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Key Points:

- Tropical cyclone-induced gravity waves have been simulated using a general circulation model (GCM) for the upper atmosphere
- The model has been used is the 3-D global ionosphere-thermosphere model with local-grid refinement (GITM-R)
- The simulation results reveal a clear evolution of traveling ionospheric disturbances (TIDs) and show a reasonable agreement with observations

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Tropical Cyclone-Induced Gravity Wave Perturbations in the Upper Atmosphere: GITM-R Simulations

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Abstract The tropical cyclone (TC)-induced concentric gravity waves (CGWs) are capable of propagating upward from convective sources in the troposphere to the upper atmosphere and creating concentric traveling ionospheric disturbances (CTIDs). To examine the CGWs propagation, we implement tropical cyclone-induced CGWs into the lower boundary of global ionosphere-thermosphere model with local-grid refinement (GITM-R) and simulate the influence of CGWs on the ionosphere and thermosphere. GITM-R is a three-dimensional non-hydrostatic general circulation model for the upper atmosphere with the local-grid refinement module to enhance the resolution at the location of interest. In this study, CGWs induced by the typhoon Meranti in 2016 have been simulated. Information of the TC shape and moving trails is obtained from the TC best-track dataset, and the gravity wave patterns are specified at the lower boundary of GITM-R (100 km altitude). The horizontal wavelength and phase velocity of wave perturbation at the lower boundary are specified to be consistent with the TEC observations. The simulation reveals a clear evolution of CTIDs, which shows reasonable agreement with the GPS-TEC observations. This is the first time the typhoon-driven TEC perturbation has been simulated in a general circulation model. To further examine the dependence of the CTIDs on the wavelength and frequency of the gravity wave perturbation at the lower boundary, different waveforms have been tested as well. The magnitude of CTIDs has a negative correlation with the period but a positive correlation with the wavelength when the horizontal phase velocities are sufficiently fast against the critical-level absorption.

1. Introduction

Previous studies showed that extreme meteorological events such as deep convection systems (Azeem et al., 2015; Lane et al., 2003), tropical cyclones (TCs) including typhoons, hurricanes, tropical storm (Chum et al., 2018; Georges, 1973; Hung & Kuo, 1978; Xiao et al., 2007; Yue et al., 2014), thunderstorms (Curry & Murty, 1974; Taylor & Hapgood, 1988), and tornadoes (Nishioka et al., 2013) can generate acoustic waves and gravity waves, which can propagate upward and have the potential to change the upper atmosphere via momentum flux and heat deposition. Bauer (1958) first showed the influence of hurricanes on critical frequency of the F2 layer of the ionosphere. Gravity waves induced by TCs were detected in the ionosphere by high-frequency Doppler sounders (Hung & Smith, 1978) and in the mesosphere by ground-based all-sky OH airglow imager (Suzuki et al., 2013). Concentric gravity waves (CGWs) have been reported in the study of Suzuki et al. (2013) for the first time. Concentric traveling ionosphere disturbances (CTIDs) caused by super typhoon Meranti have been observed in total electron content (TEC) by global navigation satellite systems (GNSS) unprecedentedly (Chou et al., 2017). During this event, the CTIDs deduced from the band-pass filtered TEC data have wave periods of 8–30 min, horizontal wavelengths of 160–200 km, and horizontal phase velocities of 106–220 m/s.

Numerical simulations have also been devoted to study the impacts of convection-induced GWs on the upper atmosphere. Mesoscale models were employed to simulate dynamics of gravity waves triggered by isolate convective source from the troposphere (Lane & Zhang, 2011) and stratosphere (42 km) (Piani et al., 2000). Based on a linear model and ray tracing method, Vadas and Fritts (2004) approximated the propagation of gravity waves from mesoscale convective complexes in the troposphere to mesosphere-lower

thermosphere (MLT) region. Simulations of mesoscale CGWs (Vadas et al., 2009) excited from convections to the MLT region were conducted afterwards. TC-induced gravity waves were mostly simulated below 30 km (Kim et al., 2005; Kuester et al., 2008). However, the gravity waves induced by intense convections may influence the atmosphere at very high altitudes due to relatively large vertical wavelengths and high horizontal phase velocities of the gravity waves which escape the critical-level absorption (Salby & Garcia, 1987; Vadas & Fritts, 2004). Using global ionosphere-thermosphere model (GITM) (Ridley et al., 2006) a three-dimensional general circulation model for the upper atmosphere, the response of upper atmosphere to gravity waves excited by tsunami and earthquake activities has been investigated (Meng et al., 2015, 2018).

