

SCHOLARLY COMMONS

Publications

6-16-2010

Atmospheric Airglow Fluctuations due to a Tsunami-driven Gravity Wave Disturbance

Michael P. Hickey Ph.D. Embry-Riddle Aeronautical University, hicke0b5@erau.edu

G. Schubert Institute of Geophysics and Planetary Physics, University of California

R. L. Walterscheid *The Aerospace Corporation*

Follow this and additional works at: https://commons.erau.edu/publication

Part of the Atmospheric Sciences Commons

Scholarly Commons Citation

Hickey, M. P., G. Schubert, and R. L. Walterscheid (2010), Atmospheric airglow fluctuations due to a tsunami-driven gravity wave disturbance, J. Geophys. Res., 115, A06308, doi: https://doi.org/10.1029/2009JA014977

This Article is brought to you for free and open access by Scholarly Commons. It has been accepted for inclusion in Publications by an authorized administrator of Scholarly Commons. For more information, please contact commons@erau.edu.



Atmospheric airglow fluctuations due to a tsunami-driven gravity wave disturbance

M. P. Hickey,¹ G. Schubert,^{2,3} and R. L. Walterscheid³

Received 26 October 2009; revised 11 January 2010; accepted 18 February 2010; published 16 June 2010.

[1] A spectral full-wave model is used to study the upward propagation of a gravity wave disturbance and its effect on atmospheric nightglow emissions. Gravity waves are generated by a surface displacement that mimics a tsunami having a maximum amplitude of 0.5 m, a characteristic horizontal wavelength of 400 km, and a horizontal phase speed of 200 m/s. The gravity wave disturbance can reach F region altitudes before significant viscous dissipation occurs. The response of the OH Meinel nightglow in the mesopause region (~ 87 km altitude) produces relative brightness fluctuations, which are $\sim 1\%$ of the mean for overhead viewing. The wave amplitudes grow as the wave disturbance propagates upward, which causes the thermospheric nightglow emission responses to be large. For overhead viewing, the brightness fluctuations are ~50% and 43% of the mean for the OI 6300 Å and O 1356 Å emissions, respectively. The total electron content fluctuation is $\sim 33\%$ of the mean for overhead viewing. For oblique viewing, the relative brightness fluctuations are slightly smaller than those obtained for overhead viewing. In spite of this, the thermospheric nightglow brightness fluctuations are large enough that oblique viewing could provide early warning of an approaching tsunami. Thus, the monitoring of thermospheric nightglow emissions may be a useful augmentation to other observational techniques of tsunami effects in the thermosphere/ionosphere system.

Citation: Hickey, M. P., G. Schubert, and R. L. Walterscheid (2010), Atmospheric airglow fluctuations due to a tsunami-driven gravity wave disturbance, *J. Geophys. Res.*, *115*, A06308, doi:10.1029/2009JA014977.

1. Introduction

[2] Hines [1972] and Peltier and Hines [1976] first considered the possibility that tsunamis could excite atmospheric acoustic gravity waves and that the gravity waves would subsequently propagate to ionospheric heights and manifest themselves as traveling ionospheric disturbances. Subsequent observations of the ionosphere following tsunami events and supporting modeling studies have confirmed the viability of this suggestion [Artru et al., 2005a, 2005b; Occhipinti et al., 2006, 2008; Hickey et al., 2009]. These studies, as well as some of the references they contain, focused on the effects of tsunami-driven gravity waves on the electron number density in the F region ionosphere and on the total electron content (TEC) fluctuations so produced. The latter is a useful quantity of interest because the GPS network can provide fairly complete global coverage of such events. Hence, interpreting the TEC signatures through

Copyright 2010 by the American Geophysical Union. 0148-0227/10/2009JA014977

modeling should be an important part of such an observational program.

[3] Other processes in the atmosphere should also be influenced by tsunamis. These include atmospheric airglow emissions, which are known to be strongly influenced by atmospheric gravity waves. Here, we specifically select for further study the OH nightglow emanating from the mesopause region near 87 km altitude, and the OI 6300 Å and O 1356 Å nightglow emissions emanating from the middle thermosphere (250–300 km altitude).

[4] The interaction of gravity waves with the mesopause region OH Meinel nightglow emissions has been extensively studied [Sivjee et al., 1987; Hecht and Walterscheid, 1991; Taylor et al., 1991; Swenson and Mende, 1994; Hickey et al., 1992; Hickey, 1993]. The OH emission layer is centered near 87 km altitude and is fairly narrow, with a full width at half maximum of about 10 km. Gravity waves of short vertical wavelength (<15 km) tend to be more difficult to observe in airglow emission measurements due to cancellation between positive and negative fluctuations that occur over a vertical wavelength within the airglow layer [Hines and Tarasick, 1987; Schubert and Walterscheid, 1988]. Tsunami-generated gravity waves tend to have large-phase speeds (~200 m/s) and large vertical wavelengths for which such cancellation should be minimal. Hence, tsunami-generated gravity waves may be observable in the OH nightglow. Additionally, although the mesopause region (and the various airglow emissions) is pervaded by

¹Department of Physical Sciences, Embry-Riddle Aeronautical University, Daytona Beach, Florida, USA.

