Observations and Interpretation of Gravity Wave Induced Fluctuations in the O I (557.7 nm) Airglow

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Observations and interpretation of gravity wave induced fluctuations in the O I (557.7 nm) airglow

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Abstract. Observations of fluctuations in the intensity and temperature of the O I (557.7 nm) airglow taken at Arecibo in 1989 are reported and interpreted on the assumption that they are caused by gravity waves propagating through the emission layer. The data give the magnitude of Krassovsky's ratio as 3.5 ± 2.2, at periods between about 5 and 10 hours. Comparison with theory shows that the gravity waves responsible for the measured airglow variations must have long wavelengths of several thousand kilometers. The observed phases of Krassovsky's ratio are in good agreement with theoretically predicted values at the long wavelengths and large periods for about half the cases. In the other cases, observed phases are near -180°, suggesting that the waves responsible for the airglow fluctuations have experienced strong reflections in the emission layer. The observations emphasize the importance of knowing the full altitude profiles of temperature and winds for extraction of wave information from the airglow fluctuations.

1. Introduction

Rocket, ground-based, and satellite observations of the O I (557.7 nm) or O(S) airglow have been a primary source of information about the structure and dynamics of the upper mesosphere at altitudes around 97 km, where the emission rate of the atomic oxygen green line is a maximum. The literature on the O I (557.7 nm) airglow has mostly focused on elucidating the relevant photochemistry and deducing the altitude profile of atomic oxygen number density from height variations of the emission rate [e.g., Chapman, 1931; Barth, 1961, 1964; Slinger and Black, 1977; Witt et al., 1979; Thomas et al., 1979; Thomas, 1981; Bates, 1981, 1988, 1992; Torr et al., 1985; McDade et al., 1986; McDade and Llewellyn, 1986; Greer, 1988; Murtagh et al., 1990; Sharp, 1991; Kita et al., 1992; Gobbi et al., 1992; Lopez-Gonzalez et al., 1992a, b; Melo et al., 1997].

The above efforts notwithstanding, O(S) observations remain a largely untapped resource of information about gravity waves in the mesosphere. In this paper we investigate how measurements of fluctuations in the O I (557.7 nm) airglow can be used to quantitatively infer the characteristics of the gravity waves driving the fluctuations. Our approach is similar to that of previous studies of gravity wave driven fluctuations in the OH [Walterscheid et al., 1987; Schubert and Walterscheid, 1988; Schubert et al., 1991; Hickey et al., 1992; Makhlouf et al., 1995] and O2 atmospheric (0-1) [Hickey et al., 1993a] airglow which make use of Krassovsky's [1972] ratio to relate the fluctuations in airglow intensity to the fluctuations in temperature of the emitting gas. We summarize the theory needed to apply Krassovsky's ratio to the O I (557.7 nm) airglow fluctuations and use this theory to interpret measurements of the airglow temperature and intensity fluctuations.

The measurements of the O(S) airglow were made at the Arecibo Observatory in Puerto Rico during the Arecibo Initiative for Dynamics of the Atmosphere (AIDA) 1989 campaign. We used a single-etalon Fabry-Perot interferometer to obtain a ground-based CCD all-sky imager to infer a value for the vertical eddy diffusion coefficient due to gravity waves in the 90–100 km altitude interval.

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2. Theory

Theories that describe how fluctuations in the intensity of the O(S) airglow are produced by gravity wave disturbances
Makhlouf et al. [1997, 1998]. Here we follow and further develop the approach of Hickey et al. [1997, 1998]. Here we follow and further develop the approach of Hickey et al. [1997].