The multiscale feature of gravity waves poses a big challenge to simulate them accurately. To capture the mesoscale components (wavelength is 100–500 km), resolution higher than subdegree is desired. GITM is a self-consistent model with flexible grid sizes and allows for non-hydrostatic solutions (Deng et al., 2008), which is ideal for studying dynamics of mesoscale CGWs in the ionosphere. Currently, GITM can run globally and regionally. However, subdegree resolution for the global simulation is computationally too expensive and unnecessary. The high resolution can be easily achieved for the regional simulation, but the large-scale background features may become unrealistic after 6-hr simulation time since the open boundary conditions have been applied in latitude and longitude (Deng et al., 2018). Therefore, GITM with local-grid refinement (GITM-R) technique, which nests a high-resolution grid in the local domain of interest within the coarser resolution of the global domain, is developed to solve for mesoscale (wavelength is 100–500 km) and small-scale (wavelength is <100 km) perturbations of the ionosphere and thermosphere (IT) system (Deng et al., 2018).

In this study, GITM-R has been utilized to simulate the perturbations in the upper atmosphere over southeast of China excited by CGWs during typhoon Meranti in 2016. The typhoon-driven TEC perturbation has been simulated for the first time in a general circulation model (GCM). The simulation results of CTIDs are compared with GNSS TEC perturbations. The dependence of ionospheric CTIDs on frequency and wavelength of CGWs forcing at the lower boundary of the model has been studied as well.

2. Methodology

2.1. Model Description and Grid Setup

As a 3-D IT model, GITM is governed by continuity, momentum, and energy equations and solves for the neutral and ion densities, velocities, and temperatures self-consistently (Ridley et al., 2006). The viscosity and thermal conductivity, which are responsible for damping/dissipation of GWs, are included in the momentum and energy equations, respectively. Based on parallel algorithms, GITM-R is developed to facilitate increasing computational demands of high temporal-spatial resolution in upper atmospheric modeling. The coarse layer and local refined layer are coupled in two-way. GITM-R is ideal to simulate the dynamics of mesoscale CGWs in the upper atmosphere.

In this study, GITM-R with two layers has been used, including a global layer with horizontal resolution of 2° long $\times$ 2° lat and a refined regional layer with resolution of 0.2° long $\times$ 0.2° lat. The regional layer is in a domain of $36^\circ \times 36^\circ$ (latitude by longitude) centered at 123°E , 20°N , which is the location of the eye of the super typhoon Meranti (1614) at 13:30 UT on 13 September 2016. The vertical domain extends from 100 to 600 km with resolution of one third of the scale height. To initialize the simulation, GITM has first run without perturbations for 24 hr to reach a quasi-steady state under relatively quiet conditions.

2.2. Concentric Gravity Wave Forcing

As GITM-R is a self-consistent model covering the altitudes of 100–600 km, we need to bridge the gap between the lower atmosphere and 100 km altitude and estimate the wave forcing at the lower boundary of 100 km. Since the direct observations of gravity wave perturbation at 100 km altitude are extremely limited, the forcing has been specified using the information from the parameters of typhoon Meranti in 2016 in the troposphere, the perturbations of GNSS TEC during typhoon Meranti (Chou et al., 2017) in the ionosphere, and GW dispersion relation (Hines, 1960; Vadas et al., 2009).