²Department of Earth and Space Sciences and Institute of Geophysics and Planetary Physics, University of California, Los Angeles, California, USA.

³Space Science Applications Laboratory, The Aerospace Corporation, Los Angeles, California, USA.



Figure 1. (a) Surface displacement (Z) calculated using equation (1) for a maximum displacement of 0.50 m, and (b) the vertical velocity spectrum (\hat{W}) associated with Z.

waves of many varying scales, they typically have smallphase speeds (~50 m/s), and hence discriminating between them and tsunami-generated gravity waves should, in principle, be straightforward. Ultimately, the viability of such observations will depend on the amplitude of the OH nightglow perturbations and on the amplitude of the tsunami.

[5] Fast gravity waves will be able to propagate into the thermosphere where they will interact with chemical processes affecting the OI 6300 Å and O 1356 Å nightglow emissions. These emissions peak in the range of 250-300 km altitude, and unlike the OH mesospheric emission, they have a large vertical extent with a full width at half maximum of ~100 km. The 6300 Å emission has also been used to study gravity wave motions [Sobral et al., 1978; Mendillo et al., 1997; Kubota et al., 2001]. The observed phase speeds were 50–100 m/s for a geomagnetically quiet period and 300–700 m/s during a geomagnetic storm [Kubota et al., 2001]. The far ultraviolet (FUV) O 1356 Å emission is produced by electron recombination in the F region ionosphere. Its measurement as a strong indicator of thermospheric ion/electron chemistry and dynamics is well established [e.g., Paxton et al., 2003; DeMajistre et al., 2007].

[6] The objective of this paper is to simulate the interaction of a tsunami-generated gravity wave disturbance with the mesospheric OH nightglow emission and with the thermospheric OI 6300 Å and O 1356 Å nightglow emissions in order to assess if ground-based airglow measurements could identify such wave events. Accordingly, we perform simulations for a very specific set of inputs (wave spectrum parameters, direction of propagation, latitude, solar and geomagnetic indices, and the undisturbed mean state), and we neglect the effects of mean winds. These inputs favor gravity wave propagation to the middle thermosphere with a strong ionospheric response, and therefore, we expect the simulated airglow response to represent a plausible upper bound to such effects. Although not specifically considered here, we note that our results could also be applied to satellite observations of these emissions.

2. Theory and Models

[7] We follow the approach of *Hickey et al.* [2009] and model the tsunami surface displacement (shown in Figure 1a) as a prescribed analytic function of position, as given by *Peltier and Hines* [1976]. The tsunami is taken to propagate at the shallow water wave speed of -200 m/s. The associated vertical velocity spectrum (Figure 1b) has peaks at ± 400 km, which, combined with a phase speed of 200 m/s, corresponds to a dominant period of 33 min. A northward propagating wave is forced at the lower boundary (z = 0) of our full-wave model by specifying the vertical velocity there. Denoting the full-wave model output as $\psi_j'(\omega,k,z)$, where subscript *j* signifies temperature, pressure, or velocity, the response to lower boundary forcing is

$$\Psi'_{j}(x+\nu t,z) = \frac{1}{2\pi} \int_{-\infty}^{\infty} \hat{W}(k,0,0) \psi'_{j}(\omega,k,z) e^{-ik(x+\nu t)} \mathrm{d}k.$$
(1)

[8] Equation (1) was implemented by forcing every wave to have a vertical velocity of 1 m/s at the lower boundary $(w'(\omega,k,0) = 1 \text{ m/s})$. The product $\hat{W}(k,0,0) \psi_j'(\omega,k,z) \Delta k$ represents the wave-number-dependent perturbation. As discussed by *Peltier and Hines* [1976], many of the waves in the spectrum are evanescent and will not propagate very far vertically. Specifically, acoustic waves of phase speed 200 m/s are subsonic and so are evanescent. Hence, the spectrum is truncated to include only propagating waves. The minimum period is 3.0 min and corresponds to a minimum horizontal wavelength of 36 km. Equation (1) is solved for a spectrum of

