Incorporation of a gravity wave momentum deposition parameterization into the Venus Thermosphere General Circulation Model (VTGCM)

A. M. Zalucha,1 A. S. Brecht,2 S. Rafkin,3 S. W. Bougher,4 and M. J. Alexander5

Received 24 June 2012; revised 12 November 2012; accepted 12 December 2012; published 31 January 2013.

[1] The gravity wave-drag parameterization of Alexander and Dunkerton (1999) was implemented into a Venus Thermosphere General Circulation Model (VTGCM) to investigate breaking gravity waves as a source of momentum deposition in Venus’ thermosphere. Previously, deceleration of zonal jets on the morning and evening terminators in models was accomplished via Rayleigh friction, a linear drag law that is not directly linked to any physical mechanism. The Alexander and Dunkerton (1999) parameterization deposits all of the momentum of a breaking wave at the breaking altitude and features a spectrum of wave phase speeds whose amplitudes are distributed as a Gaussian about a center phase speed. We did not find a combination of wave parameters (namely, center phase speed, amplitude at center phase speed, and distribution width) to produce sufficient drag in the jet cores that would bring VTGCM density and nightglow emissions into agreement with Venus Express observations. The zonal wind shear from 100 to 120 km altitude is very strong. Gravity waves launched below 100 km either break in the strong shear zones below 115 km or are reflected and do not propagate into the jet core regions where drag is needed. The results we present demonstrate that parameterizations developed for the middle atmosphere do not work in the thermosphere and that appropriate damping mechanisms other than nonlinear breaking/saturation dominate and should be accounted for at these heights.


1. Introduction

[2] Early work with terrestrial global climate models required an ad hoc forcing near the model top to prevent reflection of numerically resolved waves off the model top and to match observations of winter jets in the middle atmosphere [see Kim et al., 2003, for a review]. Specifically, this parameterization was Rayleigh friction [e.g., Leovy, 1964], a momentum sink that was linearly dependent on velocity and increased in magnitude with height. More elaborate schemes varied this forcing in time and space until the desired result was achieved. A “sponge” layer existed in the uppermost model levels where Rayleigh friction acted, but it could not reproduce the observed reversals of zonal wind direction in the mesosphere [Shepherd et al., 1996].

[3] Eventually, it was discovered that the breaking of subgrid scale gravity waves (or buoyancy or internal waves) was the mechanism responsible for momentum drag, and a parameterization of these effects was proposed by Lindzen [1981]. While the theory behind linear, small-amplitude, idealized gravity waves is well understood, these waves present a problem for global climate models because their typical wavelength is smaller than the grid spacing. Additionally, sources of gravity waves are numerous and diverse, including thunderstorms, flow over topography, and baroclinic instabilities. Gravity waves are emitted from these sources in all directions. Vertically propagating gravity waves increase in amplitude as atmospheric density decreases, until they become highly nonlinear and break, depositing their momentum and accelerating or decelerating the background flow.

[4] Several methods have been developed to determine the amount of momentum deposition from gravity waves based on large-scale model state variables. The first was originally proposed by Lindzen [1981], which assumes that waves have a single phase speed and horizontal wave number. When the waves reach the level of linear instability, they saturate and continue propagating upward by a finite distance.
by dissipating just enough momentum for the wave to remain stable. Further improvements were described by Holton [1982], Lindzen [1985], Fritts [1984], and Dunkerton [1989].

Another method, put forth by Lindzen and Holton [1968], prescribes the waves to deposit all of their momentum at the critical level (where the phase speed equals the background wind speed in the direction of propagation). This scheme allows for a spectrum of wave speeds where each wave will in general break at a different level. Alexander and Dunkerton [1999] is a hybrid of the Lindzen and Holton [1968] and Lindzen [1981] schemes, where the waves have a spectrum of wave speeds that deposit all their momentum at one level but have a breaking criterion defined by Lindzen [1981].

The Alexander and Dunkerton [1999] scheme does not contain wave saturation effects, which would distribute momentum over a wider altitude range. This scheme is in contrast to the Venus gravity wave-drug scheme of Zhang et al. [1996], which distributes the energy of the waves throughout the atmosphere based on gravity wave-drug saturation theory and empirical observations of Earth from Fritts and Lu [1993]. Recent direct numerical simulations by Fritts et al. [2009] investigate two wave breaking cases at different amplitudes (for the case of no shear), one just above the stability limit and one below. These simulations involve the primary wave that breaks, well into the turbulence spectrum. They show that the remaining momentum flux after breaking is less than 8% of the original. In the shear case, the amount remaining would generally be even less.

