Nonlinear interactions between gravity waves with different wavelengths and diurnal tide

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[1] By using a compressible nonlinear two-dimensional gravity wave model, we simulate the nonlinear interactions between gravity waves (GWs) with three different vertical wavelengths and a meridional component of the diurnal tidal wind and temperature (30°N, September) calculated from GSWM-00. We also compare the results with the simulation of GWs propagation in a background with zero wind. Consistent with the dispersion relation, the numerical experiments show that tidal wind reduces the vertical wavelengths of the GWs when it is in the same direction as the wave propagation, and thus increases the perturbative shear and the likelihood of instability and wave breaking, especially for waves with shorter vertical wavelengths. The breaking of GWs below the critical level can increase the amplitude of diurnal tidal wind due to the momentum deposition. Because the GW penetration height increases with vertical wavelength, the amplitude of diurnal tidal wave at higher altitudes is more likely to be affected by GWs with large vertical wavelengths. Therefore gravity wave breaking not only accelerates the mean winds, but also increases the amplitudes of the diurnal tide at various altitudes.


1. Introduction

[2] Gravity waves (GWs) and tides are important waves in the mesosphere and lower thermosphere (MLT) region. The amplitudes of GWs and tides grow inversely proportional to the square root of the atmospheric density. These waves become unstable and break in the MLT region, resulting in the deposition of momentum and energy and variability in the mean and tidal fields therein. The study of the gravity wave impact in the MLT region has been an active research topic in atmospheric dynamics in the last two decades [e.g., Fritts and Vincent, 1987; Thayaparan et al., 1995; Preusse et al., 2001; She et al., 2004; Fricke-Begemann and Hoffner, 2005; England et al., 2006; Ortland and Alexander, 2006].

[3] Observations of atmospheric tides are often characterized by large day-to-day variations [e.g., Thayaparan et al., 1995; Williams et al., 1999; She, 2004]. Recent lidar observations show clear evidence that the diurnal amplitudes vary by a factor of 2–3 [e.g., She et al., 2004]. The variability of the apparent tidal amplitudes can be caused by the interactions between diurnal tides, planetary waves and GWs [e.g., Fritts and Vincent, 1987; Williams et al., 1999; Preusse et al., 2001; She et al., 2004; Liu et al., 2007], either globally or locally.

[4] Numerical simulation is an effective method in studying tidal variability, especially in investigating the nonlinear interactions between GWs and tides. In order to study the effects of GWs on tides, GW parameterizations have to be used in global scale models due to the differences of the temporal and spatial scales between GWs and tides [Miyahara and Forbes, 1991; Melandrand and Ward, 1994]. Because of the different physical mechanisms and assumptions involved in the parameterization scheme, attempts to study GWs-tidal interactions using global scale models have yielded some contradictory results regarding the impact of parameterized GWs drag on the propagating diurnal tide [e.g., Melandrand, 1997; Meyer, 1999; England et al., 2006; Ortland and Alexander, 2006]. More recently, Ortland and Alexander [2006] showed that the phase of the GWs forcing relative to the tide affects the tidal amplitude and vertical wavelength. However, for a given source spectrum, the relative phase between tides and GWs forcing calculated from different GWs parameterization schemes can be quite different. It is thus necessary to study GWs propagation and breaking in the background with tidal wind and their impacts on tides.

[5] Mesoscale models are useful to study the interactions between GWs and tides [e.g., Liu and Hagan, 1998; Goya and Miyahara, 1999; Liu et al., 2000]. Liu and Hagan [1998] simulated the interactions between GWs and a meridional component of the diurnal tidal wind calculated from GSWM. The results show that the breaking of GWs...
enhances the local dynamical cooling and heating. The results of Liu et al. [2000] indicate that the tidal modulation of GWs is dependent on season and latitude, commensurate with the seasonal and latitudinal variation of tides. Because of the large variation of scales of GWs in the atmosphere, better understanding of the nonlinear interactions between GWs with different wavelengths and tides is needed.