 Table 1. Ion-Related Chemistry Used in the Model

		Reaction		Rate $(\text{cm}^3 \text{ s}^{-1})$	
1	$O^{+} + N_{2}$	\rightarrow	$NO^{+} + N$	$k_1 = 5 \times 10^{-13}$	
2	$O^+ + O_2$	\rightarrow	$O_{2}^{+} + O$	$k_2 = 2 \times 10^{-11} (T/300)^{-0.4}$	
3	$O^+ + NO$	\rightarrow	$\overline{NO^{+}} + O$	$k_3 = 8 \times 10^{-13}$	
4	$O^+ + e$	\rightarrow	$O(^{5}P) + \hbar\nu(1356 \text{ Å})$	$k_4 = 7.3 \times 10^{-13}$	
5	$O_{2}^{+} + N_{2}$	\rightarrow	$NO^+ + NO$	$k_5 = 5 \times 10^{-16}$	
6	$O_2^+ + NO$	\rightarrow	$NO^+ + O_2$	$k_6 = 4.4 \times 10^{-10}$	
7a	$O_{2}^{+} + e$	\rightarrow	$O + O(^{3}P)$	$k_{7a} = 0.62 \times 1.6 \times 10^{-7} (300/T)^{0.55}$	
7b	$O_2^+ + e$	\rightarrow	O + O(1S)	$k_{7b} = 0.08 \times 1.6 \times 10^{-7} (300/T)^{0.55}$	
7c	$O_{2}^{+} + e$	\rightarrow	$O + O(^{1}D)$	$k_{7c} = 1.30 \times 1.6 \times 10^{-7} (300/T)^{0.55}$	
8	$N_2^+ + O$	\rightarrow	$NO^{+} + N$	$k_8 = 1.4 \times 10^{-10} (300/\text{T})^{0.44}$	
9	$N_2^+ + O$	\rightarrow	$O^{+} + N_{2}$	$k_9 = 1 \times 10^{-11} (300/T)^{0.23}$	
10	$N_{2}^{+} + O_{2}$	\rightarrow	$O_2^+ + N_2$	$k_{10} = 5 \times 10^{-11} (300/\text{T})^{-1}$	
11	$N_2^+ + NO$	\rightarrow	$NO^+ + N_2$	$k_{11} = 3.3 \times 10^{-10}$	
12	$N_{2}^{+} + e$	\rightarrow	N + N	$k_{12} = 1.8 \times 10^{-7} (300/\text{T})^{0.39}$	
13	$NO^{+} + e$	\rightarrow	N + O	$k_{13} = 4.2 \times 10^{-7} (300/\text{T})^{0.85}$	
14	$O(^{1}S)$	\rightarrow	$O(^{1}D) + \hbar\nu(5577Å)$	$A_{5577} = 1.06 \text{ s}^{-1}$	
15a	$O(^{1}D)$	\rightarrow	$O(^{3}P) + \hbar\nu$	$A_{6300+6364} = 9.34 \times 10^{-3} \text{ s}^{-1}$	
15b	$O(^{1}D)$	\rightarrow	$O(^{3}P) + \hbar\nu(6300\text{\AA})$	$A_{6300} = 7.1 \times 10^{-3} \text{ s}^{-1}$	
16	$O(^{1}D) + N_{2}$	\rightarrow	$O + N_2$	$k_{16} = 2.0 \times 10 - 11 \exp[107.8/T]$	
17	$O(^{1}D) + O_{2}$	\rightarrow	$O + O_2$	$k_{17} = 2.0 \times 10^{-11} \exp[67.5/T]$	
18	$O(^{1}D) + O$	\rightarrow	O + O	$k_{18} = 8.0 \times 10^{-12}$	

800 waves (400 positive k and 400 negative k) based on values of x ranging from -1000 km to +14,400 km (so that the wave number resolution is $dk = 4.36 \times 10^{-7}$ m⁻¹) and provides good spectral resolution for all the simulations.

[9] For each minor species (including ions) n_j we solve a steady state, linearized continuity equation

$$i\omega n'_j + w'_j \frac{d\overline{n}_j}{dz} + \overline{n}_j \underline{\nabla} \cdot \overrightarrow{v}'_j = P'_j - \overline{n}_j L'_j - n'_j \overline{L}_j.$$
(2)

[10] Here ω is the wave frequency, n_j is the number density of the *j*th minor species, P_j and L_j are the chemical production and loss of the *j*th species, respectively, primed quantities are fluctuations, and symbols bearing an over-bar represent the undisturbed mean state. Also, w_j' and $\nabla \cdot \vec{v}_j'$ are the minor species' perturbation vertical velocity and velocity divergence, respectively. Given an initial undisturbed mean state and the tsunami-induced gravity wave forcing, the set of continuity equations describing all minor species in the model is solved using the method described by *Walterscheid et al.* [1987]. This process is repeated for all waves in the spectrum, providing the total response to a gravity wave disturbance.

[11] For the neutral minor species (which are assumed to have the same velocity as the major gas) the values of w' and $\underline{\nabla} \cdot \vec{v}'_j$ required in (2) are the gravity wave vertical velocity and velocity divergence, respectively. The ion velocities needed to solve (2) are calculated using the approach of *McLeod* [1965], as described in detail by *Hickey et al.* [2009]. With the neutral velocity specified as $U = u\hat{i} + v\hat{j} + w\hat{k}(\hat{i},\hat{j},\hat{k}$ are unit vectors in the *x*, *y*, *z* directions, respectively, which are positive in the southward, eastward, and upward directions, respectively) the ion velocity (\vec{v}_i) is given by

$$\begin{aligned} (\omega_i^2 + v_{in}^2)\vec{v}_i' &= \{v_{in}\omega_i v' \sin I + \omega_i^2 (u' \cos I + w' \sin I) \cos I + v_{in}^2 u'\}\hat{i} \\ &+ \{v_{in}\omega_i (w' \cos I - u' \sin I) + v_{in}^2 v'\}\hat{j} \\ &+ \{-v_{in}\omega_i v' \cos I + w_i^2 (u' \cos I + w' \sin I) \sin I + v_{in}^2 w'\}\hat{k}. \end{aligned}$$
(3)

[12] Here, ω_i is the ion gyrofrequency, and v_{in} is the ionneutral collision frequency. The geomagnetic dip angle *I* is described by equation (A4) and associated text in the study of *Hickey et al.* [2009]. At the equator $I = \pi$, and the magnetic field is horizontal and northward. The neutral perturbation velocity components in (3) are obtained directly from the full-wave model. The velocity divergence is calculated by assuming plane waves in the horizontal direction with wave numbers *k* and *l* and by finite differencing the vertical component of the ion velocity: $\nabla \cdot \overline{v}'_i = -iku'_i - ilv'_i + \partial w'_i/\partial z$.