The ground-based spectrophotometric observations provide data on the vertically integrated airglow intensity \( I = \langle I \rangle + \langle I' \rangle \) (\( I \) is intensity, angle brackets denote integration over the height of the emission region, the overbar refers to a time-averaged background state, and the prime denotes departures therefrom) and the intensity-weighted temperature \( (T_I) = \langle T_I \rangle + \langle T'_I \rangle \) (\( T_I \) = \( f dz T_I / I \)) (\( z \) is the vertical coordinate). The transfer function \( (\eta) \) between the gravity wave driven fluctuations in intensity-weighted temperature and the fluctuations in vertically integrated intensity is Krassovsky's ratio [Krassovsky, 1972, Schubert et al., 1991]

\[
(\eta) = \frac{\langle I' \rangle/\langle I \rangle}{\langle T'_I \rangle/\langle T_I \rangle} \tag{1}
\]

Airglow observations provide values for the amplitude and phase of the complex Krassovsky's ratio according to (1). Values of \( (\eta) \) from dynamical-chemical theory can be compared with the observational values of \( (\eta) \) to infer information about the airglow chemistry, the gravity wave field, the winds, and the atmospheric structure, as has been done in the above cited studies of the OH airglow.

The intensity of the atomic oxygen green line nightglow is directly proportional to the number density of the emitting species \( n(O^1S) = \bar{n}O(1S) + n'(O(1S)) \). Therefore, we can write the numerator of (1) as

\[
\langle I' \rangle = \frac{\langle n'(O^1S) \rangle}{\langle T'_I \rangle} \tag{2}
\]

The determination of the fluctuation in the number density of \( O(1S) \) uses the linearized continuity equation

\[
i_{O(1S)}' = P' - L' - \frac{\partial \bar{n}}{\partial z} - \bar{n} \nabla \cdot \mathbf{v}' \tag{3}
\]

where \( P' \) and \( L' \) are the perturbations in the chemical production and loss of \( O(1S) \), respectively; \( n' \) is the \( O(1S) \) number density perturbation about its mean value \( \bar{n} \); \( w' \) is the gravity wave vertical velocity component; \( \mathbf{v}' \) is the gravity wave velocity divergence; and \( \omega \) is the angular frequency of the gravity wave. All perturbation quantities are assumed to vary as \( \exp(\omega t - kz) \), where \( k \) is the horizontal wavenumber in the \( x \) direction. Specification of the chemical production and loss terms in (3) requires identification of the chemical reactions involved in the \( O(1S) \) chemistry. These reactions involve other minor species whose number densities must also be determined. Fluctuations in the number densities of all minor species are determined from continuity equations identical to (3).

The \( O(1S) \) chemistry is given in Table 1. We have assumed, in accordance with Bates [1988], that the production of \( O(1S) \) is by the two-step Barth process in which the intermediate state is \( O_2(c^1\Sigma_u^+) \). The reaction rates employed here are those given by Torr et al. [1985], except for the branching ratios related to the production of \( O_2(c^1\Sigma_u^+) \) and \( O(1S) \), which are taken from Lopez-Gonzalez et al. [1992a, b]. The Barth process couples the species \( O, O_2(c^1\Sigma_u^+) \), and \( O(1S) \), and the fluctuations in the number densities of these constituents must be determined by the simultaneous solution of (3) for each of these species. The specific form of these equations is given in (A3)–(A5) of Hickey et al. [1997].

The solution of the linearized continuity equations for the minor species \( O, O_2(c^1\Sigma_u^+) \), and \( O(1S) \) depends on the complex dynamical factors \( f_1, f_2, \) and \( f_3 \) that relate the velocity divergence \( \nabla \cdot \mathbf{v}' \), the vertical velocity perturbation \( w' \), and the major gas density perturbation \( n'(M) \) to the temperature perturbation \( T' \)

\[
\nabla \cdot \mathbf{v}' = f_1 \frac{T'}{T} \tag{4}
\]

\[
w' = f_2 \frac{T'}{T} \tag{5}
\]

\[
n'(M) = f_3 \frac{T'}{T} \tag{6}
\]

A model for the upward propagation of gravity waves is needed to determine the dynamical factors \( f_1, f_2, \) and \( f_3 \) and the altitude variation of \( T'/T \). We use the full wave, gravity wave model of Hickey et al. [1994, 1995, 1997, 1998] for the propagation of nonhydrostatic, linear gravity waves from the troposphere up to a maximum altitude of 500 km. It includes dissipation due to eddy processes in the lower atmosphere and molecular processes (viscosity, thermal conduction, and ion drag) in the upper atmosphere, height variations of the mean temperature and horizontal winds, and Coriolis forces. The model accurately describes the propagation of gravity waves in an inhomogeneous atmosphere.