[6] In this paper, we study the nonlinear interactions between GWs and the meridional component of the diurnal tidal wind and temperature (30°N, September) calculated from GSWM-00 [Hagan et al., 2001] in the MLT region, using the model developed by Liu et al. [2006]. The meridional tidal component is chosen in the numerical experiments, because the mean meridional wind is relatively small and we can focus on gravity-wave/tide interaction. A large mean wind, like that in the zonal direction, will affect the specific gravity wave components that can propagate into the mesosphere due to filtering, but it should not affect the essential features of tidal modulation of gravity waves or the impact on the tides. We focus on the effect of the tide on the propagation of GWs with different wavelengths and the temporal variability of the tide induced by GWs. Section 2 presents the numerical model, background tidal wind and GWs initial perturbation. Section 3 examines the effect of the tide on the variations of the dynamically and convectively unstable regions of GWs. The temporal variability of tide and background wind induced by GWs is given in section 4. The conclusions are presented in section 5.

2. Numerical Model and Initial Conditions

[7] The model solves for two-dimensional, fully nonlinear, compressible, non-hydrostatic motions in a plane atmosphere. The equations of the model are as follows [Xu et al., 2003; Liu et al., 2006]:

$$ \frac{\partial \rho}{\partial t} + \frac{\partial (\rho u)}{\partial x} + \frac{\partial (\rho w)}{\partial z} = 0, $$

$$ \frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} + w \frac{\partial u}{\partial z} + \frac{1}{\rho} \frac{\partial p}{\partial x} = K_{\text{eddy}} \left( \frac{\partial^2 u}{\partial x^2} + \frac{\partial^2 u}{\partial z^2} \right) - \alpha(z) u', $$

$$ \frac{\partial w}{\partial t} + u \frac{\partial w}{\partial x} + w \frac{\partial w}{\partial z} + \frac{1}{\rho} \frac{\partial p}{\partial z} + g = K_{\text{eddy}} \left( \frac{\partial^2 w}{\partial x^2} + \frac{\partial^2 w}{\partial z^2} \right) - \alpha(z) w', $$

$$ \frac{\partial T}{\partial t} + u \frac{\partial T}{\partial x} + w \frac{\partial T}{\partial z} + \frac{RT}{C_v} \left( \frac{\partial u}{\partial x} + \frac{\partial w}{\partial z} \right) = K_{\text{eddy}} \left( \frac{\partial^2 T}{\partial x^2} + \frac{\partial^2 T}{\partial z^2} \right) - \alpha(z) T', $$

$$ p = \rho RT, $$

where, $x$ and $z$ are the meridional and vertical coordinates, $g$ is the acceleration of gravity, $\rho$ is the density of atmosphere, $p$ is the pressure, $u$ and $w$ are the horizontal and vertical velocities respectively, $T$ is the temperature, $C_v$ is the specific heat at constant volume, $R$ is the gas constant for dry air. $K_{\text{eddy}}$ is the eddy diffusion and defined as,

$$ K_{\text{eddy}} = \begin{cases} 0, & N^2 > 0 \\ \frac{50}{1 + \tanh \left( \frac{z - 85}{15} \right)}, & N^2 \leq 0. \end{cases} $$

[8] So, the diffusion coefficients gradually increase to 100 m² s⁻¹ in the regions where the waves will break (60–120 km) and at the time the wave will becomes convectively unstable ($N^2 \leq 0$). The eddy diffusion formulation used in the model is a simplified and first order formula to account for the subgrid turbulence. Although a first order turbulence model cannot simulate the energy density and transport of the turbulence, like the second order (or higher) turbulence model [e.g., Liu et al., 1999], it nevertheless can qualitatively represent the effects of turbulence mixing on the GWs around the onset of instability. By using (6), the wave will not be damped below 60 km or before the GWs become unstable. In the equations, $u'$, $w'$ and $T'$ are the fluctuations associated with GWs, which are calculated as follows:

$$ f'(x, z, t) = f(x, z, t) - \frac{1}{T_e} \int_0^{T_e} f(x, z, t) \, dt, $$

where $f$ can be $u$, $w$ or $T$. $f'$ is the horizontal average over one wavelength, which is time-dependent, and $T_e$ is the horizontal wavelength of GWs. $\alpha(z)$ is the Rayleigh friction coefficient and the same as that used by Xu et al. [2003].