Just like Earth, Venus’s atmosphere also contains gravity waves and planetary-scale waves. Their existence is easily confirmed using photographs of Venus’ clouds [e.g., Belton et al., 1976b; Belton et al., 1976c; Belton et al., 1976a; Rossow et al., 1980; Peralta et al., 2008]. Some of the wave structures are seen as bands at ultraviolet wavelengths; these bands are traveling waves because the bands do not align with any circle of latitude, and the bands travel at a different velocity than the background velocity [Schubert, 1983]. The most notable feature is the dark horizontal “Y” [e.g., Belton et al., 1976b; Belton et al., 1976c; Belton et al., 1976a; Rossow et al., 1980; Schubert, 1983].

Gravity waves are a suggested source for the deceleration of Venus’ zonal flow. There have been very limited observations; however, evidence of gravity waves exists in Pioneer Venus (PV) probe and Orbiter Atmospheric Drag data [Seiff, 1991] and PV Orbiter Neutral Mass Spectrometer (ONMS) data [Kasprzak et al., 1988, 1993]. Kasprzak et al. [1988] examined the PV ONMS data and found wave structures with a wavelength range from 100 to 600 km, and the amplitudes varied depending on the molecular weight of the species (He, N, O, N2, and CO2). Other observations by PV probes and bus were minimal and only able to detect wave perturbations but not characterize them [Seiff et al., 1980; Seiff, 1991; von Zahn et al., 1980]. Moreover, vertical wavelengths above the cloud tops (~20 km) and below (~5–10 km) were deduced from temperature observations [Taylor et al., 1980]. Clouds have been visually observed with a wide horizontal scale from 100 km to hundreds of kilometers with periods of 4 to 6 days [Schubert, 1983]. Additional details on waves from PV observations are discussed in Schubert [1983].

More recently, the Venus Express (VEx) mission has been augmenting knowledge of gravity waves in Venus’ atmosphere by identifying gravity waves in regular cloud patterns [Markiewicz et al., 2007; Peralta et al., 2008]. Waves were also observed in vertical variations of infrared temperature soundings, radio occultation temperatures, upper atmosphere number densities, and other atmospheric quantities [e.g., Schubert, 1983; Covey and Schubert, 1981a, 1981b, 1982; Kasprzak et al., 1988; Bouger et al., 1997]. Furthermore, thermal tides have been observed in thermal structure data and in the atmospheric circulation [Schofield and Taylor, 1983; Limaye, 1990].

The Visible and Infrared Thermal Imaging Spectrometer (VIRTIS) instrument on VEx has detected perturbations in CO2 non-local thermodynamic equilibrium emissions [Garcia et al., 2009]. From these observations, they are able to obtain wave structures with horizontal wavelengths ranging from 90 to 400 km. Additionally, derived horizontal phase velocities (magnitude and direction) are consistent between orbits and are on average of 70 m s\(^{-1}\) westward and 30 m s\(^{-1}\) northward [Garcia et al., 2009], Garcia et al. [2009] claim these observed waves are generated from the polar vortex (note that in our study we do not consider wind ducting from distant locations). Moreover, VIRTIS and Venus Monitoring Camera observed visible trains of oscillating cloud brightness in the UV for an upper cloud layer (~66 km) on the dayside and thermal radiation for the lower cloud layer on the nightside [Peralta et al., 2008].

Wavelengths and phase speeds are also derived from these observations. Peralta et al. [2008] observed wavelengths of 60–150 km, which propagate westward with phase velocities similar to the zonal flow and are confined to horizontal wave packets of 400 to 1800 km in length. They find no correlation between the waves and surface topography, latitude, local time, or wind structure. The perturbations in the observations help provide information on gravity waves in Venus’ atmosphere, which are very important in constraining gravity wave formulations within 3-D models.

Wave drag was first modeled as part of a Venus thermosphere general circulation model (VTGCM) by Bougher et al. [1988], who found good agreement with observations. Mayor et al. [1988] investigated Venus gravity waves using a spectral model and made many conclusions, the most relevant here being the wave amplitudes are a factor of 3 to 5 larger during the night than during the day and waves excited at 130 km are rapidly attenuated (both conclusions in agreement with observations). Alexander [1992] used a Venus gravity wave-drug parameterization to describe the thermospheric superrotation and found that small- to medium-scale gravity waves in the thermosphere can supply the necessary acceleration of the mean flow if the waves have predominately large westward phase speeds. Zhang et al. [1996] adapted the Fritts and Lu [1993] terrestrial gravity wave-drug scheme to Venus, which treats the full wave spectrum (i.e., is not monochromatic) and distributes wave momentum according to wave saturation principles; however, the Fritts and Lu [1993] scheme is no longer used in current terrestrial models [Lawrence, 1997] due to the extreme sensitivity of the parameterization to arbitrary limit parameters. Zhang et al. [1996] also argued that the superrotation in Venus’ thermosphere was caused by vertically propagating gravity waves and was not a remnant of the lower atmosphere circulation.
2. Gravity Wave-drag Parameterization

The gravity wave-drag parameterization is described in section 2; the VTGCM in section 3. VTGCM results are in section 4. The effect of wave parameters on wave breaking level is shown in section 5. Section 6 discusses the results.