### Table 1. Chemistry of the \( O(1S) \) Nightglow Model

<table>
<thead>
<tr>
<th>Reaction</th>
<th>Rate of Reaction*</th>
</tr>
</thead>
<tbody>
<tr>
<td>( O + O + M \rightarrow O_2 + M )</td>
<td>( k_1 = 4.7 \times 10^{-33} ) ( \text{(300/T)}^2 )</td>
</tr>
<tr>
<td>( O + O + M \rightarrow O_2(c^1\Sigma_u^+) + M )</td>
<td>( k = \mathcal{C} k_1, \mathcal{C} = 0.03 )</td>
</tr>
<tr>
<td>( O_2(c^1\Sigma_u^+) + O \rightarrow O_2(b^3\Pi_g) + O )</td>
<td>( k_2 = 5.0 \times 10^{-13} )</td>
</tr>
<tr>
<td>( O_2(c^1\Sigma_u^+) + O \rightarrow O_2(a^3\Pi_g) + O )</td>
<td>( k_3 = 3.0 \times 10^{-11} )</td>
</tr>
<tr>
<td>( O_2(c^1\Sigma_u^+) + O \rightarrow O_2 + O(1S) )</td>
<td>( k = 5 \times 10^{-11} )</td>
</tr>
<tr>
<td>( O_2(c^1\Sigma_u^+) + O \rightarrow O_2 + O(1S) )</td>
<td>( k = 5 \times 10^{-12} )</td>
</tr>
<tr>
<td>( O_2(c^1\Sigma_u^+) + O \rightarrow O_2 + O(1S) )</td>
<td>( k = 5 \times 10^{-13} )</td>
</tr>
<tr>
<td>( O_2(c^1\Sigma_u^+) + O \rightarrow O_2 + O(1S) )</td>
<td>( k = 5 \times 10^{-14} )</td>
</tr>
<tr>
<td>( O_2(c^1\Sigma_u^+) + O \rightarrow O_2 + O(1S) )</td>
<td>( k = 5 \times 10^{-15} )</td>
</tr>
</tbody>
</table>

*Units are \( s^{-1}, \text{cm}^{-3} \text{~s}^{-1} \), and \( \text{cm}^{-6} \text{~s}^{-1} \) for unimolecular, bimolecular, and termolecular reactions, respectively.

3. Inputs and Derived Quantities for the Basic State

Mean state quantities required for the full wave computations are provided by the mass spectrometer/incoherent scatter (MSIS-90) model [Hedin, 1991]. The altitude profile of mean state temperature is shown in Figure 1. The mesopause is at an altitude of 98 km where the temperature is 181.4 K. Mean state major gas number density \( \bar{n}(M) \) versus altitude is shown in Figure 2. Major gas number density decreases nearly exponentially with height in the altitude interval 75–110 km with an approximate scale height \( H \) of 6 km. Momentum and thermal diffusivities are plotted as a function of altitude in Figure 3. The molecular coefficients of viscosity \( \mu_m \) and thermal conductivity \( k_m \) are taken from Rees [1989] and are used to calculate the molecular momentum diffusivity \( \eta_m = \mu_m / \rho \) and molecular thermal diffusivity \( k_m / \rho C_p \). The eddy momentum diffusivity \( \eta_e \) approximates that given by Strobel [1989]. The eddy thermal diffusivity \( k_e \) is calculated from the eddy momen-
Figure 1. Altitude profile of mean state temperature from the MSIS-90 model [Hedin, 1991].

Figure 2. Altitude profile of mean state major gas number density from the MSIS-90 model [Hedin, 1991].