[9] A staggered grid system is chosen in the computation for spatial discretization. The staggered grid system implicitly increases computational accuracy and more naturally observes conservation laws [Walterscheid and Schubert, 1990; Liu et al., 1999]. A second order center difference schemes are used in the vertical direction, and the fourth-order compact difference schemes are used in the horizontal direction [Santhanam et al., 2003].

[10] Temporal integration is advanced semi-implicitly by using the time splitting method [Ferziger and Perić, 1996]. The explicit scheme used in the horizontal direction is the third-order total variation diminishing (TVD) type Runge-Kutta method [Shu, 1988]. The FICE (Full-Implicit-Continuous-Eulerian) scheme is applied in the vertical direction because of its good stability [Hu and Wu, 1984; Zhang and Yi, 1998]. For the spatial discretization, there are 50 grids in the meridional direction and 340 grids in the vertical direction. According to the Courant-Friedrichs-Lewy (CFL) conditions for the explicit third-order TVD type Runge-Kutta method, the time step is set to 1.0 s for integration in the horizontal direction. While the symmetrical time splitting method requires that one horizontal time step matches with two half vertical time step to increase the resolution in the vertical direction, so each of the half vertical time step is 0.5 s [Ferziger and Perić, 1996].

[11] Periodic lateral boundary condition is used. The zero flux condition is used for the velocities and temperature on the upper-boundary, while the density and pressure are
approximated in the hydrostatic equilibrium and state equation. The lower-boundary condition is fixed on the ground and a small Rayleigh friction is applied near the ground to remove the reflection from the ground.

The validity of the model has been tested and presented by Liu et al. [2006]. The energy density and dispersion relation of the model coincide well with the linear theory when the model satisfies the conditions of small amplitude and isothermal atmosphere.

The background atmosphere is isothermal with \( T = 239 \text{ K} \) superposed with the meridional component of diurnal tidal wind and temperature (30°N, September, Shown in Figure 1) calculated from GSWM-00 [Hagan et al., 2001]. The background wind and temperature constitute the time independent parts of \( u \) and \( T \) in equations (1)–(5). The vertical range of the computational domain extends from the ground to 170 km. The horizontal width is equal to a horizontal wavelength.

Initial perturbations to wind, temperature and density fields are introduced at the height of 40 km. A GW packet is induced by adjustment process of this initial perturbation. The horizontal velocity perturbation has the following spatial dependence [Zhang and Yi, 1998; Xu et al., 2003; Liu et al., 2006].

\[
u'(x, z, t = 0) = A \cos(k_x x + k_z (z - z_0)) \exp \left[-\ln2 \left( \frac{z - z_0}{\lambda_z} \right)^2 \right] \exp \left( \frac{z - z_0}{2H} \right)
\]

where \( k_x = 2\pi/l_x \) and \( k_z = -2\pi/l_z \) are wave numbers in the horizontal and vertical directions respectively, \( z_0 = 40 \text{ km} \), \( H \) is the scale height, \( l_x \) and \( l_z \) are the horizontal and vertical wavelength respectively. \( \lambda_z \) is the half width of the wave packet. \( u'(x, z, t = 0), p'(x, z, t = 0), \rho'(x, z, t = 0) \) and \( T'(x, z, t = 0) \) are derived from \( u'(x, z, t = 0) \) using the polarization equations for linear GWs. The initial perturbations are superposed on the background field in our simulations. According to linear theory, the maximum horizontal velocity amplitude \( A \) that the wave can reach before the onset of convective instability is [Fritts, 1984, equations (19) and (20)],

\[
A_{\text{break}} = |u - c| \approx |N/k_z|,
\]

where \( N \) is the buoyancy frequency, \( u \) the background wind and \( c \) the phase speed of the gravity wave. The dispersion relation between the intrinsic phase speed, \( N \) and \( k_z \) is a valid approximation if the intrinsic frequency of the wave is much higher than the inertial frequency. With the same amplitude at the wave source, a wave with a longer vertical wavelength (\( k_z \) smaller) will reach larger amplitude and thus higher altitude before breaking. In (8) we set the half-width of the wave packet equal to the vertical wavelength so that the wave packet will at least contain two full waves.

