airglow response to a passage of wave trains and is defined as

$$\eta = \frac{I' / \bar{I}}{T' / \bar{T}}, \quad (1)$$

where $I'(T')$ is the wave-perturbed airglow intensity fluctuation (airglow temperature fluctuation) and $\bar{I}(\bar{T})$ is the average of the airglow intensity fluctuation (airglow temperature fluctuation) over the emission layer [Krassovsky, 1972]. From equation 1, we can see that η is the ratio of relative intensity fluctuation to the relative temperature fluctuation. Since our simulation results of airglow intensity and airglow temperature fluctuations are normalized to their initial values and not to their average values, we need to convert our fluctuation data to relative fluctuation data before we can calculate the Krassovsky ratio as defined in Eq. (1). Our airglow intensity and airglow temperature fluctuations in the models are calculated using the following expressions

$$flu_I = \frac{I'}{I_0} \text{ and } flu_T = \frac{T'}{T_0}, \quad (2)$$

where $I'(T')$ is the same as in Eq.(1), and where $I_0(T_0)$ is the initial airglow intensity (airglow temperature) value. Thus, the relative airglow intensity fluctuation is related to our airglow intensity fluctuation as

$$\frac{I'}{\bar{I}} = \frac{I'}{I_0} \frac{I_0}{\bar{I}} = (flu_I) * \left(\frac{I_0}{\bar{I}}\right) \quad (3)$$

Likewise, the relative airglow temperature fluctuation is related to our temperature fluctuation as

$$\frac{T'}{\bar{T}} = \frac{T'}{T_0} \frac{T_0}{\bar{T}} = (flu_T) * \left(\frac{T_0}{\bar{T}}\right) \quad (4)$$

So the Krassovsky ratio can be calculated using our fluctuation data multiplied by a factor C

$$\eta = \frac{I' / \bar{I}}{T' / \bar{T}} = \frac{flu_I * (I_0 / \bar{I})}{flu_T * (T_0 / \bar{T})} = C * \frac{flu_I}{flu_T}, \quad (5)$$

where $C = \frac{I_0 / \bar{I}}{T_0 / \bar{T}}$ and can be easily calculated using our initial and average values of the

airglow intensity and airglow temperature from our simulation results. Also, we note that the Krassovsky ratio can be expressed in terms of amplitude and phase angle as

$$\eta = \eta_{\max} e^{-i\phi}, \quad (6)$$

where $\eta_{\max}$ is the amplitude and ϕ is the phase difference between I' and T' . So to find the amplitude of the Krassovsky ratio for the three airglow emissions, we first find the maximum fluctuation amplitude of airglow and that of airglow temperature. Then $\eta_{\max}$ is calculated by taking the ratio of the fluctuation amplitudes multiplied by the factor C.

To find the phase difference ϕ , we first find the time difference Δt between the maximum fluctuation amplitude of airglow and that of airglow temperature, and then we use $\phi = \frac{\Delta t}{T} \times 360^\circ$ to find ϕ , where T is the forcing wave period. Our calculation results of the C value, the maximum airglow intensity and airglow temperature fluctuation amplitude ($I_{\max}$ and $T_{\max}$), the amplitude and phase of the Krassovsky ratio ($\eta_{\max}$ and ϕ) for the three airglow emissions in response to the 30-km wave packet are listed in Table 1. We can see that the amplitude of the Krassovsky ratio gets smaller with increasing altitude for the 30-km wave packet. The phase difference between airglow intensity and its temperature is larger for $O(^1S)$ greenline while the other two have a similar phase difference.

5. Discussion

We note that our simulation results of wave-induced airglow temperature variations are rather small. Bear in mind that the spectral full wave model we employed in this study only simulates a packet of linear dissipative gravity waves. So it does not include irreversible changes induced by wave-mean flow interactions like wave breaking or encountering a critical level in the model. The wave-induced secular variations and fluctuations of airglow intensity-weighted temperatures presented in the current paper are due to the wave-perturbed temperatures coupled with changes in the altitude of the airglow layers. The simulated secular changes of airglow temperatures are different from the temperature change brought about by the wave-mean flow interactions. For example, a substantial temperature rise was observed (~ 40 K) during the ALOHA-93 campaign. It was demonstrated that the wave-mean flow interaction at the critical level was strong

enough to raise the temperature in a few hours accompanied by a large change in the wind shear [Huang et al.,1998; 2002]. Our secular changes of airglow temperatures are also different from the dynamical heating due to the divergence of the wave-driven sensible heat flux [Hickey and Walterscheid, 1994] or the wave-induced chemical heating from the exothermic reactions [Huang and Hickey, 2012].