Figure 3. Height profiles of momentum \( \eta \) and thermal \( \kappa \) diffusivities.

Figure 4. Height profiles of mean state minor species number densities \( \bar{n}(O), \bar{n}(O_2(c^1 \Sigma_u^-)), \) and \( \bar{n}(O(^1S)) \). \( \bar{n}(O) \) is from the MSIS-90 model for April 10 at 18°N [Hedin, 1991]. The values of \( \bar{n}(O_2(c^1 \Sigma_u^-)) \) and \( \bar{n}(O(^1S)) \) are from (7) and (8), respectively.

The increase of both momentum and thermal diffusivity with height between 75 and 90 km altitude is thus due to the dominance, and increase with height, of both eddy diffusivities in this altitude range. The increase of both diffusivities with height above the mesopause is due to the dominance, and increase with height, of the molecular diffusivities.

The solution of the linearized continuity equations for the minor species \( O, O_2(c^1 \Sigma_u^-), \) and \( O(^1S) \) also requires mean state values for \( \bar{n}(O), \bar{n}(O_2(c^1 \Sigma_u^-)), \) and \( \bar{n}(O(^1S)) \). We employ the altitude profile of undisturbed atomic oxygen density derived from the MSIS-90 model for April 10 at 18°N for approximately local midnight conditions (Figure 4). At an altitude of 97.9 km, \( \bar{n}(O) \) is a maximum and varies little with height in the altitude range 90–100 km. The chemistry of Table 1 allows us to determine \( \bar{n}(O_2(c^1 \Sigma_u^-)) \), and \( \bar{n}(O(^1S)) \) as
Figure 5. Altitude profiles of the background intensities of the O(1S) and O₂(″Σg″) airglows.

The O(1S) emission intensity resulting from the chemical scheme of Table 1 is shown in Figure 5 \( I \propto A_{3577}[O(1S)] \). The O(1S) emission layer peaks near 95 km altitude with a full width at half maximum of about 9 km. The O₂(″Σg″) emission intensity resulting from the chemical scheme of Table 1 is also shown in Figure 5 \( I \propto A_{4310}[O₂(″Σg″)] \). The O₂(″Σg″) emission layer peaks near 93.6 km altitude with a full width at half maximum of about 10.8 km.

4. Dynamical Factors \( f_1, f_2, \) and \( f_3 \): A Comparison of Full Wave and WKB Results

As discussed in section 3, the dynamical effects of gravity waves on airglow intensity are controlled by the factors \( f_1, f_2, \) and \( f_3 \). In this section we discuss how these factors vary with gravity wave period. We also discuss how the dynamical factors, computed using the full wave (FW) theory, differ from the dynamical factors based on the WKB approximation employed in many of our previous papers [e.g., Schubert et al., 1991; Hickey et al., 1993a]. The factors are shown as a function of period in Figure 6 for a horizontal wavelength \( \lambda_x \) of 1000 km evaluated at an altitude of 95 km. Results for both the FW theory and WKB approximation are included. In earlier work, there were gaps in the plotted results where the waves were purely evanescent, i.e., \( k_x \) purely imaginary [Waterscheid et al., 1987; Schubert et al., 1988]. The absence of similar gaps in the present plots is due to wave dissipation. With dissipation \( k_x \) has a real (propagating) component in the gap regions. We will see later that the airglow observations suggest that we are dealing with waves having \( \lambda_x \) of order 1000 km.

The magnitude of \( f_1 \) generally decreases with increasing period from a value near \( 10^{-1} \) at a period of 100 s to a value between \( 10^{-3} \) and \( 10^{-4} \) at a period of \( 10^2 \) s. The factor \( f_1 \) is almost pure imaginary (it has a phase of \(-90°\) except at periods of \( 10^2 \) to \( 10^3 \) s, when its phase differs by a small amount from \(-90°\), and at periods in and near a period interval of wave evanescence, which occurs at periods between \( 10^2 \) and \( 10^3 \) s).