3. Venus Thermosphere General Circulation Model

3.1. Model Description

The VTGCM is a 3-D finite-difference hydrodynamic model of the Venus upper atmosphere [e.g., Bougher et al., 1988, 1997, 2008] that is based on the National Center for Atmospheric Research (NCAR) terrestrial Thermospheric Ionosphere General Circulation Model. The VTGCM solves the time-dependent primitive equations for the neutral upper atmosphere: temperature, neutral-ion densities, and three-component neutral winds. The model domain covers a 5° by 5° latitude-longitude grid, with 69 evenly spaced log-pressure levels in the vertical, extending from approximately ~70 to 300 km (~70 to 200 km) at local noon (midnight).

Formulations for CO$_2$ 15 μm cooling, wave drag, and eddy diffusion are parameterized within the VTGCM using standard aeronomical formulations. “Exact” (line-by-line radiative transfer model) CO$_2$ 15 μm cooling rates for a given temperature and composition profile are taken from Roldán et al. [2000]; cooling rates for the simulated VTGCM temperatures and species abundances are calculated (from these exact rates) based upon a slight modification of a parameterization scheme utilized previously [e.g., Bougher et al., 1986]. The corresponding O-CO$_2$ collisional relaxation rate adopted for typical benchmark VTGCM simulations is now $3 \times 10^{-12}$ cm$^3$ s$^{-1}$ at 300 K [Bougher et al., 1999]. This value provides strong CO$_2$ 15 μm cooling that is consistent with the use of EUV-UV heating efficiencies of ~20–22%, which are in agreement with detailed offline heating efficiency calculations of Fox [1988]. The near-IR heating term is incorporated using offline simulated look-up tables, updated recently using Roldán et al. [2000] rates. The most notable
consequence from updating the new IR rates is the doubling of the 4.3 μm heating around 115 km on the dayside. These changes imply a warmer atmosphere closer to solar zenith angle (SZA) of 0, and a corresponding enhancement of CO₂ 15 μm cooling within the VTGCM.

[19] During the spin-up phase, Rayleigh friction was prescribed in order to mimic general wave-drag effects on the mean flow according to

$$ F_{RF} = \lambda_{RF} (u - u_{SR}). \tag{5} $$

where $F_{RF}$ is the drag force, $\lambda_{RF}$ is a damping rate, $u$ is the VTGCM calculated zonal wind, and $u_{SR}$ is the specified zonal wind speed to approximate the retrograde superrotating during the VEx sampling period [see maximum value of 1 (Brecht et al., 1988)]. The maximum $u_{SR}$ value was set to 75 m s$^{-1}$. The damping rate is expressed as

$$ \lambda_{RF} = \lambda_{o} \sqrt{\frac{p_{\text{break}}}{p}}, \tag{6} $$

where $\lambda_{o}$ is the maximum $\lambda_{RF}$ ($\lambda_{o} = 0.5 \times 10^{-4}$ s$^{-1}$), $p_{\text{break}} = 2.24 \times 10^{-7}$ μbar is the breaking level, and $p$ is pressure. $\lambda_{RF}$ is referred to as the symmetric Rayleigh drag term; $u_{SR}$ is the asymmetric Rayleigh drag term. The $\lambda_{RF}$ and $u_{SR}$ terms are based upon empirical matches to Pioneer Venus Orbiter (PVO) and VEx observations that contain a specified exponential profile dependent on cos(latitude). The resulting $F_{RF}$ term is then added to the momentum equation. For more detail on the Rayleigh friction formulations, see Brecht et al. [2011].

[20] The eddy diffusion coefficient on the nightside is prescribed in the form $K = \frac{D}{n}$ with units of cm$^2$ s$^{-1}$ where $n$ is the total number density and $D$ is a constant [von Zahn et al., 1979]. The nightside eddy diffusion has a prescribed maximum value of $1 \times 10^{-7}$ cm$^2$ s$^{-1}$, and the dayside has a constant value for the entire upper atmosphere of $1 \times 10^{-9}$ cm$^2$ s$^{-1}$. For more details about the VTGCM, see Brecht et al. [2011].