Three sets of numerical experiments have been done for comparative studies. The parameters used in our numerical experiments are shown in Table 1. \( C_{ge} \) is the vertical
group velocity, \(C_{px}\) is the horizontal phase velocity, \(T_G\) is the period. The wave characteristics have been chosen so that the vertical group velocities are about the same in a windless and isothermal atmosphere, they are all about 3.0 m s\(^{-1}\) as shown in Table 1. The vertical group velocities are relatively fast, so that we can ignore the vertical

<table>
<thead>
<tr>
<th>Case</th>
<th>(l_x), km</th>
<th>(l_z), km</th>
<th>(C_{px}), m s(^{-1})</th>
<th>(C_{gz}), m s(^{-1})</th>
<th>(T_G), min</th>
<th>(A), m s(^{-1})</th>
<th>(\lambda_z), km</th>
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</thead>
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<tr>
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<td>10</td>
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<td>2.96</td>
<td>29.94</td>
<td>55.66</td>
<td>3.0</td>
<td>10</td>
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<tr>
<td>Case B</td>
<td>20</td>
<td>400</td>
<td>2.95</td>
<td>59.20</td>
<td>112.62</td>
<td>4.0</td>
<td>20</td>
</tr>
<tr>
<td>Case C</td>
<td>30</td>
<td>850</td>
<td>3.06</td>
<td>86.69</td>
<td>163.42</td>
<td>5.0</td>
<td>30</td>
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</tbody>
</table>

Table 1. The Parameters of Initial GWs Perturbation

Figure 2. The spatial distribution of potential temperature field (contour line) and the dynamically (regions between shadow and dash line) and convectively unstable regions (shadow), the left and right column are the results for case A in the background without and with tide, respectively. Detailed explanation can be seen in text.
propagation of tide in the process of GWs propagation. The specific amplitudes are selected so that the waves will break at similar altitudes (~100–110 km) without tide. The momentum fluxes of the 3 cases are $1.7 \times 10^{-3}$ Pa, $1.5 \times 10^{-3}$ Pa, and $1.6 \times 10^{-3}$ Pa, respectively, and comparable to longer-term averages of the convectively generated GWs [e.g., Fritts and Alexander, 2003]. We note that the molecular diffusion is generally important above 100 km. For the waves considered in these cases, however, the molecular diffusion will become significant when it is on the order of $l_z^2 / T_G$, which translates to altitudes of 130–150 km. Therefore ignoring molecular diffusion should not fundamentally change our conclusions in these idealized simulations.

3. The Effects of Tide on GWS Propagation

[16] Both tidal wind and temperature affect the gravity wave propagation by modifying the intrinsic phase speed and buoyancy frequency, respectively, as shown in the dispersion relation of a gravity wave \(|k| \approx N|c-u|\) (second half of equation (9)). Therefore in regions where the buoyancy frequency (thus the atmospheric static stability) increases (decreases) and/or the intrinsic phase speed decreases (increase), the vertical wave number will increase

![Figure 3](image-url). Same as Figure 2, but for case B.
and the perturbative wind shear increases (decreases) accordingly.

The diurnal tidal wind and temperature used in the simulation are shown in Figure 1. For the purpose of comparative study we also simulate the nonlinear GWs propagation in the background without the tidal field. Figures 2–4 give the spatial distribution of the potential temperature field (contour lines), convectively unstable ($N^2 < 0$, shadow), and dynamically unstable ($0 < Ri < 0.25$, the regions between the shadow and dash line, $Ri = N^2/u^2_z$ is Richardson number) regions during the GWs propagation in the background without (Figures 2a, 2c, 3a, 3c, 4a, and 4c) and with tide (Figures 2b, 2d, 3b, 3d, 4b, and 4d). The dynamically unstable region is generally in close proximity to the convectively unstable region [Fritts, 1984] as shown in Figures 2–4. We should note that additional calculations have been conducted using finer horizontal resolutions, and the large-scale features are similar to the results presented here.