The WKB approximation does a reasonable job of representing the FW \( f_1 \) except at evanescent periods wherein the WKB approximation fails to account for the nonmonotonic behavior of \( f_1 \). The magnitude of \( f_1 \) versus period shows nonmonotonic behavior at periods in and around the evanescent period interval. Peaks and troughs in \( |f_1| \) versus period occur at the ends of the evanescent period interval.

At periods in excess of about \( 3 \times 10^4 \) s (for \( \lambda_x = 1000 \) km) or at wave phase speeds smaller than about 50 m s\(^{-1}\), the vertical wavelength of the gravity wave is small enough for severe damping to occur. This is illustrated in Figure 7 which compares the ratio of the wave kinetic energy (KE) at 100 km altitude to that at 85 km altitude as a function of wave period on the basis of the FW description of wave propagation. It is seen that dissipation starts to become important at periods of
about $2 \times 10^4$ s (phase speeds of about 50 m s$^{-1}$). At a given phase speed, additional calculations show that dissipation is more severe for waves with larger horizontal wavelengths. Waves with phase speeds of about 10 m s$^{-1}$ are highly dissipated with more than 99.999% of the energy lost between 85 and 100 km altitude.

The magnitude of $f_2$ also generally decreases with increasing period except for more complex behavior associated with the period interval of evanescence; the WKB approximation works rather well in calculating $f_2$ at most periods, but it fails in and around the evanescent region (Figure 6a). The FW calculation of $f_2$, as is the case with $f_1$, is influenced by severe wave damping for periods in excess of about $3 \times 10^4$ s. The factor $f_3$ is essentially of unity magnitude at long gravity wave periods. In the evanescent period interval, $f_3$ differs substantially from unity. The WKB calculations of $f_2$ and $f_3$ do not do well at evanescent periods.

5. Krassovsky’s Ratio

Figure 8 shows the amplitude and phase of Krassovsky’s ratio $\langle \eta \rangle$ for the O($^1S$) airglow as a function of period for a horizontal wavelength of 1000 km. Both FW and WKB results are shown. For periods greater than about $3 \times 10^3$ s, the amplitude of $\langle \eta \rangle$ varies from about 4 to 30 and the phase of $\langle \eta \rangle$ varies between about $+30^\circ$ and $-60^\circ$ except near the longest periods plotted where the full wave phase rotates rapidly through large positive values. For the most part, computation of $\langle \eta \rangle$ using the WKB approximation for the waves yields results in approximate agreement with those of the FW calculation. Differences in the magnitude of $\langle \eta \rangle$ between the FW and WKB models are about 20% at a period of $10^4$ s (phase speed of 100 m s$^{-1}$), with the FW results being the larger.

The theory discussed above for determining Krassovsky’s ratio for the O($^1S$) airglow can also provide $\langle \eta \rangle$ for the O$_2$ atmospheric airglow. The O$_2$ atmospheric airglow at 864.5 nm results from the decay of O$_2$ ($b^1\Sigma^+_g$) (Table 1). A comparison of $\langle \eta \rangle$ versus period at a horizontal wavelength of 1000 km for the O($^1S$) and O$_2$ atmospheric airglows is shown in Figures 9a and b. The Krassovsky ratios for the two airglows are very similar, though there are some quantitative differences in detail. If both airglows are described by the chemistry in Table 1, then observations of both signals provide essentially redundant information about their source region and the driving gravity wave field.

In sections 6 and 7 we discuss observations of intensity and temperature for the O($^1S$) airglow and the values of Krassovsky’s ratio inferred therefrom. We then compare the observed values of Krassovsky’s ratio with values of Krassovsky’s ratio from our theoretical model and discuss the implications of the comparison.

