Atmospheric and Ionospheric Responses to Acoustic and Gravity Waves Driven by Earthquakes and Tsunamis

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Atmospheric and Ionospheric Responses to Acoustic and Gravity Waves Driven by Earthquakes and Tsunamis

by

Pavel A. Inchin

A Dissertation Submitted to the Physical Sciences Department in Partial Fulfillment of the Requirements for the Degree of

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Atmospheric and Ionospheric Responses to Acoustic and Gravity Waves Driven by Earthquakes and Tsunamis

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This Dissertation was prepared under the direction of the candidate’s Dissertation Committee Chair, Dr. Jonathan B. Snively and has been approved by the members of his dissertation committee. It was submitted to the College of Arts and Sciences and was accepted in partial fulfillment of the requirements for the Degree of Doctor of Philosophy in Engineering Physics

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Abstract

Natural hazards (NHs) are causes for significant concern due to their potentially catastrophic impacts on society. The study of their effects on the upper atmosphere can provide significant insight into the coupled nature of geophysical processes, and drives important applications for radio communication and navigation, leading potentially to the development of NH early-warning systems. Although general concepts underlying their coupling mechanisms are well understood, recent computational capabilities, supported by high temporal and spatial density of observations, have reached the level where such processes can be modeled with unprecedented realism for detailed case studies. The studies provided in this thesis are aimed to advance the understanding of the mechanisms of the generation and propagation of acoustic and gravity waves (AGWs) triggered by earthquakes and tsunamis, and their effects on mesopause air-glow and ionospheric plasma.

First, we examine coupling mechanisms based on case studies for two inland earthquakes. We demonstrate that the incorporation of near-epicentral seismic wave dynamics, based on earthquake finite-fault models, provides marked improvement for the simulation of realistic AGWs and coseismic ionospheric disturbances (CIDs). Particularly, earthquake rupture propagation (and its direction) plays important role in AGW and CID asymmetries. The regime of propagation of the AGWs, driven by large earthquakes, can be weakly to strongly nonlinear, leading to substantially different dynamics in comparison with linear assumptions. Global Navigation Satellite System signals’ derived total electron content (TEC) observations may supplement
seismological studies through the investigation of finite-fault models and their ability to reproduce detected ionospheric perturbations. In this case, numerical simulations are the most comprehensive way to resolve AGW propagation through the whole range of altitudes and to subsequently reproduce CIDs accurately. Electron density depletion processes and resonance of AGWs between thermosphere and ground, leaking energy into upper atmospheric layers, can contribute to long-lived CIDs that remain observable in TEC after the event.

This thesis also investigates the characteristics and propagation regimes of AGWs driven by tsunamis, based on realistic and parametric case studies with demonstrative and simplified bathymetry variations and sources. We show that AGW propagation is markedly affected by variations in atmospheric state, as well as nonlinear effects. Substantial amplification of AGWs reaching thermosphere may lead to their self-acceleration and breaking that, along with dissipative mechanisms, leads to the excitation of secondary AGWs of a broad range of periods. Bathymetry variations, resulting in focusing, reflection and refraction of ocean waves, play a crucial role on AGW characteristics, dispersion, and amplitudes.

Finally, we demonstrate that AGWs from large earthquakes and tsunamis can be sufficiently intense to drive strong perturbations in mesospheric nighttime airglow emissions, readily measurable from ground or space by contemporary imagers. New targeted observations above regions at high risk of seismic hazards have the potential to provide an additional source of data for tsunami early-warning systems, as well as new diagnostics of surface displacements for seismological studies.
List of Acronyms

AW acoustic wave
AGW acoustic-gravity wave
SAGW secondary acoustic-gravity wave
TAGW tsunamigenic acoustic-gravity wave
GW gravity wave
RW Rayleigh wave
CID coseismic ionospheric disturbance
TEC total electron content
sTEC slant total electron content
vTEC vertical total electron content
CMT centroid moment tensor
CFL Courant–Friedrichs–Lewy condition
GPS Global Positioning System
GNSS global navigation satellite system
NH natural hazards
LOS line-of-sight
IPP ionospheric pierce point
MA mesospheric airglow
IVER integrated volume emission rate
VER volume emission rate
BWT brightness-weighted temperature
EIA equatorial ionospheric anomaly
<table>
<thead>
<tr>
<th>Acronym</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>MPI</td>
<td>message passing interface</td>
</tr>
<tr>
<td>SWE</td>
<td>shallow water equation</td>
</tr>
<tr>
<td>ITD</td>
<td>initial tsunami distribution</td>
</tr>
<tr>
<td>MAGIC</td>
<td>Model for Acoustic-Gravity wave Interactions and Coupling</td>
</tr>
<tr>
<td>GEMINI</td>
<td>Geospace Environment Model of Ion-Neutral Interactions</td>
</tr>
<tr>
<td>UT</td>
<td>universal time</td>
</tr>
</tbody>
</table>
Contents

List of Tables ix

List of Figures x

1 Introduction 1
   1.1 Natural Hazards-Atmosphere interactions ......................... 1
   1.2 Detection methods and instrumentation .............................. 2
   1.3 Motivation and Dissertation Organization .......................... 8