The non-hydrostatic dispersion relation for the gravity wave (GW) (Hines, 1960) can be described as

$$m^2 = \frac{k^2 N^2}{\omega^2} - \frac{1}{4H^2} - k^2, \quad (1)$$

where k and m are GW horizontal and vertical wavenumber components, ω is the wave's intrinsic frequency, N is the buoyancy frequency, and H is the density scale height. For excited CGWs with $m^2 \gg \frac{1}{4H^2}$, the dispersion relation can be expressed as

$$\omega^2 = \frac{k^2 N^2}{k^2 + m^2} = N^2 \cos^2 \beta, \quad (2)$$

β is the angle of wave vector with respect to the horizontal component. The conditions for wave to be reflected and ducted are (Deng & Ridley, 2014)

$$\lambda > \lambda_n = 4\pi H \sqrt{\frac{N^2}{\omega^2} - 1}, \quad (3)$$

where λ is the horizontal wavelength of gravity wave and λ_n is the cutoff wavelengths for the non-hydrostatic condition. According to neutral density and temperature from International Civil Aviation Organization (ICAO) Standard Atmosphere from sea levels to 87 km and parameters in GITM from 100 to 600 km, we can easily calculate the scale height H and buoyancy frequency N from the sea level to 600 km, which are at the range of ~ 4 – 40 km and ~ 0.01 – 0.04 rad/s. The angle β is $\sim 60^\circ$ to 85° , and the cutoff wavelength is ~ 330 – 730 km when the frequency is 0.92 mHz. The wavelength of forcing wave at the lower boundary of simulation box is 170 km. It is significantly less than the cutoff wavelength λ_n . Consequently, the wave propagates to higher altitudes. The magnitude of the CGWs forcing at lower boundary is estimated from the relationship between convective available potential energy (CAPE) of single convective plume and horizontal velocity. In Vadas and Liu (2009), horizontal velocity magnitudes of gravity waves induced by a single convective plume with CAPE of 800 J/kg at 87 km is ~ 20 m/s when horizontal wavelength is ~ 170 km and horizontal phase velocity is ~ 150 m/s. As the CAPE of typhoon Meranti before landing can reach $1,000$ J/kg and the amplitude grows with altitude as $e^{z/2H}$ (Hines, 1960), we can roughly estimate the horizontal velocity amplitudes at 100 km as 100 m/s.

The horizontal wavelength and horizontal phase velocity of CGWs during Meranti are estimated from the time evolution of band-pass filtered total electron contents (TECs) shown in Chou et al. (2017). The observations exhibit TEC perturbation amplitude of ~ 0.3 – 0.8 TECu, wave period of ~ 8 – 30 min, horizontal wavelength of ~ 160 – 200 km, and horizontal phase velocity of ~ 106 – 220 m/s. In this study, wave with period and horizontal wavelength of ~ 18 min and 170 km has been examined initially. The dependence on the period and wavelength has also been investigated in section 3. The center of the wave is derived from the tropical cyclone best track dataset and is around 123.0° W, 20.3° N at $8:00$ UT on 13 September 2016.

Typhoon is one of the largest and most intense convective systems in the troposphere. It has the most common size of ~ 300 – 600 km and can reach 800 km and more, which is classified by the radius of the area in which the wind speed exceeds 15 m/s. The horizontal size of forcing at 100 km altitude, D_h , can be estimated from its horizontal body diameter D_{tp} in the troposphere and wave vector angle β :

$$D_h = D_{tp} + \frac{H}{\tan \beta}, \quad (4)$$

where $H = 100$ km. According to the angle β and the general typhoon size, the horizontal scale of the implemented lower boundary forcing in the simulations is 800 km. The Meranti-induced CGWs forcing at 100 km is specified as 170 km for the horizontal wavelength and 157 m/s horizontal phase velocity. The amplitude of horizontal wind perturbation is 100 m/s. The circular perturbation added at 100 km altitude is centered at 123.0° W, 20.3° N with radius of 800 km. Based on this setting, the concentric