As was discussed in HG2014, the production of O_2 atmospheric bands and $O(^1S)$ greenline depends strongly on the branching ratios in the three-body recombination reactions. Recall that the branching ratio values used in our MACD model were $(7 \times 10^{-5}, 0.04)$ for (ϵ, α) shown in Table 1 in HG2014. We compared our airglow intensity results to the results when another simulation run used the (0.11, 0.8) combination as in Hickey et al. [1993] to see if the wave-induced effects would be any different when different branching ratios are used. We found that our wave-induced secular variation and fluctuation of $O(^1S)$ greenline are the same as those when the (0.11, 0.8) combination is used. The secular variation and fluctuation of $O(^1S)$ greenline temperature also show similar results to those of the $O(^1S)$ greenline intensity, i.e., the actual values of $O(^1S)$ greenline intensity and temperature are different when different branching ratios are used, but the percent change magnitudes and phases remain the same. As for $O_2(0,1)$ atmospheric band, it is found that our secular variations and fluctuations of $O_2(0,1)$ atmospheric band intensity and temperature are larger than those using the (0.11, 0.8) branching ratios and the phases are nearly identical. However, the $O_2(0,1)$ atmospheric band temperature has a maximum fluctuation amplitude that is $\sim$ three times smaller and the secular variation shows a decreasing trend when the (0.11, 0.8) combination is used.

Simultaneous measurements of both emissions would provide a means to better assess the branching ratio values.

The phase relationship of airglow intensity or airglow temperature in multiple airglow layers can provide information of the gravity wave propagating through the airglow layers. For example, if the airglow intensity (or airglow temperature) located at a higher altitude leads the airglow intensity (airglow temperature) located at a lower altitude, it indicates downward phase propagation, i.e., upward energy propagation of the gravity wave. The amplitude ratio of airglow intensity or airglow temperature fluctuations from multiple airglow layers also reveals whether the gravity wave's amplitude is increasing, decreasing, or constant [Liu and Swenson, 2003; Vargas et al., 2007]. Another phase relationship that is of interest to the scientific community is the phase between airglow intensity and airglow temperature. Observations have shown that sometimes airglow temperature leads airglow intensity [Chung et al., 2006; Takahashi et al., 2011] and sometimes it is the opposite [Taori and Taylor, 2006]. Our simulation results show that all three airglow intensities under study lead their respective temperatures, though OH(8,3) airglow initially shows the opposite before the time at 210 min. The observations of O₂(0,1) atmospheric band in Maui, Hawaii, analyzed by Taori and Taylor, showed that the waves observed had periods ranging from 1-12 hours and the η value increases with increasing wave periods while the phase ϕ showing a decreasing trend. Their observations also showed that O₂(0,1) atmospheric band intensity leads its temperature. We note several things regarding their observations and analysis. First, the observations in their Figure 1 show a rather large variation of the airglow intensity (~

90% to 78%) and airglow temperature (16%) from their mean values. Second, as a side note, the increasing trend in the left panel of their Figure 1 looks very much like the secular increase shown in Figure 2 of HG2014. Taori and Taylor removed the trends (the large variability) by first doing a simple sinusoidal best fit on the data and then subtracting the fit from the observation before finding wave fluctuation from the data of de-trended, residual mean deviations. Essentially, they treated the large variability as of long-period tidal influences where we think it might be nothing more than the wave-induced secular variations. Would the results of their analysis be any different had they considered it as secular variations remain open and require further investigations. As cautioned by Huang and Hickey [2008], the wave-induced secular variations can oftentimes be mistaken as long-period or short-period waves in the airglow observations, and so care must be taken when analyzing the data from observations. The observations of $O_2(0,1)$ atmospheric and $OH(6,2)$ airglow at Rikubetsu, Japan showed that the wave periods range from 2 to 10 hrs and the airglow temperatures lead the airglow intensities for a wave with a period of 2.8 ± 0.2 h [Takahashi et al., 2011]. Also, some of their observations showed that $O_2(0,1)$ atmospheric band leads $OH(6,2)$ airglow and other observations showed no phase difference between the two emissions suggesting that the waves were evanescent. Another study by Chung et al. [2006] reported their observations of $O_2(0,1)$ atmospheric band and $OH(6,2)$ airglow at King Sejong Station in Antarctic over a two-year time period. Their findings found that there were two dominant waves with periods of 4 and 6 h that seem to be of tidal origin and airglow temperatures lead airglow intensities for both airglow emissions.