[13] The chemical reactions and production and loss rates for the ionospheric model are shown in Table 1. They are taken from the works of *Schunk and Sojka* [1996], *Rees* [1989], and *Meléndez-Alvira et al.* [1999]. The OI 6300 nightglow chemistry is also included in Table 1. It is similar to that used by *Mendillo et al.* [1997] and includes only the charge exchange reaction producing O_2^+ and the dissociative recombination reaction of O_2^+ . This assumption was shown by them to work well in reproducing observed large-scale wave structures in electron density. Other reaction rates are taken from the works of *Solomon et al.* [1987] and *Torr et al.* [1990].

[14] The OH nightglow emission model includes the chemistry shown in Table 2. These same reactions have been previously used by *Hickey et al.* [2003] to model the effects of a gravity wave packet on the chemical exothermic heating in the mesopause region and by *Huang and Hickey* [2008] to model gravity-wave-driven secular variations in the OH (8-3) nightglow. Table 2 includes options for either the (8-3) or the (6-2) emission chemistry, with an obvious notation. Here, we use the (6-2) emission parameters, but note that the results for the (8-3) emission (not shown) are very similar.

3. Results

[15] Simulations are performed using the same undisturbed mean state described by *Hickey et al.* [2009] appropriate for equatorial regions. The assumed waveform that

 Table 2. Chemical Reactions and Kinetic Parameters Used to Evaluate the OH Nightglow^a

	Reaction	Reaction Rate ^b
R ₁	$O + OH(\nu = 0) \rightarrow H + O_2$	4.0×10^{-11}
R_2	$H + O_2 + M \rightarrow HO_2 + M$	$2.1 \times 10^{-32} \exp(290/T)$
R_3	$O + HO_2 \rightarrow OH(\nu = 0) + O_2$	4×10^{-11}
R ₄	$O + O + M \rightarrow O_2 + M$	$4.7 \times 10^{-33} (300/T)^2$
R.5	$O + O_2 + M \rightarrow \tilde{O_3} + M$	$1.0 \times 10^{-34} \exp(510/T)$
R ₆	$H + O_3 \rightarrow OH^*(\nu = 8) + O_2$	$0.27 \times 1.4 \times 10^{-10} \exp(-470/T)$
R ₇	$H + O_3 \rightarrow OH^*(\nu = 0) + O_2$	$0.73 \times 1.4 \times 10^{-10} \exp(-470/T)$
R ₈	$OH^*(\nu = 8) \rightarrow OH(\nu = 3) + \hbar\nu$	0.569
R ₈	$OH^*(\nu = 6) \rightarrow OH(\nu = 2) + \hbar\nu$	1.609
R ₉	$OH^*(\nu = 6.8) + O \rightarrow H + O_2$	$2.0 \times 10^{-10} (2.5 \times 10^{-10})$
R ₁₀	$OH^*(\nu = 6.8) + O_2 \rightarrow OH(\nu - 1) + O_2$	$3.0 \times 10^{-12} (8.0 \times 10^{-12})$
R ₁₁	$OH^*(\nu = 6,8) + N_2 \rightarrow OH(\nu - 1) + N_2$	$9.1 \times 10^{-14} (7.0 \times 10^{-13})$
		× /

^aParameters are provided for the 6-2 transition and for the 8-3 transition (in parentheses).

^bRate constants are in units of cm⁶ s⁻¹ for termolecular reactions, cm³ s⁻¹ for bimolecular reactions, and s⁻¹ for emissions.

mimics a tsunami at the lower boundary is also the same, and we consider meridional wave propagation only. We first consider the results for the mesospheric OH nightglow emissions, followed by the results for the thermospheric nightglow emissions (6300 and 1356 Å). Because the thermospheric nightglow emission chemistries are tied to the ion chemistry, we also show the response of the total electron content to the gravity wave disturbance.

3.1. OH Nightglow Fluctuations

[16] The altitude profile of the mean undisturbed OH nightglow volume emission rate (VER) is shown in Figure 2. The maximum VER occurs near 87 km with a value of 1.16×10^{10} photons/s/m³. The emission layer vertical thickness has a full width at half maximum of ~10 km.

[17] The temperature perturbation due to the wave disturbance in the upper mesosphere/lower thermosphere is shown as a function of horizontal position and altitude in Figure 3. In the vicinity of the OH peak (~87 km), the maximum temperature fluctuations are at most about 1 K and occur during the first main cycle of the disturbance. The maximum perturbation temperature amplitude in the second cycle is ~0.5 K near 87 km altitude. Because the wave amplitude increases with increasing altitude, the temperature perturbations are slightly larger in the upper region of the emission layer.