3.2. Summary of “Mean” VTGCM Case Results

[21] The results for this study are based upon a VTGCM “mean” case, which is representative of mean conditions during the VEx sampling period [see Brecht et al., 2011; Brecht et al., 2012], and are produced using a Rayleigh friction scheme to provide the best match with observations. Ideally, simulations with properly represented gravity waves will closely resemble this simulation and the observations. The VEx mean conditions are depicted by statistically averaged NO and O₂ nightglow emission maps (observed emissions averaged spatially and temporally) [Gérard et al., 2008; Piccioni et al., 2009; Soret et al., 2012]. Specific parameters in the VTGCM are tuned to achieve this mean condition, which are given as follows: solar minimum fluxes (F10.7-cm=70 units), the maximum nightside eddy diffusion coefficient ($1.0 \times 10^{-7}$ cm$^2$ s$^{-1}$), and the wave-drug parameter ($0.9 \times 10^{-4}$ s$^{-1}$). This mean case produced a maximum NO UV nightglow emission near the equator at 108 km. The calculated peak vertical intensity was 2.28 kR (Rayleigh = 10$^6$ photons cm$^{-2}$ s$^{-1}$ sr$^{-1}$) with a hemispheric average intensity of 0.78 kR. The O₂ IR nightglow emission was also calculated near the equator but at 102 km. The O₂ IR nightglow emission peak vertical intensity for a three-body reaction (O + O + CO₂) yield of 75% was 2.19 MR with a corresponding hemispheric average intensity of 0.53 MR. These nightglow peak intensities are within the VEx observational ranges, and the hemispheric averages agree favorably with the VEx observations [Brecht et al., 2011].

[22] Monitoring and modeling the nightglow emission gives rise to an understanding of the global wind system. The VTGCM neutral zonal winds near the equator at the morning terminator were $-108$ m s$^{-1}$ at 110 km, $-166$ m s$^{-1}$ at 120 km, and 150 m s$^{-1}$ near 180 km. The evening terminator winds were $112$ m s$^{-1}$ at 110 km, $191$ m s$^{-1}$ at 120 km, and $289$ m s$^{-1}$ near 180 km. The evening terminator winds are faster than the morning terminator winds because the wave-drag term, Rayleigh friction, is prescribed asymmetrically in local time in order to mimic the observed upper atmosphere RSZ winds. The prescribed RSZ winds are very weak (<10 m s$^{-1}$) from ~80 km to 112 km, and above 110 km the emergence of modest RSZ winds approach ~100 m s$^{-1}$ above ~130 km. This RSZ profile is slightly different than the profile described in Brecht et al. [2011] due to the updated VTGCM temperatures. This RSZ profile is needed to produce the O₂ IR nightglow peak intensity near midnight and the NO UV nightglow peak intensity to be positioned near 01:00 LT.

[23] The convergence of the global wind system on the nightside not only produces night airglow emission, but it also contributes to the nightside heat balance [Brecht et al., 2011]. As discussed previously, the 4.3 μm heating on the dayside near 115 km and the corresponding day-to-night global circulation produces a downswwelling component on the nightside, which results in dynamical heating near midnight. This mean simulation produces a maximum temperature on the nightside of 198 K at 104 km with a corresponding total dynamical heating rate of 71 K day$^{-1}$. This nightside peak temperature is in accord with available spacecraft and ground based observations [Bertaux et al., 2007; Bailey et al., 2008; Brecht et al., 2011]. For more details on the mean case, parameters, or observations, see Brecht et al. [2011].

4. Implementation and Results

[24] We initialized the combination VTGCM/gravity wave-drag simulations from the VTGCM mean state (section 3.2) that had symmetric Rayleigh friction (i.e., $u_{SR}=0$ in equation (5)) during its spin-up phase. Figure 1 shows a longitude-height cross section of the zonal winds at latitude 2.5° for this state. The zonal winds are symmetric in local time because of the symmetric Rayleigh friction. At the start of the combination VTGCM/gravity wave-drag simulations, the gravity wave-drag term was turned on. In the numerical gravity wave-drag scheme, e is discretized into $n_c = 1200$ bins of width $\Delta e = 0.5$ m s$^{-1}$ centered on $e_o$. $n_c$ is sufficiently large enough for the effect of the waves to vary smoothly with phase speed, while small enough to maintain computational efficiency.

[25] Note that faster harmonics propagate with a greater horizontal tilt and can leave the model’s vertical grid column (at a given horizontal resolution) while propagating to the top of the model. For them, the approximation of vertically propagating gravity waves is no longer suitable, and 1-D gravity wave (GW) parameterizations are not applicable. The discussion of this issue in Yiğit et al. [2008, 2009] shows that waves faster than 100 m s$^{-1}$ should not be considered in
also see section 5). Values for the wavespectrum that we use (discussed below, see also section 5).

Waves are assumed to be launched by clouds, and $c_0$ is set to the wind speed at the cloud tops. $B_r$ controls the initial amplitude of waves and must be large enough for waves to influence the atmosphere but not so large that waves break too low in the atmosphere.