Now, we compare the processes of GWs propagation in the background with and without tidal wave. The horizontal phase velocity is 29.94 m s$^{-1}$ in case A, so that the critical level for the GWs packet is around 95 km for case A with tide (Figure 1). From Figure 2, we find that the

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**Figure 4.** Same as Figure 2, but for case C.
GWs become convectively unstable at about 14750 s without tide, and extensive overturning is still seen at 20,000 s. In the presence of the tide, the GWs become convectively unstable at about 8750 s, and by 15,000 s the GWs break down and no coherent wave structure is seen between 90–100 km. It should be noted that there is a positive shear layer between 90–100 km, and $|u-c|$ is approaching zero in this range. Consequently, the wave overturning amplitude is smaller (according to equation (9)) and it thus occurs earlier and is weaker in the presence of the tide. It is also worth noting that the coherent wave structure persists much longer in the absence of tide. This is

Figure 5. Vertical profiles of the horizontal mean wind (first column), Gaussian fitting mean wind (second column) and tidal wind (third column) at several times during the gravity wave propagation in the tidal background.
likely due to the weaker dissipative damping of GWs in the absence of the tide. As mentioned above, the tidal wind acts to reduce the GWs intrinsic phase speed and thus the vertical wavelength between 90-100 km. Because the dissipative damping rate is inversely proportional to the square of the vertical wavelength, the GWs with shorter vertical wavelength in the presence of the tide decay much faster.

In case B, the horizontal phase velocity of GWs is 59.2 m s$^{-1}$, so the GWs do not encounter a critical level, but the decrease of the vertical wavelength due to the increase of background wind $u$ and the decrease of $|u-c|$ affects the GWs overturning time, breaking, and the wave damping in a similar way as in case A. Figures 3a and 3b indicate that GWs begin to overturn at about 13,000 s without tide and 10,000 s with tide. At about 16,000 s, the GW breaks into turbulence extensively between 90 km and 100 km in the tidal background. On the contrary, the wave structure and strong overturning persist at 20,000 s without tide.

The results in Figure 4 indicate that the characteristics of potential temperature field and unstable ranges for GWs with 86.69 m s$^{-1}$ horizontal phase speed are similar as that in case B. In case C the difference of the initial overturning times between the two simulations with and without tidal background is 2350 s, shorter than the difference of 3000 s in case B, which is in turn shorter than the difference of 6000 s in case A.

The comparative results from the three cases show that the modulated effects of tide on the initial overturning time and height, the wave breaking amplitude, and the damping rate of the GWs are stronger for waves with shorter vertical wavelength.

4. The Effects of GWS on Tide and Background Wind

Figures 5a, 5b, and 5c present the vertical profiles of the horizontal mean wind (averaged over the horizontal domain, i.e., a horizontal wavelength, $u_{\text{mean}}$) at several times during the GWs propagation in the tidal background. The mean wind accelerations occur mainly in the height ranges 80—120 km. In comparison, the accelerations have a Gaussian shape in simulations without tide (not shown in paper). These phenomena can be observed frequently from satellite and lidar observations and result in the reversal zonal mean wind in summer at middle latitudes [e.g., Zhang and Shepherd, 2005; Li et al., 2005].

To better understand the effects of breaking gravity waves on tide, the horizontal mean wind $u_{\text{mean}}$ shown in the first column of Figure 5 is partitioned into a "tidal" part plus a "background" change. With only 5 h simulation, we cannot extract a tidal component from time series analysis. Here we assume that the background change follows a Gaussian shape, with the Gaussian peak at the altitude where the gravity wave forcing peaks, and approximate the background flow by fitting the mean wind to a Gaussian profile using least squares method ($u_{\text{fit}}$ in the second column of Figure 5). The third column of Figure 5 is the residual by subtracting the fitting mean wind $u_{\text{fit}}$ from the total mean wind $u_{\text{mean}}$, and considered here as the tidal component $u_{\text{tide}}$. This partition is somewhat arbitrary, but still enables us to qualitatively assess the effects of gravity wave breaking on the tide and background flow.

The results in Figures 5a2, 5b2, and 5c2 indicate that breaking GWs accelerate the background wind and induce a jet flow peaking at about 100 km. The peak value of the fitting mean wind increases from ~9 m s$^{-1}$ in case A to ~50 m s$^{-1}$ in case C at 17,000s. The increasing acceleration rate with the intrinsic phase speed of the wave is in general agreement with the linear saturation theory [Lindzen, 1981].