[18] The resulting OH (6-2) VER fluctuations are shown as a function of horizontal position and altitude in Figure 4. The largest perturbations occur just below the layer peak (near 86 km altitude), with the largest amplitude being $\sim 1.5 \times 10^8$ photons/s/m³ or $\sim 1\%$ of the mean at this height. A phase reversal is seen in the VER fluctuations about the OH centroid height (87 km), indicative of a dominance of vertical advection in the continuity equation for the minor species (as confirmed by further numerical analysis, not shown here).

[19] Vertical integration of the VER profiles shown in Figure 4 provides the OH (6-2) brightness variation, shown relative to its respective mean value as a function of horizontal position in Figure 5. This waveform looks strikingly similar to the original lower boundary/tsunami disturbance shown in Figure 1a. The maximum brightness fluctuation occurs at $x \sim 100$ km, with an amplitude of $\sim 1.1 \times 10^6$ Rayleighs (R), which is almost 1% of the mean brightness (1.4 × 10⁸ R).

3.2. Thermospheric Nightglow Fluctuations

[20] Figure 6 shows altitude profiles of the mean undisturbed electron density, the OI 6300 Å VER and the O 1356 Å VER. The electron number density profile has been modeled as a Chapman layer with the F2 peak at 300 km altitude with a maximum number density of 10^{12} m⁻³ and with an E layer peak at 105 km altitude and a number density of 1.25×10^7 m⁻³. Implicit in this choice of parameters is the assumption of nighttime conditions at moderately high-solar activity. The maximum unperturbed O 1356 VER occurs near 301 km altitude with a value of ~7.2 × 10⁵ photons/s/m³. The full width at half maximum (FWHM) for the O 1356 Å VER profile is about 100 km. The OI 6300 Å VER peaks at 254 km altitude with a value of ~8.6 × 10⁷ photons/s/m³ and has a FWHM of 63 km.

[21] The thermospheric temperature fluctuation due to the gravity wave disturbance is shown as a function of hori-



Figure 2. Mean OH (6-2) volume emission rate $(10^6 \text{ photons/s/m}^3)$.



Figure 3. Mesosphere/lower thermosphere temperature fluctuations (in K) due to tsunami-driven gravity wave disturbance.

zontal position and altitude in Figure 7. The maximum amplitude of the leading cycle of the disturbance is \sim 130 K (or 18% of the mean) and occurs in the 200–250 km altitude region. At greater heights, the wave disturbance is strongly dissipated by viscosity and thermal conduction, and the wave amplitudes decrease with increasing height above \sim 250 km altitude while the vertical wavelength increases.

[22] The corresponding OI 6300 Å VER fluctuation is shown as a function of horizontal position and altitude in Figure 8. The largest perturbations occur near 254 km altitude (in the vicinity of the peak of the mean emission) and at $x \sim 350$ km with an amplitude of $\sim 6 \times 10^7$ photons/s/m³. The wave disturbance exhibits two strong cycles, with the leading part of the fluctuation being negative.

[23] The relative brightness fluctuations for overhead viewing are shown as a function of altitude in Figure 9. In order to provide early warning of an approaching tsunami, ground-based observations would not be made looking in the zenith but instead would be made looking obliquely through the atmosphere. We have therefore also performed



Figure 4. OH (6-2) volume emission rate fluctuations (photons/s/m³) as a function of position.



Figure 5. OH brightness fluctuations. The mean OH brightness is $\sim 1.44 \times 10^8$ R, and so the fluctuations shown above are $\sim 1\%$ brightness fluctuations.

calculations of the brightness of the OI 6300 Å emission along a line of sight inclined at an angle of ~29° from the horizontal. Our Cartesian-based model cannot account for Earth curvature effects, but these effects should be small for the large elevation angles considered here (at the ~254 km height of the peak VER, the horizontal distance is ~500 km). For overhead viewing, the maximum OI 6300 Å airglow response of ~55% occurs at a position $x \sim 350$ km, which is ~150 km retarded with respect to the original tsunami disturbance. In the case of oblique viewing, the maximum airglow response of ~30% occurs at $x \sim 0$, which is ~200 km (~15 min) ahead of the tsunami.

[24] The O 1356 Å VER fluctuations are shown as a function of horizontal position and altitude in Figure 10. The largest perturbations occur at the layer peak (near 300 km altitude), with the largest amplitude being $\sim 4 \times 10^5$ photons/s/m³ or ~55% of the mean at this height. At lower altitudes, the relative VER fluctuations are larger (76% at 250 km altitude) because of the small O 1356 Å scale height on the layer bottom side, while at greater altitudes the relative VER fluctuations are smaller (45% at 350 km altitude) as a result of viscous dissipation and the larger O 1356 Å scale height.

[25] Integration over altitude of the VER provides the O 1356 Å brightness. The O 1356 Å relative brightness perturbation (which is the fluctuation divided by the mean) is shown as a function of horizontal position in Figure 11. Also shown is the total electron content fluctuation relative to the TEC mean. The shape and phases of the two waveforms are almost identical, and both waveforms look strikingly similar to the original lower boundary/tsunami disturbance shown in Figure 1a. The maximum relative TEC and brightness fluctuations both occur at $x \sim 200$ km with values of ~33% and 43% for the TEC and O 1356 Å, respectively. Therefore, the relative ultraviolet fluctuations are about 30% greater than the relative TEC fluctuations. [26] The brightness fluctuations for oblique viewing (at $\sim 29^{\circ}$ to the horizontal) shown in Figure 11 have smaller relative amplitudes compared to those for zenith viewing. For oblique viewing, the relative perturbation amplitudes in the TEC and O 1356 Å nightglow are a factor of ~ 4 and ~ 2.5 smaller than for zenith viewing, respectively. The phase of the perturbations for oblique viewing are advanced by ~ 400 km (or 30 min) relative to overhead viewing, demonstrating that, for ground-based observations of the O 1356 Å emission, oblique viewing could provide a reasonable warning of an approaching tsunami.