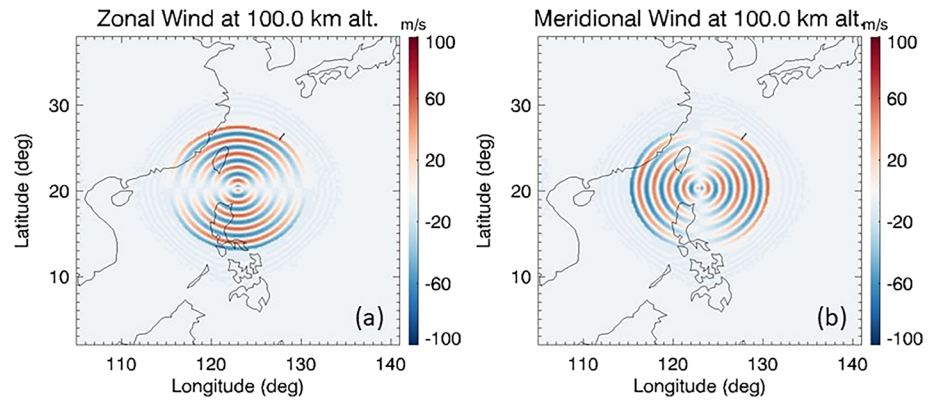


Figure 1. The perturbation forcing of zonal (a) and meridional wind (b) at the lower boundary of GITM.

cosine waves have been imposed to the horizontal winds at lower boundary of GITM, as shown in Figure 1:

$$\delta u_{x,y} = A \times \cos\left(\frac{2\pi}{\lambda}r - \omega t\right), \quad (5)$$

where A is the amplitude of CGWs, λ and ω is the horizontal wavelength and frequency, r is the distance referring to the eye of typhoon, and t is the time. Since the typhoon typically can last for several days and is a relatively long-lasting system, the typhoon-induced CGWs can stay for more than 10 hr. The GITM simulations start at 08:02 UT, and the gravity wave forcing is constant for 4 hr.

3. Results and Discussion

Utilizing GITM-R and CGWs forcing we estimated, we simulate the ionospheric perturbations over the southeastern coast of China induced by the typhoon Meranti in 2016. Different forcing patterns have also been applied to examine the sensitivity of CTIDs to the wavelengths and frequencies.

3.1. Simulation of Typhoon Meranti

Figure 2 displays the GITM-simulated total electron content perturbations (δTEC_{GITM}) around the Chinese southeastern coast at four epochs. Ionospheric slant TEC data are total number of electrons observed by terrestrial-based GNSS receivers along oblique signal path from satellites. The slant TEC can be converted to vertical TEC under certain assumptions. In our simulations, TEC_{GITM} is integrated from 100 to 350 km vertically, which is considered as a fraction of the total TEC. Above 350 km altitude, the electron density perturbation from GITM simulations conducted in this study may be under the influence of upper boundary conditions. As the largest typhoon-induced fluctuations of electron densities occur around the F2 peak, which is below 350 km, integrating to 350 km includes most of the investigated variations. δTEC_{GITM} represents the difference from the non-typhoon background case. The maximum value of $|\delta TEC|$ is 0.019, 0.064, 0.086, and 0.093 when δt is 5 min, 15 min, 35 min, and 1 hr. The magnitude increases with time, and the TEC perturbations expand radially outward from the wave center. The magnitude of the TEC perturbations is not circularly symmetric, especially at 1 hr. The TEC perturbations to the southward direction of storm center is stronger than other directions at 1 hr, which may be caused by the variation of background winds and the effect of magnetic field. When the background wind speed and wave phase velocity are comparable and in the same direction, critical level filtering will prevent GWs from propagating upward (Vadas et al., 2009). Meanwhile, the stronger perturbations in the south may be related to the preferential plasma motion toward the magnetic equator (Zettergren & Snively, 2015).

In order to understand the horizontal propagation characteristics of CTIDs, the perturbations of TEC_{GITM} are plotted as a function of distance northwestward off the wave center, as indicated by the redline in Figure 2c. Figure 3 displays horizontal propagation velocities within 4 hr after forcing imposed. The horizontal phase velocity and period vary along the time and distance. Within 4 hr, the horizontal phase velocity