While there is some guidance for the choice of $B_r$, $C_0$, $C_r$, $C_w$, and the launch level from terrestrial cases, the appropriate values for Venus are unknown. A wide range of parameters were tried. For $B_r$, values spanning many orders of magnitude, specifically $10^{-6}$, $10^{-5}$, $10^{-4}$, $10^{-3}$, $10^{-2}$, $10^{-1}$ kg m$^{-2}$ s$^{-1}$, and also finer increments of 0.0008, 0.0009, 0.0011, 0.0012, 0.002, 0.003 kg m$^{-2}$ s$^{-1}$ were tried. For $C_0$, values of 0, 10, 20, 30, 45, 90, 180, and $-90$ m s$^{-1}$ were tried in the zonal direction and 0, 90, and 180 m s$^{-1}$ in the meridional direction. $C_w$ was also set to the local wind speed at the launch level, $u$ and $v$. For $C_r$, values of 40, 80, 90, 120, and 200 m s$^{-1}$ were tried. Values for $C_w$ were 0.0004167 (1/2400), 0.0008333 (1/1200), 0.0016667 (1/600) 0.008, 0.009, 0.01, and 0.1. Launch levels ranged from the bottom level (70 km) of the VTGCM, 90 km, 100 km, and from multiple levels from 70 to 90 km and 70 to 100 km.

Despite considering a very wide range of cases, no simulations provided a suitable match to VEx observations. Because the parameters are not linearly independent, we cannot conclude that it is impossible to find a working combination; however, we will show a range of cases that suggest it is unlikely. The description of the best case follows. As a first demonstrating case (case 1), the gravity wave-drag parameters were set to $B_r = 10^{-5}$ kg m$^{-2}$ s$^{-1}$, $C_u = 80$ m s$^{-1}$, $C_w = 0$, and $C_r = 1/1200$ for all azimuthal directions, and further integrated the model for 9 Earth days of model time. We have run the model simulations out to 63 Earth days but find no significant change from the 9 day results. Ideally, the gravity wave-drag scheme will perturb the mean state to a new steady state in which gravity waves provide additional momentum drag in the upper atmosphere. Figure 2 shows the resulting longitude-height cross section of the gravity wave drag, and Figure 3 shows the corresponding zonal winds. Westerly drag exists where there are easterly winds, and easterly drag exists where there are westerly winds. Drag is absent from the jet cores near the morning and evening terminators (90° and $-90°$ longitude, respectively) and above 110 km altitude. This is an undesirable result because winds must decelerate at all longitudes. Figure 4 is the difference in winds between the combination VTGCM/gravity wave-drag results and the symmetric Rayleigh friction winds (i.e., Figure 3 minus Figure 1). The locations of nonzero differences between these two winds occurs at the locations of nonzero gravity wave drag, as expected, but magnitudes of the zonal winds in the centers of the jet cores are hardly affected by the addition of gravity wave drag using this gravity wave parameter set.

PVO and VEx observations of density and nightglow emissions, interpreted using general circulation models with Rayleigh friction [e.g., Bouger et al., 2008; Brecht et al., 2011], show that the zonal winds at the morning terminator (near 90° longitude) should be weaker than on the evening terminator (near near $-90°$ longitude). Figure 5 shows a VTGCM simulation with Rayleigh friction applied in a manner that is asymmetric in local time (i.e., $u_{SR} \neq 0$ in equation (5)). Gravity wave drag is turned off in this simulation.
To obtain this wind pattern using gravity wave drag, the magnitude of the drag on the morning terminator (where the winds are easterly) must be stronger. The zonal $c_w$ was shifted to $90 \text{ m s}^{-1}$ (as suggested by VEx observations at the cloud tops by Moissl et al. [2009]), such that the amplitude of the phase speed spectrum is weighted towards $\omega$. 

**Figure 2.** Instantaneous case 1 gravity wave drag ($10^{-3} \text{ m s}^{-2}$) at latitude 2.5° after 9 Earth days of VTGCM/gravity wave-drag integration. The subsolar point (local time 12 h) is at 180° longitude. The morning and evening terminators are located at 90° and −90° longitude, respectively. Contour intervals are $10^{-3} \text{ m s}^{-2}$ with a maximum and minimum of $\pm 6 \times 10^{-3} \text{ m s}^{-2}$.

**Figure 3.** Instantaneous case 1 zonal winds ($\text{m s}^{-1}$) at latitude 2.5° after 9 Earth days of VTGCM/gravity wave-drag integration. The subsolar point (local time 12 h) is at 180° longitude. The morning and evening terminators are located at 90° and −90° longitude, respectively.
**Figure 4.** Difference between instantaneous case 1 VTGCM/gravity wave-drag zonal winds and zonal winds with symmetric Rayleigh friction during spin-up (m s\(^{-1}\)). Latitude is 2.5°. The VTGCM/gravity wave-drag model has been integrated for 9 Earth days, starting from the symmetric Rayleigh friction configuration. The subsolar point (local time 12 h) is at 180° longitude. The morning and evening terminators are located at 90° and −90° longitude, respectively. There is a ∼20 m s\(^{-1}\) reduction in the winds near 135 km altitude and ±130° longitude and a ∼10 m s\(^{-1}\) reduction near 130 km altitude and ±150° longitude, but otherwise the wind magnitudes are unchanged in the jet cores (±90° and above 100 km).