6. Instrument Description and Observations

The airglow data presented in this paper were obtained with a Fabry-Perot interferometer (FPI) located at the Arecibo
tures of the F region. To determine line-of-sight components is a sensitivity to the lower thermospheric winds and tempera-
tures without a contamination from the winds and tempera-
tures <20% to the total 557.7 nm emission rate. The result,
ally, the O(S) emission from the upper thermosphere con-
trates against the hotter F region component within the spec-
tral window of the instrument, which then manifests itself as a
continuum or constant level in the measurements. Addition-
ally, the O(S) 557.7 nm emission was scanned twice in wavelength
(both increasing and decreasing pressure), usually sampling 12
points across the line with a total integration time of 14 s per
spectral point. This yields about 2.8 min between line-of-sight
measurements and roughly 25 min resolution for the deriv-
ation of the full vector components.

For typical green line emission rates of about 100 R, the
integration time is sufficient to collect some 2000 photocounts
under the measured line profile, resulting in statistical uncer-
tainties of <10%. However, these errors increase with dimin-
ished emission strengths. The wind, temperature, intensity, and
background level are determined by a nonlinear least squares
fit to the near-Gaussian line shape, a procedure which also
determines the uncertainties in the parameters from the re-
duced chi-square of the fit. The errors were also used as
weighting factors in the analysis that determines the vector
components.

We applied the analysis of Burnside et al. [1981], who deter-
mined F region winds and temperature from O(^3D) 630.0 nm
observations, to our 557.7 nm spectral measurements. For
winds the method assumes that the components of the hori-
Zontal neutral wind velocity can be represented by a Taylor
expansion about a point directly above the observatory. Only
linear terms in the expansion are preserved; that is, we in-
cluded only the mean flow and the constant horizontal velocity
gradients. For temperatures and intensities we both retained
the line-of-sight components and also averaged the data in
each map to compare with the theory.

For absolute intensity calibration, photometric observations
of the 557.7 nm emission intensity were also made in the zenith
with a higher time resolution (30 s) than what was used for the
Fabry-Perot interferometer. The measured overhead 557.7 nm
intensity was determined against a 14C standard source, and
this in turn, was used to cross calibrate the signal measured by
the FPI. The relative error of the photometric signal is <1%
for emission strengths ≈100 R. Further details of the Arecibo
instruments, the observational methods, and the analysis tech-
niques used are described by Burnside et al. [1981].

Temperature and intensity data as a function of time were
detrended by the removal of a best quadratic fit. The de-
trended data were filtered, interpolated to an evenly spaced grid, smoothed, and windowed. The filtering involved fitting
the unequally spaced data with a Fourier series by least squares
down to the mean Nyquist frequency. The series was truncated
at high frequency to preclude spectral contamination due to
large point-to-point oscillations in the data. This also serves as
an antialiasing measure for the smoothing step. Smoothing was
done by a simple three-point averaging. A Hanning data win-
don was then applied to the smoothed data. The smoothed
data served as input to the spectral analysis. Both temperature
and intensity power spectra were smoothed with a Bartlett
spectral window with 6 degrees of freedom (bandwidth of
1.5/(τ/2), where τ is the length of the data set for a given night).
The above data analysis procedure removes all waves with
periods less than about an hour.

Data that yielded reliable estimates of Krassovsky's ratio are
shown in Figure 10. All data shown have been azimuthally

Figure 9. (a) Amplitude and (b) phase of Krassovsky's ratio (η)
versus period for the O(^1S) and O_2 atmospheric airglows.
Results are presented for a horizontal wavelength λ_0 of 1000
km using the full wave FW theory.
Figure 10. Intensity and temperature fluctuations in observations of the O(1S) airglow carried out at Arecibo in 1989. The solid lines connect data that were detrended by the removal of a best quadratic fit. The dashed curves show the final results of filtering, interpolating, smoothing, and windowing. Dates in the upper left of each panel are in the format day, month, year.
Table 2. Values of Krassovsky's Ratio ($\langle \eta \rangle$) From Observations of O I Airglow