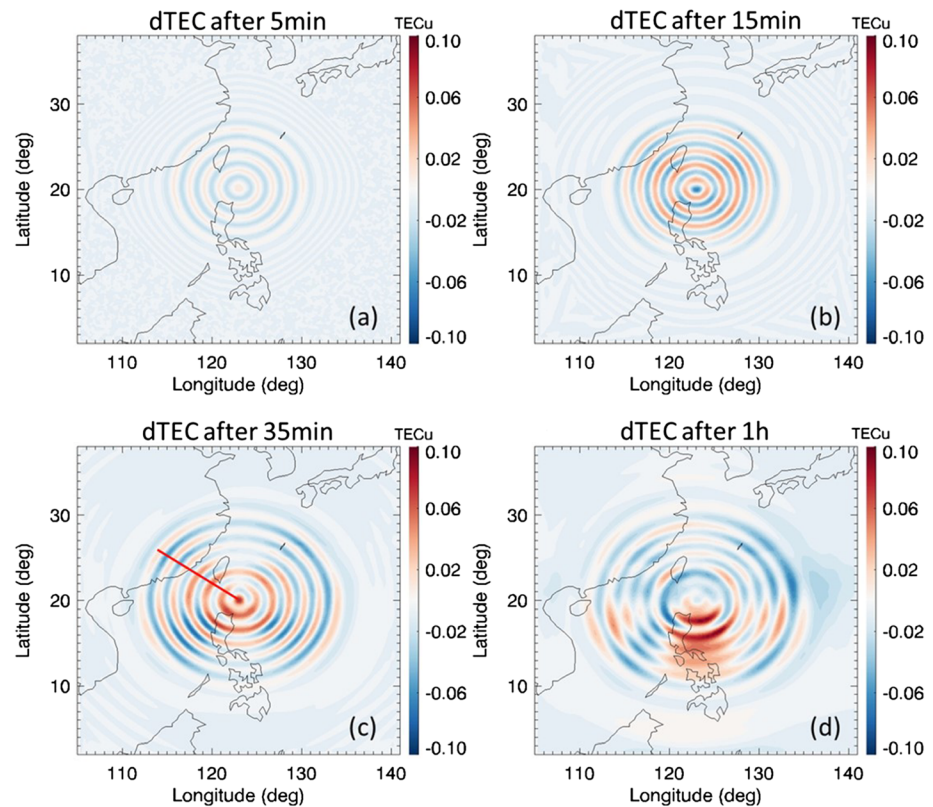


Figure 2. Snapshots of simulated δTEC_{GITM} for the typhoon Meranti at 5 min, 15 min, 35 min, and 1 hr after forcing imposed at the lower boundary.

varies from ~ 160 to ~ 300 m/s, and the period varies from 5 to 17 min. The horizontal phase velocity increases when the distance is greater than 1,000 km. From 40 min to 4 hr, within 1,000 km the horizontal phase velocity and period are about 168.83 m/s and 17 min, respectively. During this period, the magnitude of δTEC_{GITM} increases with time. The phase velocity and period shown in Figure 3 agree well with those presented in filtered GNSS TEC data (Chou et al., 2017).

Figure 4 shows the altitude-longitude view of percentage of simulated neutral density perturbations and electron density perturbations along latitude 20.3°N , the latitude of the typhoon eye, within 30 min after imposing CGWs at the lower boundary. As discussed in Deng et al. (2008, 2011), the change of forcing can trigger acoustic waves propagating upward during the first 20 min. Indeed, the signature of acoustic waves propagating vertically has been shown in Figure 4a. At 9 min, neutral density perturbations (panel a) show acoustic waves propagating upward beyond 200 km and cause up to 4.7% perturbations in neutral density.

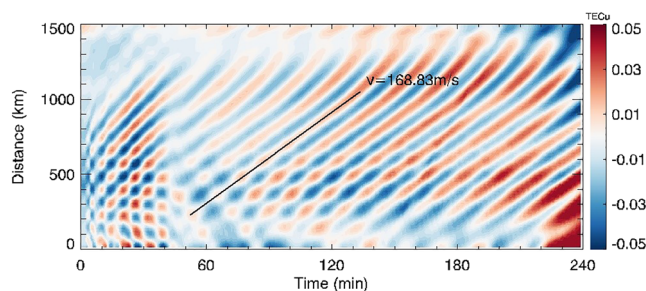


Figure 3. Temporal variation of δTEC along radial direction westward of the wave center up to 1,500 km, indicated by the redline in Figure 2. The gradient of the slant line denotes the horizontal phase velocity of the CTIDs.