**Figure 5.** Initial VTGCM zonal winds (m s\(^{-1}\)) at latitude 2.5° and asymmetric Rayleigh friction. Gravity wave drag has not been turned on. The subsolar point (local time 12 h) is at 180° longitude. The morning and evening terminators are located at 90° and −90° longitude, respectively.
westerly waves and repeated the combination VTGCM/gravity wave-drag simulations (case 2) starting from the symmetric spun up state (Figure 1). Figure 6 shows a longitude-height cross-section of the gravity wave drag, and Figure 7 shows the zonal winds after 9 Earth days of model integration with gravity wave drag turned on. The resulting drag force is now asymmetric in local time, as expected and desired. The magnitude of the zonal wind on the edges of the jet has been modiﬁed near the areas where there is nonzero gravity wave drag, but the drag in the jet core is insufﬁcient. Figure 8 shows the difference between the VTGCM/gravity wave-drag zonal winds for case 2 and the zonal winds with asymmetric Rayleigh friction during spin-up. Significant (≥100 m s−1) deviations exist between the VTGCM/gravity wave-drag zonal wind results and the asymmetric wind pattern that has been shown to match data with asymmetric Rayleigh drag; the character of the asymmetry in the zonal wind structure needed to match VEx observations is not reproduced.

5. Investigation of Breaking Level

Neither conﬁguration of the combination VTGCM/gravity wave drag from section 3 was able to produce drag that penetrated the jet core. Since this gravity wave-drag scheme has previously only been applied to Earth, we know the gravity wave parameterization must be tuned to achieve the best match of VTGCM output ﬁelds to VEx observations. The parameter space is large. Of particular phase speed and longitude pair before total internal reﬂection occurred. Thus, the waves have not deposited any momentum in this column. This behavior occurs for positive phase speeds near 90° longitudes, and negative phase speeds near −90° longitude, i.e., in the jet cores when phase speed and zonal wind are of different signs. It also occurs for very high-magnitude phase speed waves at any longitude. At ±90° longitude, when the phase speed and zonal wind are of different sign (ﬁrst and third quadrants), the breaking level is deﬁned but is low (≤105 km). At these low altitudes and higher atmospheric densities, the breaking waves have only a weak forcing effect.

On the jet flanks (0°–45°, 135°–180°, −180°–(−135°), and −45°–0° longitude), waves of both signs break at higher altitudes. The gravity wave-drag force deﬁned by equation (4) depends on both the amplitude of the waves as a function of phase speed and inversely on the density as a function of height. The combination of these parameters will determine the sign of the waves that dominate, and they are not easily read off of the wave breaking plots. No waves break above 135 km because we have deﬁned the turbopause to be at this level, and all waves that penetrate this high are forced to decay exponentially with height above this location.

Table 1 shows additional combinations of gravity wave parameters. Figure 10 shows the wave breaking height
**Figure 7.** Instantaneous case 2 zonal winds (m s$^{-1}$) at latitude 2.5° after 9 Earth days of VTGCM/gravity wave-drag integration. The subsolar point (local time 12 h) is at 180° longitude. The morning and evening terminators are located at 90° and −90° longitude, respectively.

**Figure 8.** Difference between instantaneous case 2 zonal winds and zonal winds with asymmetric Rayleigh friction during spin-up (m s$^{-1}$). Latitude is 2.5°. The VTGCM/gravity wave-drag model has been integrated for 9 Earth days. The subsolar point (local time 12 h) is at 180° longitude. The morning and evening terminators are located at 90° and −90° longitude, respectively. Significant ($\geq 100$ m s$^{-1}$) differences occur at positive longitudes, indicating that the gravity wave drag has not produced enough change in the winds to slow them down there.
for case 2 (wave amplitude weighted towards westerly phase speeds). As in case 1 (symmetric wave amplitude), the jet cores either contain waves that break too low or undergo total internal reflection. The breaking level has increased in altitude for the negative phase speeds and decreased for the positive phase speeds in the flanks of the jets.

Table 1. Parameter Sets for Wave Breaking Analysis

<table>
<thead>
<tr>
<th>Case</th>
<th>$B_o$ (kg m$^{-2}$ s$^{-1}$)</th>
<th>$c_w$ (m s$^{-1}$)</th>
<th>$c_m$ (m s$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>$10^{-5}$</td>
<td>0</td>
<td>80</td>
</tr>
<tr>
<td>2</td>
<td>$10^{-5}$</td>
<td>90</td>
<td>80</td>
</tr>
<tr>
<td>3</td>
<td>$10^{-8}$</td>
<td>0</td>
<td>80</td>
</tr>
<tr>
<td>4</td>
<td>$10^{-2}$</td>
<td>0</td>
<td>80</td>
</tr>
<tr>
<td>5</td>
<td>$10^{-5}$</td>
<td>0</td>
<td>40</td>
</tr>
</tbody>
</table>

Figure 9. Gravity wave breaking level (km) for case 1. Contour intervals are 5 km. The morning and evening terminators are located at 90$^\circ$ and −90$^\circ$ longitude, respectively.