<table>
<thead>
<tr>
<th>Case</th>
<th>Period, hours</th>
<th>Amplitude of $\langle \eta \rangle$</th>
<th>Phase of $\langle \eta \rangle$, deg</th>
<th>Coherence</th>
</tr>
</thead>
<tbody>
<tr>
<td>March 30, 1989</td>
<td>5.7</td>
<td>2.2</td>
<td>-42.</td>
<td>0.7</td>
</tr>
<tr>
<td>April 7, 1989</td>
<td>10.0</td>
<td>3.1</td>
<td>-46.</td>
<td>0.8</td>
</tr>
<tr>
<td>April 7, 1989</td>
<td>8.0</td>
<td>3.0</td>
<td>-58.</td>
<td>0.6</td>
</tr>
<tr>
<td>April 8, 1989</td>
<td>10.1</td>
<td>1.3</td>
<td>-124.</td>
<td>0.8</td>
</tr>
<tr>
<td>April 8, 1989</td>
<td>8.0</td>
<td>1.5</td>
<td>-117.</td>
<td>0.75</td>
</tr>
<tr>
<td>April 9, 1989</td>
<td>9.9</td>
<td>2.1</td>
<td>-154.</td>
<td>0.9</td>
</tr>
<tr>
<td>April 9, 1989</td>
<td>4.9</td>
<td>2.2</td>
<td>-144.</td>
<td>0.9</td>
</tr>
<tr>
<td>April 9, 1989</td>
<td>3.3</td>
<td>2.0</td>
<td>-127.</td>
<td>0.8</td>
</tr>
<tr>
<td>May 1, 1989</td>
<td>9.3</td>
<td>7.0</td>
<td>-155.</td>
<td>0.85</td>
</tr>
<tr>
<td>May 1, 1989</td>
<td>4.7</td>
<td>7.6</td>
<td>-150.</td>
<td>0.8</td>
</tr>
<tr>
<td>May 2, 1989</td>
<td>9.5</td>
<td>6.5</td>
<td>-6.0</td>
<td>0.6</td>
</tr>
<tr>
<td>May 2, 1989</td>
<td>4.7</td>
<td>5.3</td>
<td>11.</td>
<td>0.7</td>
</tr>
<tr>
<td>May 8, 1989</td>
<td>7.4</td>
<td>1.4</td>
<td>-3.7</td>
<td>0.75</td>
</tr>
</tbody>
</table>

7. Comparison of Observed and Theoretical Values of Krassovsky's Ratio

The observations of Krassovsky's ratio pertain to long periods between about 5 and 10 hours. Theoretical values of Krassovsky's ratio at periods of 5 and 10 hours are shown in Figure 11 as a function of horizontal wavelength. Since the long-period value of $|\langle \eta \rangle|$ is $3.5 \pm 2.2$, it is clear from Figure 11a that the observed waves must have horizontal wavelengths of several thousand kilometers. At these long wavelengths, there is little variation of $|\langle \eta \rangle|$ with period in the period range of 5 to 10 hours. At wavelengths of several thousand kilometers and for periods of 5 to 10 hours, Figure 11b shows that the phase of $\langle \eta \rangle$ is between about $-50^\circ$ and $-25^\circ$. These values are in good agreement with the group of phase observations between about $10^\circ$ and $-60^\circ$. However, the group of phase observations between about $-125^\circ$ and $-155^\circ$ must represent waves reflected in the emission layer; reflections not accounted for in the theoretical calculations.

8. Summary and Discussion

From the observations of Krassovsky's ratio and the theory of wave-driven fluctuations in the O I airglow discussed above, it can be concluded that the gravity waves responsible for the measured airglow variations must have long wavelengths of several thousand kilometers. The near $-180^\circ$ phase of $\langle \eta \rangle$ for a subset of the observations is indicative of wave reflection in the emission layer. Wave reflection can result from altitude variations of temperature or wind velocity, but reflections have not been fully accounted for in the theoretical calculations.