The percentage of electron density at the same moment (panel c) propagates upward above 250 km and results in $\pm 0.9\%$ change in electron density. At 30 min, the acoustic waves already propagate out of the simulation domain (Deng et al., 2008), and the gravity waves cause up to $\pm 5\%$ and $\pm 1\%$ perturbations in neutral density and electron density, respectively. The electron density perturbations below 250 km are larger than above, which might be a consequence of dissipation processes caused by viscosity and thermal conductivity. The inflection in electron density perturbations at 105 and 150 km may be caused by the great height gradients of temperature and plasma density. Meanwhile, thermospheric ducted secondary wave excitation (Snively & Pasko, 2008) can play an important role in GW propagation at mesosphere and low-thermosphere altitudes. Unfortunately, the secondary forcing has not been included in our current simulations due to several reasons. First, the wave perturbation we

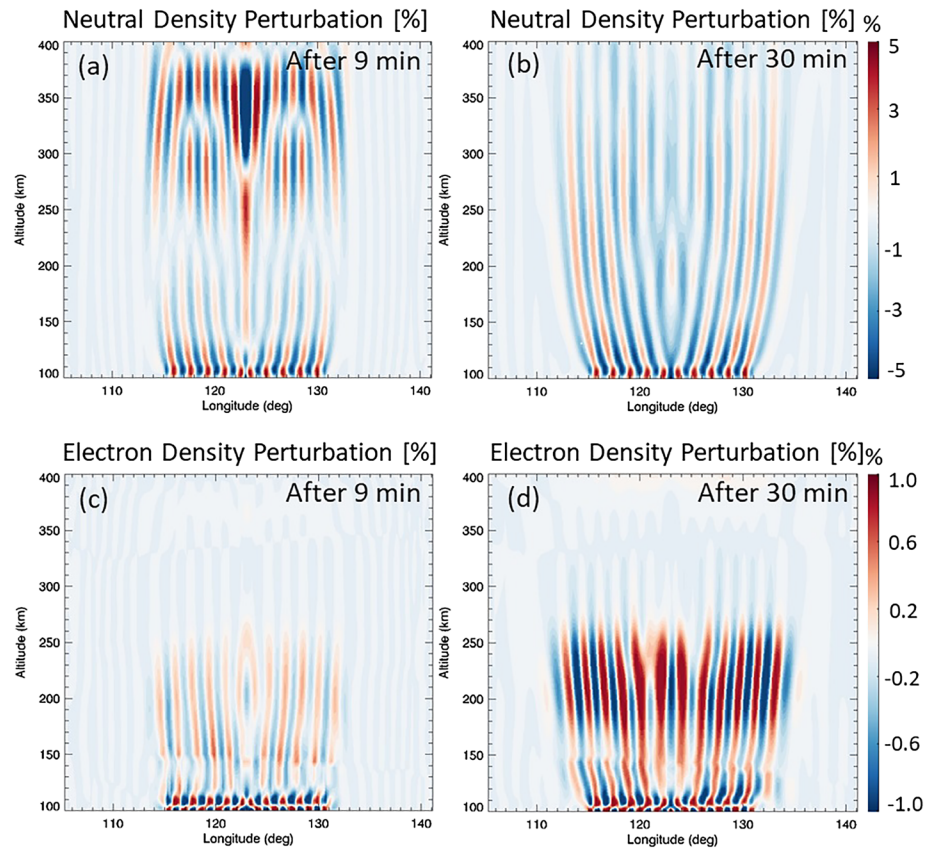


Figure 4. Snapshots of GITM-simulated neutral density perturbations (panels a and b) and electron density perturbations (panels c and d) at latitude 20.3°N from 9 and 30 min after forcing imposed.

imposed at lower boundary (period of 18 min and wavelength of 170 km) has a much larger period and longer wavelength than the short-period and small-scale gravity waves (5 min and 30 km) shown in Snively and Pasko (2008). Therefore, the secondary forcing related to ducted waves is not efficient for those waves. Second, the resolution of GITM (20 km in horizontal direction) is much lower than the resolution in Snively and Pasko (2008) (1–3 km in X direction). So the small-scale duct waves with wavelength of 30 km cannot be resolved in GITM simulations. The implication can be the underestimation of short-period and small-scale gravity waves in the GITM simulations. Since the gravity waves which influence TEC most efficiently have a wavelength around 150 km and the main goal of this study is to compare GITM simulations with GNSS TEC observations, this underestimation will not impact the main conclusion of this paper.