As we mentioned in the preceding paragraph, our results show that airglow intensity leads airglow temperature for the airglow under study. A modeling work by Liu and Swenson [2003] (LS2003 hereafter) shows the opposite phase relationship for the OH and O₂ airglow. Given that some observations have shown phase relationships that either agree with our results or with the LS2003 result, it is premature to draw a definitive conclusion on the phase relationship between airglow and its temperature. Here we only want to note the differences of our models and their model to better understand the possible causes for this discrepancy. First of all, our models are time-dependent and so they simulate the transient nature of wave-airglow interactions, whereas the model in LS2003 simulates the steady-state of the interactions. Our damping profile from eddy and molecular diffusions are altitude-dependent, whereas they use a constant damping rate for the altitudes of interest. Both studies use MSIS90 model as the reference model but ours is for latitude 18 N and theirs is for latitude 35 N. Lastly, our vertical wavelength is ~ 6 km and theirs start from 10 km. All these model differences could give rise to the noted discrepancy of the results.

The Krassovsky's ratio can be used to deduce the chemical processes in the airglow chemistry responsible for its production and it can also be used for tentative estimation of the long-term and seasonal variations of minor species involved in the airglow chemistry [Khomich et al., 2008]. The Krassovsky ratio depends on quite a few parameters, for example, the wave periods, vibrational levels, quenching, horizontal wavelengths, wave dissipation rate, variations of local oxygen photochemistry, height variation of the emission layer, and temperature definition [Schubert et al., 1991, Hickey et al., 1993;

Makhlouf et al., 1995; Liu and Swenson, 2003; Vargas et al., 2007]. Since there are rather many factors affecting the Krassovsky ratio, it is not surprising to find a rather wide range of the ratio reported in the literature. Takahashi et al. reported a value of 5.1 for $O_2(0,1)$ atmospheric band and 14 for $OH(6,2)$ airglow for an upward propagating wave with a period of 2.8 ± 0.2 h. On the other hand Chung et al. reported the η value ranging from 2 to 8 for $O_2(0,1)$ atmospheric band and 5 to 6 for $OH(6,2)$ airglow for the dominate waves they observed. A look at their tables 2-5 listing the η values and phases ϕ for the 2- and 4-h waves, it can be seen that the phases ϕ are rather different despite the fact that the oscillations were identified to have the same wave periods. The reports of these observations we discussed in our paper were used to show the diverse nature of phase relationships and Krassovsky ratio. Apparently, there is still a lot that needs to be investigated in order to fully understand the complex nature of airglow intensities and airglow temperatures when there are waves propagating through the airglow layers.

In our study, we used the intensity-weighted temperature for the calculations of the Krassovsky ratio. We obtained the η amplitude for each airglow emission (9.68, 5.76, and 4.09 for $OH(8,3)$ airglow, $O_2(0,1)$ atmospheric band, and $O(^1S)$ greenline, respectively) and found it to be decreasing with increasing height. Our values are within the ballpark of the values reported in the literature. The study by Makhlouf et al. [1995] using the $OH(6,2)$ airglow shows that for a small-scale gravity wave the amplitude of η is larger when the intensity-weighted temperature is used in the calculation, shown in their Figure 6. For a wave of period of 20 min, it shows that the η amplitude is overestimated by ~23% when compared to the value using the rotational temperature in the calculation.

As for the phase ϕ , their study shows that the phase is underestimated by about 12.5% when the intensity-weighted temperature is used in the calculation comparing to the phase calculated using the rotational temperature for a small-scale gravity wave with a period of 20 min, as shown in their Figure 8. Their study also shows that both the amplitude and phase of Krassovsky ratio depend on vibrational level of OH airglow. In their Figure 11, they showed the results for three different OH vibrational levels (1, 6, and 9). Specifically, their η value for a wave period of 20 min is ~ 8 for the OH vibrational level $v=9$. Their results indicate that the difference between levels 6 and 9 is about 12.5% for the η value and very small for the phase ϕ for a wave with a period of 20 min. Since these values for higher vibrational levels show smaller deviations, their results of vibrational level at 9 are somewhat indicative of the results for the OH(8,3) airglow.