![Figure 11](image-url)
calculations because the observations do not adequately constrain the vertical profiles of winds and temperature. The full wave theoretical model is able to treat gravity wave propagation with arbitrary vertical variations in winds and temperature had we measurements of the full altitude profiles of wind velocity and temperature simultaneous with the airglow observations. Hickey et al. [1997, 1998] have already demonstrated the utility of simultaneous O$^1$S airglow and wind velocity measurements in modeling the gravity wave field.

Wind velocities obtained with the data reported here are variable and refer to only the emission layer. Gravity wave modeling requires the full altitude profile of wind velocity. Wind velocities were determined by several techniques [Hines et al., 1993; Roper et al., 1993; Bird et al., 1993] as part of the AIDA'89 campaign, but uncertainties in these velocities are substantial [Hines et al., 1993]. We have investigated the effects of winds on Krassovsky's ratio with idealized altitude profiles of wind velocity including the Cooperative Institute for Research in the Atmosphere (CIRA) zonal mean wind velocity model for April [CIRA, 1986]. The CIRA winds compare qualitatively with the measured winds over the several days of observation [e.g., Roper et al., 1993]. The results of these calculations (not shown) indicate that Krassovsky's ratio is essentially unaffected by the winds at the large wavelengths and long periods of interest here. Wind profiles not modeled could have more significant effects on $\eta$. One observation worth noting, although its implications are not fully understood, is that the largest values of $[\eta]$ occurred on the only two days in Table 2, May 1, 1989, and May 2, 1989, on which the winds at the level of the emission layer were eastward.

A notable aspect of the data shown in Table 2 is that differences between values of $[\eta]$ among various wave periods on a given day are significantly less than the differences among various days for a given wave period. Also characteristic of the data is the preponderance of large negative values of phase ($< -120^\circ$) from April 8, 1989, to May 1, 1989, with several instances of phases $< -150^\circ$. Large negative values of phase can occur as a result of substantial wave reflection [Hines and Tarasick, 1994; Makhlof et al., 1995]. Wave energy is reflected by thermal and Doppler gradients [Schubert and Walterscheid, 1984; Chimonas and Hines, 1986; Wang and Tuan, 1988], and the reflection is especially strong when gradients give contiguous regions of vertical propagation and evanescence. The waves that are most subject to reflection are those with long vertical scales. For the hour-period waves under consideration, waves with long vertical scales have very long horizontal wave-velocities (>200 m s$^{-1}$). Vertical scales are lengthened by Doppler shifting when a component of the background wind is opposed to the direction of wave propagation. Calculations with standard models of temperature structure indicate that the lower 30 km or so of the thermosphere is a favored region for strong wave reflection. This is a region of high static stability bounded by regions of evanescence for waves with sufficiently long vertical scales. The lower boundary of this region is located near the mesopause. Thus a large reflected component is favored by large horizontal wavelengths, airglow layer centroids above the mesopause, and strong winds opposed to the direction of wave propagation.

The period from April 8, 1989, to May 1, 1989, may have been a period when the O$^1$S airglow layer was largely situated above the mesopause. If so, the dynamical forcing of the airglow fluctuations would have occurred in a region where waves with long horizontal wavelengths might experience strong reflection, especially if they were subject to significant Doppler shifting to higher intrinsic frequencies. We do not know the horizontal wavelengths of the waves in question. However, the data are averaged over all look directions, and this process would effectively eliminate all waves with wavelengths less than about 350 km and would severely attenuate waves at somewhat longer scales. In addition, we do not know how the direction of propagation of the waves and do not know how the winds in the airglow layer would project on the wavenumber vector. However, the winds in the region might have been rather strong. The winds for the period when large phases were noted may have approached 100 m s$^{-1}$ during the night [Roper et al., 1993; Bird et al., 1993]. Our wind data give speeds of 20 to 30 m s$^{-1}$ for these times.

We cannot rule out a tidal cause for the wave periods that are not too different from 12 and 6 hours [Morton et al., 1993], especially 12 hours. Insofar as we are aware, values of $[\eta]$ for tidally driven 557.7 nm airglow fluctuations are not presently available.

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