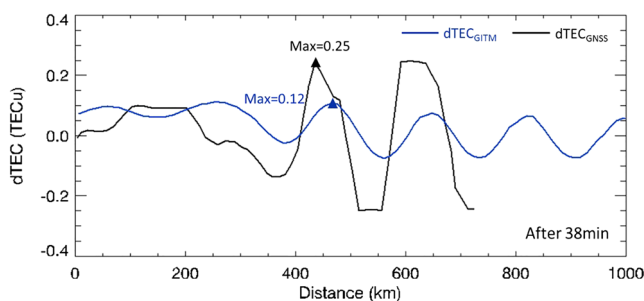


Figure 5. Comparisons between the perturbations of GITM-R simulated TEC_{GITM} (blue line) and GNSS observed TEC_{GNSS} (black line) along the radius outward from wave center of typhoon Meranti.

To compare the waveforms of CTIDs between GITM-R simulated TEC perturbation (δTEC_{GITM}) and GNSS-observed TEC (δTEC_{GNSS}) along distance, we extract the δTEC_{GNSS} from two-dimensional GNSS TEC map with a band-pass filtering in Figure 2h of Chou et al. (2017) along the radial direction from typhoon eyes (123°E, 20.3°N) to northwestward at point (120°E, 30°N). δTEC_{GITM} is also extracted along the red line in Figure 2c at 38 min. The comparison of δTEC_{GITM} (blue line) with δTEC_{GNSS} (black line) along the distance is shown in Figure 5. The wavelengths of δTEC_{GNSS} and δTEC_{GITM} are about 170 km and 180 km, and the amplitudes of δTEC_{GNSS} and δTEC_{GITM} are about 0.25 TECu and 0.12 TECu, respectively. The wavelength of simulated perturbations is in a good agreement with the observation. The amplitude of simulated perturbations is, however, about half of that in the observation. That may

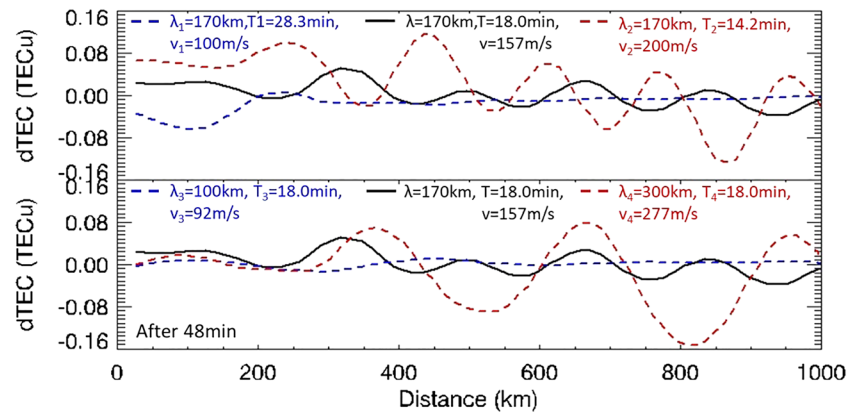


Figure 6. Comparison of GITM simulated δTEC_{GITM} after imposing typhoon Meranti waveform ($\lambda = 170$ km, period $T = 18.0$ min) with other four waveforms with different frequencies ($\lambda_1 = 170$ km, $T_1 = 28.3$ min; $\lambda_2 = 170$ km, $T_2 = 14.2$ min) shown in top and different wavelengths ($\lambda_3 = 100$ km, $T_3 = 18.0$ min; $\lambda_4 = 300$ km, $T_4 = 18.0$ min) shown in bottom.