Figure 10. Gravity wave breaking level (km) for case 2. Contour intervals are 5 km. The morning and evening terminators are located at 90$^\circ$ and −90$^\circ$ longitude, respectively.

Figure 11. Gravity wave breaking level (km) for case 3. Contour intervals are 5 km. The morning and evening terminators are located at 90$^\circ$ and −90$^\circ$ longitude, respectively.

Figure 12. Gravity wave breaking level (km) for case 4. Contour intervals are 5 km. The morning and evening terminators are located at 90$^\circ$ and −90$^\circ$ longitude, respectively.

shows the breaking level for the case with $B_o$ increased by 3 orders of magnitude. Comparing these along with case 1 shows that the breaking level decreases in altitude with increasing $B_o$ because the wave reaches critical amplitude lower in the atmosphere. Still, the waves at the longitudes of the jet cores either break too low or undergo total internal reflection. Figure 13 shows the breaking level for the case of $c_w$ decreased by a factor of 2, which narrows the distribution of wave amplitudes. In the jet flanks, the wave breaking level decreases with decreasing $c_w$, and the jet cores remain untouched.

6. Discussion

Rayleigh friction provides a source of momentum dissipation in general circulation models (GCMs); however, it does not have a physically grounded basis. Rayleigh friction is useful to provide a context for the magnitudes
and altitudes of momentum deposition needed in the upper atmosphere for reproducing VEx observations. Gravity waves launched from the cloud tops are a mechanism for depositing momentum when they propagate upwards, then become unstable and break. Gravity waves are small in spatial and temporal scales compared with the resolution of a GCM. Therefore, some parameterization is still needed.

Although there are several different gravity wave parameterizations in common use in Earth middle atmosphere models, McLandress and Scinocca [2005] showed that these can all be tuned to give rather similar results. One difference among these parameterizations is whether or not they include the process of total internal reflection. The Alexander and Dunkerton [1999] parameterization includes this process. Others do not, although Scinocca [2002] described a method for adding to the process to the Warner and McIntyre [2001] parameterization, and this modified parameterization is now in use at the European Center for Medium Range Weather Forecasting [Orr et al., 2010]. Our results suggest that if we chose a different parameterization for Venus that does not include reflection, that while we might obtain drag in the jet cores, this would be spuriously due to the absence of this apparently important process.

The gravity-wave-drag scheme presented in Alexander and Dunkerton [1999] (and later Ortland and Alexander [2006]) has been successfully used in Earth upper atmosphere models. Venus has very high-speed thermospheric jets on the morning and evening terminators that in the absence of any momentum drag produce strong winds that are inconsistent with VEx observations. Three key parameters need to be adjusted for application to Venus: the initial amplitude of the gravity wave spectrum \( B_0 \), the spectrum width \( c_w \), and the offset of phase speeds \( c_o \). The first two parameters have less connection to a large-scale, model-resolved physical parameter, while we interpret \( c_o \) to be related to the model background wind speed at the launch level (in this case, the cloud tops). \( B_0 \), \( c_o \), and \( c_w \) were varied over a large area of parameter space, namely, \( B_0 = 10^{-5} \) to \( 10^{-2} \) kg m\(^{-2}\) s\(^{-1}\), \( c_o = 0 \) to 90 m s\(^{-1}\), and \( c_w = 40 \) to 80 m s\(^{-1}\), the results of which were presented in section 5, as well as various combinations of these parameters and more extreme values. In every case, within the morning and evening terminator jets, the waves either are totally internally reflected, thus prohibiting them from breaking and depositing any momentum or breaking at too low an altitude (\( \leq 105 \) km) where the winds are weak. The low-level shear is too great for waves of any physically plausible configuration to propagate upwards into the high-speed regions of the jets to potentially decelerate the winds.

The version of the Alexander and Dunkerton [1999] gravity wave-drag scheme implemented in the VTGCM was able to modify the winds in the jet flanks, but is peripheral to the main goal of decelerating the winds in the jet core. It was found that lower \( B_0 \) corresponds to higher breaking level, for reasons as follows. The background atmospheric density decreases with height, while wave amplitude grows with decreasing density. Waves propagate upwards until the density becomes low enough that they are no longer in the linear (stable) regime and therefore break. A group of waves that have an initially lower amplitude (lower \( B_0 \)) will be able to propagate to lower densities (higher altitudes) before breaking. This behavior is represented mathematically in the breaking equations, equation (2). \( \rho, u, \) and \( N \) all depend on \( z \), but the \( \rho \) factor is the strongest because it has an exponential dependence. Thus, a lower \( B_0 \) results in a smaller \( \rho \) (higher \( z \)) when the left side of the equation reaches 0.