4. Discussion

[27] Unlike the ionospheric response to a gravity wave, which depends on the wave orientation with respect to the geomagnetic field [*Hooke*, 1968], in the absence of mean winds the OH response to a gravity wave disturbance is independent of the direction of wave propagation [*Hickey et al.*, 1992]. The mean winds will influence the OH night-glow response, but because the phase speed of the wave disturbance (~200 m/s) significantly exceeds the mean wind speeds throughout the lower and middle atmosphere, sensitivity of the OH nightglow response to propagation direction should not be large.

[28] The OH nightglow response is ~1% of the mean for a 50 cm surface displacement amplitude at the model lower boundary. A signal of this amplitude should be discernible against the background of waves, since the former is fast (200 m/s) compared to the latter (10–50 m/s) and has a much larger (~400 km) horizontal wavelength than the latter (a few tens to a few hundred kilometers). *Hecht et al.* [2002] discussed observations of a fast (~160 m/s) disturbance in



Figure 6. Mean electron density (solid curve, m^{-3}), mean OI 6300 Å volume emission rate (long-dashed curve, photons/s/m³), and mean O 1356 Å volume emission rate (dashed triple-dotted curve, photons/s/m³).



Figure 7. Thermospheric temperature fluctuations (in K) due to tsunami-driven gravity wave disturbance.

the OH (4-2). This wave had a horizontal wavelength of 35 km, and the brightness amplitude was ~2%. The larger amplitude and shorter wavelength of this wave would make it easier to observe than a tsunami-generated gravity wave, but the observation nonetheless supports the plausibility of using the OH nightglow to observe fast moving waves associated with larger tsunami-driven events in the upper mesosphere. The OH nightglow can only be observed at night and when there is fairly good seeing conditions, which would place further constraints on the observations. However, OH nightglow measurements may provide a reason-

ably low-cost means to usefully augment other measurement techniques.

[29] Results for the oblique viewing of the OH Meinel mesospheric emission (not shown) is similar to that obtained for overhead viewing, which is mainly due to the relatively low altitude of these emissions and the thinness of the OH layer.

[30] We have used the OH (6-2) Meinel emission for this study, but several other emissions are commonly used. Broadband imaging, covering several emission bands, could



Figure 8. OI 6300 Å volume emission rate fluctuations (photons/s/m³) as a function of position.



Figure 9. OI 6300 Å relative brightness fluctuations for overhead and oblique viewing. The zenith brightness of the 6300 is 5.78×10^6 R. The brightness for the 29° line of sight is 1.19×10^7 R.

provide a larger signal than would be obtained using a single band, which would add to the usefulness of this approach.

[31] Ion diffusion has not been considered here, but its effects should not be large for the large vertical wavelength disturbance encountered here [*Occhipinti et al.*, 2006]. Also, like *Occhipinti et al.* [2006], we have found that the electron

density perturbations are dominated by dynamic effects and are not very sensitive to the chemistry in the model.

[32] We have estimated the time taken for the gravity wave disturbance to propagate to 300 km altitude (the peak height of the O 1356 Å emission) by considering the dominant wave in the spectrum (400 km wavelength and 33 min period) and calculating its vertical group velocity. The vertical group speed is ~20 m/s in the lower 100 km of the atmosphere, and it rapidly increases to ~60 m/s in the thermosphere. This implies a vertical propagation time of ~2 h between the surface and 300 km altitude. This time delay is only relevant during the initial setup phase of gravity wave production and propagation. After that the disturbance achieves a steady state in the frame of reference moving with the tsunami (see discussions by *Peltier and Hines* [1976]).

[33] We have used a large vertical displacement of 50 cm at the surface based on the Sumatra tsunami event of 26 December 2004. Because our model is linear, we can rescale the results to accommodate more modest forcing amplitudes. For a 5 cm vertical displacement at the sea surface, the OI 6300 Å and O 1356 Å fluctuations would be $\sim 3\%$ -5% of the mean (depending on viewing geometry), which may still be large enough to be observable.

[34] The northward propagation direction at the equator considered here provides the greatest electron (and hence, O 1356 Å and OI 6300 Å) response [*Hooke*, 1968; *Occhipinti et al.*, 2008; *Hickey et al.*, 2009]. For zonally propagating gravity waves at the equator, the response is substantially reduced, which would make the gravity waves difficult to observe in both the TEC and in the OI 6300 Å and O 1356 Å measurements. However, further calculations (not shown here) and those of *Occhipinti et al.* [2008] show that the electron response for zonal propagation is comparable to that for meridional propagation at midlatitudes. Hence, except at very low latitudes, tsunami-driven fluctuations in TEC and in the thermospheric nightglow emissions considered here should be observable for zonal propagation. Addi-



Figure 10. O 1356 Å volume emission rate fluctuations (photons/s/m³) as a function of position.