The results in Figure 5b3 and 5c3 show that the tidal amplitude is increased due to the nonlinear interactions between GWs and tide in all three cases, especially at the height range 90—120 km, with the maximum amplitude increase from the initial 45 m s$^{-1}$ to 60 m s$^{-1}$ in case C. In case C, the tidal amplitude changes are mainly confined below 100 km, because the wave encounters a critical level between 88—100 km. The third column in Figure 5 also indicates that the tidal phase shifts downward due to the nonlinear interactions between GWs and tide [Liu et al., 2000]. The temporal evolution of the total horizontal mean wind is shown in Figure 6. The net acceleration of the mean flow due to gravity wave breaking, and the tidal structure of the accelerated mean flow are evident in the figure.

To further test the GWs impact on tidal amplitude and phase, we repeat the numerical experiments by only changing the sign of the GWs phase velocities. Figure 7 presents the total mean wind $u_{\text{mean}}$, fitting mean wind $u_{\text{fit}}$ and the tidal wind $u_{\text{tide}}$ for case C. The results show the GWs breaking affects the tidal amplitude and phase in a similar fashion, though mainly through the negative (southward) branch of the tidal wind.

We have also repeated case B, replacing the isothermal atmosphere with realistic temperature profile. The onset of the instability with the realistic temperature occurs at an earlier time compared with the isothermal case. The impact on the tides is nevertheless similar between the two experiments. We have also tested the sensitivity of the simulation results to the eddy diffusion coefficient by repeating case B with different values of $K_{\text{eddy}}$. It is found that the mean
wind structures are quite similar, though the vertical shear of the mean wind and the maximum mean wind are weaker at the end of the simulation in cases with larger diffusion coefficients.

[28] It should be noted that the tidal wind and temperature background in these simulations are stationary. According to previous studies by Zhong et al. [1995] and Eckermann and Marks [1996], the gravity wave interaction with a time-varying background flow differs from the interaction with a stationary background flow because the time-varying background flow changes the apparent phase speed of the gravity wave with respect to the ground. As a result, the gravity wave impact on the tide may be over-estimated if the time variation of the background is ignored. On the other hand, the results from our simulations may still be a reasonable approximation because these high frequency gravity waves with small horizontal wave numbers (thus with large horizontal and vertical phase speeds). This is because the temporal change of the apparent gravity wave frequency is approximately proportional to \( k_x U_T / \omega \), and with \( k_x \) small and \( \omega \) large, \( \omega / dt \) becomes relatively less important. For the three cases considered here, the maximums of \( k_x U_T / dt \) (\( U_{T \text{max}} = 40 \) m/s for meridional wind component of diurnal tide) are \( 1.82 \times 10^{-7} \) rad s\(^{-2}\), \( 4.55 \times 10^{-8} \) rad s\(^{-2}\), and \( 2.14 \times 10^{-5} \) rad s\(^{-2}\), compared with the respective frequencies of \( 1.88 \times 10^{-3} \) rad s\(^{-1}\), \( 9.29 \times 10^{-4} \) rad s\(^{-1}\), and \( 6.41 \times 10^{-4} \) rad s\(^{-1}\) at the launch level.

[29] By using a compressible and nonlinear two-dimensional numerical model, and including the meridional component of the diurnal tidal wind and temperature (30°N, September) from GSWM-00 as background wind and temperature, we studied the propagation and breaking of GWs with increasing vertical wavelength in the tidal background and the impact of these waves on the tide. Compared with GWs propagation in the background without tide, the diurnal tidal wind affects the initial time of the GWs overturning, GWs breaking amplitude, and the dissipative damping rate of the GWs. The modulating effects of diurnal tidal wave on GWs with shorter (longer) wavelength are stronger (weaker).

[30] The GWs breaking not only accelerates the background wind, but also increases the amplitudes of the tidal wind amplitude for the set of GWs used in these simulations. It is more likely for GWs with smaller vertical wavelength to encounter critical layers and to increase the amplitude of diurnal tidal wind below the critical layers. GWs with longer vertical wavelength have larger horizontal phase velocity and are less likely to encounter critical level. They can thus propagate up to a higher altitude and amplify the tidal amplitude in a larger height range. The tidal phase shifts downward as a result of the GWs and tide interaction.

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