It was also found that lower \( c_w \) corresponds to lower breaking level. \( c_w \) controls the width of the initial amplitude distribution; lower \( c_w \) makes the distribution narrower. For a given phase speed \( c \), lower \( c_w \) causes the initial amplitude of that particular wave to be decreased (except at the center of the distribution \( c = c_w \) where the initial amplitude remains the same). Lower initial wave amplitude corresponds to higher breaking level, for the same arguments as above.

For a wind distribution that is symmetric in local time, shifting \( c_o \) off of 0 causes the breaking level to be asymmetric in local time. \( c_o \) controls the phase speed where the maximum initial wave amplitude occurs. The breaking of each wave is controlled by its phase speed relative to the background wind. For \( c_o = 0 \), waves with opposite signed phase speeds will have the same initial amplitude and same relative speed with respect to the background wind and will thus break at the same level (but with opposite signed drag). For \( c_o \neq 0 \), the distribution of amplitudes is not symmetric about \( c = 0 \). Thus, waves with opposite signed phase speeds will now have different initial amplitudes and different breaking levels.

The intermittency factor \( \varepsilon \) enters into the force equation but not the breaking level. \( \varepsilon \) is a factor between 0 and 1 that allows for gravity waves that are not necessarily being produced continuously at all times. \( \varepsilon \) can be increased or decreased to produce the desired force of the gravity waves as \( B_0 \) is changed. However, since no waves are penetrating the jet cores, the force is zero, and \( \varepsilon \) is not relevant.

Several parameters have remained fixed in the analysis. The launch level is always taken to be the cloud tops at the base of the VTGCM. Presumably, the source of the gravity waves is mesoscale convection in the clouds that penetrates into stable layers. We have tried to raise the launch level to 100 km, but then the waves that would have broken in the 70–100 km region break at 100 km instead, which is still too low to affect the jet cores. The wave breaking criteria above 100 km is still the same, and the behavior above unchanged.
from the 70 km launch level case within the jet core magnitudes. While we were unable to obtain realistic drag with gravity wave launched at cloud levels, it is possible that secondary waves may be a source for waves at higher altitudes. For example, cloud top waves may break in the strong shear zones between 100 and 115 km and generate secondary waves, which can subsequently penetrate into the jet cores and give the drag forces needed to slow the winds. Hence, the addition of this process is not likely to assist in weakening the overly strong jets in the Venus thermosphere. 

[47] Medvedev et al. [2011] and Medvedev and Yi it [2012] recently adapted a thermospheric gravity wave parameterization to Mars from a terrestrial scheme by Yiğit et al. [2008], which includes the effects of gravity wave saturation and breaking, gravity wave dissipation due to molecular viscosity, thermal conduction, ion drag, and radiative damping. The Medvedev et al. [2011] gravity wave saturation and breaking scheme is a spectral parameterization that distributes the wave amplitudes as a Gaussian over phase speeds with the same free parameters $c_w$, $B_w$, and $c_r$ as in Alexander and Dunkerton [1999] (our equation (2)). The Medvedev et al. [2011] and Medvedev and Yi it [2012] scheme is different than the Alexander and Dunkerton [1999] scheme in that the waves can be saturated at multiple heights and are not completely removed at a single breaking level. However, since we have shown that gravity waves are unable to propagate to high levels due to total internal reflection or breaking at low levels, the gravity wave saturation and breaking specifications of Medvedev et al. [2011] and Medvedev and Yi it [2012] would not be effective momentum flux deposition mechanisms in the VTGCM. Moreover, the role of the radiative damping used in their model is in fine tuning and shaping the momentum deposition patterns.

[48] Currently, the bottom boundary condition of the VTGCM is static, but a time-varying boundary condition (at fixed pressure) in temperature, 3-D winds, and height surfaces would enable upward propagating planetary waves and tides to be present to reduce the strong shear in the 100–115 km layer. This might allow the parameterized gravity waves to penetrate the jet core without reflection and slow the winds in the thermosphere. The gravity wave-drag model would then be applied as described here. Hence, we would be considering a combination of dynamic processes rather than addressing them in isolation from one another.

[49] We noted secondary wave generation might be another important process in the Venus atmosphere that could provide the needed drag on thermospheric winds. One additional process that is neglected in gravity wave parameterizations but could be important on Venus is horizontal wave propagation. Sato et al. [2012] describe how this process can focus wave momentum into the jet core of the Earth’s middle atmosphere, and if something similar occurs on Venus, it could also assist in slowing the thermospheric winds.

[50] Acknowledgments. This work was supported by the NASA Venus Express Participating Scientist Program, NNX10AI35G.

References


ZALUCHA ET AL.: GRAVITY WAVES IN A VENUS MODEL