Figure 11. Total electron content (TEC) fluctuations relative to the mean TEC and the O 1356 Å relative brightness fluctuation. For zenith viewing, the mean total electron content is 16.6 TECU and the mean O 1356 Å brightness is $\sim 7.82 \times 10^4$ R. For the oblique viewing considered these means are increased by a factor of ~2.06.

tionally, the midlatitude TEC and OI 6300 Å and O 1356 Å responses should be less dependent on the direction of tsunami (and wave) propagation.

[35] We have only considered nighttime ion/electron chemistry and hence our brightness calculations for the OI 6300 Å and O 1356 Å emission are only appropriate for the nightglow. Future research will focus on including daytime photochemistry in the modeling of tsunami-driven wave effects on the ionosphere/visible/FUV airglow.

5. Conclusions

[36] Simulations obtained using a spectral full-wave model have shown that relative fluctuations in the OI 6300 Å and O 1356 Å emissions due to a gravity wave disturbance forced by a tsunami at the lower boundary of the atmosphere should be at least as large as those expected to occur in the total electron content. Monitoring of the emissions from space-based platforms (especially for the O 1356 Å far ultraviolet emission) should provide useful additional information on tsunami-driven atmospheric waves. Satellite measurements of the O 1356 Å, most probably made in the nadir, could provide worldwide coverage of tsunami effects in the ionosphere. Whether or not such observations would be useful in the early detection of tsunamis would depend, amongst other things, on the number of satellites and the distribution of orbits, how frequently the satellite data are relayed to ground stations, and the time taken to analyze and distribute the data. The capability to remotely sense these disturbances using the FUV airglow may be an important

technique appropriate for the monitoring of planetary atmospheres in order to learn more about their atmospheric waves and their origin [e.g., *Garcia et al.*, 2009]. In the case of ground-based measurements (either OI 6300 Å airglow or GPS measurements of TEC), oblique viewing could provide early warning of an approaching tsunami. Finally, we conclude that, while it may be possible to observe the effects of tsunami-driven gravity waves in mesospheric airglow emissions, such observations would most likely provide significantly less warning of an approaching tsunami than would observations of thermospheric airglow emissions, because the much higher altitude of the latter makes them observable to greater distances.

[37] Acknowledgments. M.P.H. was supported by the National Science Foundation under grants ATM-0408407 and ATM-0639293. G.S. was supported by a grant from the NASA Planetary Atmospheres program. R.L.W. was supported by NASA grant NNX08AM13G.

[38] Robert Lysak thanks Giovanni Occhipinti and another reviewer for their assistance in evaluating this paper.

References

- Artru, J., V. Ducic, H. Kanamori, P. Lognonne, and M. Murakami (2005a), Ionospheric detection of gravity waves induced by tsunamis, *Geophys. J. Int.*, 160, 840–848.
- Artru, J., P. Lognonne, G. Occhipinti, F. Crespon, R. Garcia, E. Jeansou, and M. Murakami (2005b), Tsunami detection in the ionosphere, *Space Res. Today*, 163, 23–27.
- DeMajistre, R., L. J. Paxton, and D. Bilitza (2007), Comparison of ionospheric measurements made by digisondes with those inferred from ultraviolet airglow, *Adv. Space Res.*, 39, 918–925.
- Garcia, R. F., P. Drossart, G. Piccioni, M. Lopez-Valverde, and G. Occhipinti (2009), Gravity waves in the upper atmosphere of Venus revealed by CO₂ non local thermodynamic equilibrium emissions, *J. Geophys. Res.*, 114, E00B32, doi:10.1029/2008JE003073.
- Hecht, J. H., and R. L. Walterscheid (1991), Observations of the OH Meinel (6,2) and O₂ atmospheric (0,1) nightglow emissions from Maui during the Aloha-90 campaign, *Geophys. Res. Lett.*, 18, 1341–1344.
- Hecht, J. H., R. L. Walterscheid, M. P. Hickey, R. J. Rudy, and A. Z. Liu (2002), An observation of a fast external atmospheric acoustic-gravity wave, J. Geophys. Res., 107(D20), 4444, doi:10.1029/2001JD001438.
- Hickey, M. P. (1993), A review of gravity wave interactions with selected mesopause nightglow emissions, *Trends Geophys. Res.*, 2, 487–499.
- Hickey, M. P., G. Schubert, and R. L. Walterscheid (1992), Seasonal and latitudinal variations of gravity wave-driven fluctuations in OH nightglow, J. Geophys. Res., 97, 14,911–14,922.
- Hickey, M. P., T.-Y. Huang, and R. L. Walterscheid (2003), Gravity wave packet effects on chemical exothermic heating in the mesopause region, *J. Geophys. Res.*, 108(A12), 1448, doi:10.1029/2002JA009363.
- Hickey, M. P., G. Schubert, and R. L. Walterscheid (2009), The propagation of tsunami-driven gravity waves into the thermosphere and ionosphere, J. Geophys. Res., 114, A08304, doi:10.1029/2009JA014105.
- Hines, C. O. (1972), Gravity waves in the atmosphere, Nature, 239, 73-78.
- Hines, C. O., and D. W. Tarasick (1987), On the detection and utilization of gravity waves in airglow studies, *Planet. Space Sci.*, 35, 851–866.
- Hooke, W. H. (1968), Ionospheric irregularities produced by internal atmospheric gravity waves, J. Atmos. Terr. Phys., 30, 795–823.
- Huang, T.-Y., and M. P. Hickey (2008), Secular variations of OH nightglow emission and of the OH intensity-weighted induced by gravity wave forcing in the MLT region: On the distribution and transport of minor species, Adv. Space Res., 41, 1477–1486, doi:10.1016/j. asr.2007.10.020.
- Kubota, M., H. Fukunishi, and S. Okano (2001), Characteristics of medium- and large-scale TIDs over Japan derived from OI 630-nm nightglow observations, *Earth Planets Space*, 53, 741–751.
- McLeod, M. A. (1965), Sporadic E theory: I. Collision-geomagnetic equilibrium, J. Atmos. Sci. , 23, 96–109.
- Meléndez-Alvira, D. J., R. R. Meier, J. M. Picone, P. D. Feldman, and B. M. McLaughlin (1999), Analysis of the oxygen nightglow measured by the Hopkins Ultraviolet Telescope: Implications for ionospheric partial radiative recombination rate coefficients, J. Geophys. Res., 104, 14,901– 14,913.