6. Conclusions

In this study, we present the results of gravity-wave induced airglow intensity-weighted temperature variations simulated by the OHCD and the MACD model. Three airglow temperatures, OH(8,3) airglow, O₂(0,1) atmospheric band, and O(¹S) greenline, are under the investigation. A small-scale gravity wave packet that has a forcing wave period of 20 min and a forcing horizontal wavelength of 30 km is used for the study. Our simulation results show that the 30-km wave packet does not induce a significant secular increase or fluctuation in the airglow temperatures. The wave-induced secular variations in the airglow temperatures show that the O₂(0,1) atmospheric band temperature has a net

0.75% increase and the $O(^1S)$ greenline temperature has a net 0.18% increase, whereas the OH(8,3) airglow temperature shows a net 0.05% decrease.

The wave-induced airglow temperature fluctuations show that $O(^1S)$ greenline and OH(8,3) airglow temperatures fluctuate at a similar maximum amplitude (0.48% - 0.45%) and the $O_2(0,1)$ atmospheric band temperature fluctuates at a smaller maximum amplitude (0.32%). Note that our simulation results of airglow intensities and temperatures also show that the wave-induced airglow intensity fluctuations are an order of magnitude larger than the airglow temperature fluctuations. These results are similar to the results reported in Huang and Hickey [2008] where they only considered OH(8,3) airglow. That the airglow intensity fluctuations are larger than the temperature fluctuations is not surprising. The reported values of Krassovsky ratio in the literature are all greater than 1, indicating that airglow responds to a passage of waves much greater than airglow temperature does.

We also present the phase difference of airglow intensities, airglow temperatures, and airglow intensity and its temperature for the three airglow emissions. Our results show that the airglow emission located at a higher altitude leads the airglow emission located at a lower altitude, indicating an upward energy (or rather downward phase) propagating gravity wave, consistent with the wave packet used in the simulation. Our calculated values of Krassovsky ratio for the three airglow emissions are within the ballpark of the values reported in the literature. The reported airglow observations seem to indicate that the phase relationships of airglow and Krassovsky ratio are rather complex and their

behaviors are not always consistent among the studies. Since they depend on quite a few factors, it is not surprising that the results of their studies do not always converge. However, it is apparent that more observations and modeling are required to better understand how gravity waves affect our atmosphere.

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Figure Captions

Figure 1. The wave-induced variations of $O(^1S)$ greenline temperature (solid curve), $O_2(0,1)$ atmospheric band temperature (dashed curve), and $OH(8,3)$ airglow temperature (dash-dotted curve) as a function of time.

Figure 2. The wave-induced fluctuations of $O(^1S)$ greenline temperature (solid curve), $O_2(0,1)$ atmospheric band temperature (dashed curve), and $OH(8,3)$ airglow temperature (dash-dotted curve) as a function of time.

Figure 3. The wave-induced secular variations of $O(^1S)$ greenline temperature (solid curve), $O_2(0,1)$ atmospheric band temperature (dashed curve), and $OH(8,3)$ airglow temperature (dash-dotted curve) as a function of time.

Figure 4. The phase difference (over 2π) of any two of the airglow intensities as a function of time. The curve with solid circles denotes the phase difference between $O(^1S)$ greenline intensity and $OH(8,3)$ airglow intensity, the curve with solid triangles denote the phase difference between the $O_2(0,1)$ atmospheric band intensity and $OH(8,3)$ airglow intensity, and the curve with solid squares denotes the phase difference between $O(^1S)$ greenline intensity and $O_2(0,1)$ atmospheric band intensity.

Figure 5. Same as Figure 4 but for airglow temperatures.

Figure 6. The phase difference (over 2π) between airglow intensity and its weighted temperature as a function of time. The curve with solid circles is for OH(8,3) airglow, the curve with triangles is for O₂(0,1) atmospheric band, and the curve with solid squares is for O(¹S) greenline.

Table 1. The amplitude and phase of Krassovsky ratio for the three airglow emissions in response to the 30-km wave packet.

	OH(8,3)	O ₂ (0,1)	O(¹ S)
C	0.91	0.88	0.85
I _{max}	2.41E-2	1.08E-2	1.15E-2
T _{max}	2.27E-3	1.65E-3	2.40E-3
η _{max}	9.68	5.76	4.09
φ	-35.10°	-33.30°	-54.90°

Figure 1

Figure 2

Figure 3

Airglow Temperature Secular Variations

T=20 min, Lx=30 km, Lat=18 N

Figure 4

Phase Difference of Airglow Intensities

T=20 min, Lx=30 km, Lat=18 N

Figure 5

Phase Difference of Airglow Temperatures

T=20 min, Lx=30 km, Lat=18 N

Figure 6