be due to the following two reasons. First, the integration paths are different. δTEC_{GNSS} is calculated from the slant TEC, while δTEC_{GITM} is the perturbation of vertical TEC from altitudes below 350 km. Second, the typhoon-induced CGWs can disturb neutral components temperature and density as well as neutral winds at the 100 km altitude according to the polarization relationship. However, in our current study only the horizontal neutral wind perturbations have been utilized to specify the wave perturbation at the lower boundary of the simulation domain. Therefore, it is reasonable that δTEC_{GITM} shows a comparable and relatively smaller magnitude than δTEC_{GNSS} . Currently, for the real event study it is very challenging to fully specify the 2-D perturbation at the lower boundary of the model (100 km) from the observations. Our approach to impose the perturbation at the lower boundary according to the TEC observations serves as a modeling experiment. Since the nonlinear processes in the dynamics and ion-neutral coupling may play significant roles when the waves propagate from lower boundary to the F-region altitude and change the ionosphere, perturbation of neutral wind at 100 km and perturbation in TEC are not necessary to be consistent. Our results indicate that the nonlinear processes are not dominant for the particular wavelength and frequency we studied.

3.2. Sensitivity to the Wavelength and Frequency

In order to examine the dependence of the CTIDs on the period and wavelength of gravity waves at the lower boundary, different waveforms have been applied to specify the wave forcing at the lower boundary (~ 100 km altitude). The waveform of typhoon Meranti (wavelength $\lambda = 170$ km, period $T = 18$ min) is used as the control case. For the first group as shown in the top panel of Figure 6, the period varies from $T_1 = 28.3$ min in Case 1 to $T_2 = 14.2$ min in Case 2, while the horizontal wavelength kept to be the same. The δTEC_{GITM} along the radial direction (red line in Figure 2c) after 48 min are shown in Figure 6, and the amplitude increases from 0.01 to 0.13 TECu when the period decreases. For the second group as shown in the bottom panel of Figure 6, the wavelength varies from $\lambda_3 = 100$ km in Case 3 to $\lambda_4 = 300$ km in Case 4, while the period kept to be the same. The amplitude of δTEC_{GITM} increases from 0.01 to 0.15 TECu when the wavelength increases. Therefore, the magnitude of CTIDs has a negative correlation with the period but a positive correlation with the wavelength. The possible physical reason is that when the period is shorter, the forcing frequency is closer to the buoyancy oscillation frequency of the atmosphere. Therefore, the wave can more efficiently cause the disturbance in the thermosphere and ionosphere. When the wavelength is longer, the horizontal phase speed is larger. It means that the GWs can more quickly propagate out and create larger perturbation in the nearby regions. It is noticeable that the magnitudes of δTEC_{GITM} in the long period case (case 1) and the short wavelength case (case 3) are far less than those in the other cases. One possible reason is that the horizontal phase velocity of these two cases is around 100 m/s, which is comparable to background wind speed. Therefore, it is more likely for these GWs with small horizontal phase velocity to be influenced by the critical level filtering effect of the winds.

4. Conclusions

We have simulated the dynamics of TC-induced CGWs using a 3-D nonhydrostatic local-refined model in the upper atmosphere, GITM-R. The forcing of CGWs at the lower boundary (100 km altitude) is specified according to the typhoon parameters of Meranti in 2016 and the perturbations of GNSS TEC observed during this event and the GW dispersion relationship. The simulated TEC perturbations reveal a clear evolution of ionospheric CTIDs caused by TC-induced CGWs. The perturbations of TEC_{GITM} reach 0.12 TECu after 38 min with the forcing. Compared to the observed TEC perturbations, GITM-R simulations successfully reproduce the wavelength and horizontal phase velocities of the TC-induced CTIDs. Within 1,000 km radius from the center, the horizontal phase velocity and period of CTIDs gradually stabilize at 168.83 m/s and 17 min. The strongest perturbations of electron density occur below 250 km. Four additional waveforms of the lower boundary forcing have also been implemented to the lower boundary to examine the sensitivity of CTIDs to the wavelength and frequency. The magnitude of CTIDs shows a negative correlation with period but a positive correlation with wavelength.

Data Availability Statement

The modeling results are stored at <http://doi.org/10.5281/zenodo.3530952>.

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