- Mendillo, M., J. Baumgardner, D. Nottingham, J. Aarons, B. Reinisch, J. Scali, and M. C. Kelley (1997), Investigations of thermosphericionospheric dynamics with 6300-A images from the Arecibo Observatory, J. Geophys. Res., 102, 7331–7343.
- Occhipinti, G., P. Lognonne, E. A. Kherani, and H. Hebert (2006), Threedimensional waveform modeling of ionospheric signature induced by the 2004 Sumatra tsunami, *Geophys. Res. Lett.*, 33, L20104, doi:10.1029/ 2006GL026865.
- Occhipinti, G., E. A. Kherani, and P. Lognonne (2008), Geomagnetic dependence of ionospheric disturbances induced by tsunamigenic internal gravity waves, *Geophys. J. Int.*, 173, 753–765.
- Paxton, L. J., D. Morrison, D. J. Strickland, M. J. McHarg, Y. Zhang, B. Wolven, H. Kil, G. Crowley, A. B. Christensen, and C.-I. Meng (2003), The use of far ultraviolet remote sensing to monitor space weather, *Adv. Space Res.*, 31(4), 813–818.
- Peltier, W. R., and C. O. Hines (1976), On the possible detection of tsunamis by a monitoring of the ionosphere, *J. Geophys. Res.*, 81, 1995–2000.
- Rees, M. H. (1989), *Physics and Chemistry of the Upper Atmosphere*, Cambridge Univ. Press, New York.
- Schubert, G., and R. L. Walterscheid (1988), Wave-driven fluctuations in OH nightglow from an extended source region, J. Geophys. Res., 93, 9903–9915.
- Schunk, R. W., and J. J. Sojka (1996), USU model of the global ionosphere, in *STEP: Handbook of Ionospheric Models*, edited by R. W. Schunk, pp. 153–171, Utah State University/SCOSTEP, Logan.

- Sivjee, G. G., R. L. Walterscheid, J. H. Hecht, R. M. Hamwey, G. Schubert, and A. B. Christensen (1987), Effects of atmospheric disturbances on polar mesopause airglow OH emissions, J. Geophys. Res., 92, 7651–7656.
- Sobral, J. H. A., C. Carlson, D. T. Farley, and W. E. Shwartz (1978), Nighttime dynamics of the F region near Arecibo as mapped by airglow features, J. Geophys. Res., 83, 2561–2566.
- Solomon, S. C., P. B. Hays, and V. J. Abreu (1987), The auroral 6300 Å emission: Observations and modeling, J. Geophys. Res., 93, 9867–9882.
- Swenson, G. R., and S. B. Mende (1994), OH emission and gravity waves (including a breaking wave) in all-sky imagery from Bear Lake, UT, *Geophys. Res. Lett.*, 21, 2239–2242.
- Taylor, M. J., D. N. Turnbull, and R. P. Lowe (1991), Coincident imaging and spectrometric observations of zenith OH nightglow structure, *Geophys. Res. Lett.*, 18, 1349–1352.
- Torr, M. R., D. G. Torr, P. G. Richards, and S. P. Yung (1990), Mid- and low-latitude model of the thermospheric emissions 1. O⁺(²P) 7320 Å and N⁺₂(²P) 3371 Å, *J. Geophys. Res.*, *95*, 21,147.
- Walterscheid, R. L., G. Schubert, and J. Straus (1987), A dynamical-chemical model of wave-driven fluctuations in the OH nightglow, J. Geophys. Res., 92, 1241–1254.

M. P. Hickey, Department of Physical Sciences, Embry-Riddle Aeronautical University, 600 S. Clyde Morris Boulevard, Daytona Beach, FL 32114, USA. (hick0b5@erau.edu)

G. Schubert and R. L. Walterscheid, Space Science Applications Laboratory, The Aerospace Corporation, Los Angeles, CA 90009, USA.