2 Modeling and Data Analysis Methodology 12
   2.1 Previous modeling efforts ........................................... 12
   2.2 Proposed modeling approach .......................................... 16
      2.2.1 Simulation of offset and seismic wave propagation .............. 16
      2.2.2 Simulation of tsunami propagation ................................ 19
      2.2.3 Simulation of AGW propagation in neutral atmosphere ............ 21
      2.2.4 Simulation of ionospheric plasma responses to AGWs ............ 23
      2.2.5 Coupling of numerical models and data flow ................... 27

3 Ionospheric responses to AGWs driven by the 2015 Nepal Mw7.8 Earthquake 29
   3.1 Earthquake Characteristics and Observed Coseismic Ionospheric Dis-
      turbances ....................................................................... 30
   3.2 Model Configuration ....................................................... 40
   3.3 Seismic Waves Dynamic Simulation Results .......................... 44
4 Constraining finite-fault kinematics of the 2016 M7.8 Kaikoura Earthquake from ionospheric measurements 65

4.1 The 2016 M7.8 Kaikoura earthquake 66
   4.1.1 Earthquake characteristics 66
   4.1.2 Observable TEC perturbations 69

4.2 Modeling approach and assumptions 72

4.3 Simulation results 76
   4.3.1 AW dynamics 76
   4.3.2 Ionospheric plasma responses to AWs 82
   4.3.3 Seismo-ionospheric imagery 89

4.4 Discussion 94

4.5 Conclusion and future work 99

5 The dynamics of tsunamigenic acoustic-gravity waves 102

5.1 Numerical simulation approach 103

5.2 Tohoku-Oki tsunami case study 105

5.3 Effect of tsunami scattering on TAGWs 117

5.4 Discussion and conclusion 128

6 Mesopause airglow responses to AGWs from earthquake and tsunamis 133

6.1 Introduction 133

6.2 Mesopause airglow disturbances driven by nonlinear infrasonic acoustic waves generated by large earthquakes 134
   6.2.1 Modeling approach and assumptions 134
   6.2.2 Model Simulation Results 140
   6.2.3 Discussion 145
   6.2.4 Conclusions 151

6.3 Mesopause airglow responses to tsunamigenic acoustic-gravity waves 153
6.4 Earthquake kinematics constraint based on mesopause airglow response observations 156
6.4.1 Conclusions 161

7 Conclusions and future work 163
7.1 Main outcomes 163
7.2 Future work 167
List of Tables

3.1 Kinematic slip model parameters based on Yue et al. [2016] . . . . . . . 41
3.2 Simulation configurations used. . . . . . . . . . . . . . . . . . . . . . . . . . . 42
4.1 Finite-fault models and configuration of the PF used in the analysis. . . . . 90
List of Figures

1.1 (a) 20-min high-pass filtered vTEC overlapped on a geographical map during Moore Tornado, 2013. (b) Time-latitude diagram of 20-min high-pass filtered vTEC along the longitude of Moore city (97.48°W). The figure is adapted from [Snively et al., 2021] .......................... 4

1.2 Mapped 30 min high-pass filtered vTEC observations during the 2011 Tohoku-Oki earthquake. ................................................................. 5

1.3 Wavelets and time-series of filtered (a) TEC, (b) Z-component of magnetometer and (c) seismometer observations during the 2010 M8.8 Chilean earthquake. ................................................................. 6

1.4 Earthquake/tsunami-atmosphere-ionosphere coupling processes. Earthquakes and tsunamis produce crust and ocean surface displacements that serve as a source of AGWs in atmosphere. AGWs reaching upper atmosphere trigger perturbations in airglow, ionospheric plasma and electromagnetic field that can be detected with modern instruments. ........................... 9

2.1 The scheme of models utilized in numerical simulations reported in the thesis. The arrows indicate data flow directions between models. .......................... 28
3.1 Line-of-sight vertical surface deformation after the Nepal 2015 Mw7.8 Gorkha earthquake adapted from Lindsey et al. [2015]. GCMT solution position, epicenter used in Yue et al. [2016] model and the high-rate data GNSS stations positions used for seismic dynamic and CIDs analysis at near-field region are presented. The beach ball is based on the GCMT moment tensor. The moment rate function is adopted from Yue et al. [2016].

3.2 (a) Station locations map and their IPP positions with GPS PRN23 at the region where far-field ionospheric disturbances were mostly detected (on average ~1400 km from the epicenter). (b) Wavelets and 10 min lowpass filtered data for DGRP, PBRI and CUSV GNSS stations.

3.3 (a) Tracks of IPP positions for several high-rate GNSS stations near the epicenter with GPS satellite PRN16 (except LCK4-PRN27 and NAGP-PRN27 pairs, which are marked separately). Filled circle at the end of every IPP track presents final time tag and white circle along the track - time of CIDs detection. (b) Absolute vTEC time series are shifted one to another (not equally) for better visibility. For PRN16 the tracks are structured from East (top) to West (bottom). Real absolute values scale is valid only for station RMTE and presented for reference.
3.4 Travel-time diagrams (distance of IPP positions from epicenter (ordinate) with time (abscissa)) for vTEC measurements with IPP positions to north from epicenter using (a) GPS PRN 16 and 26, (b) to south using GPS PRN 3 and 23 (satellites moving south) and (c) to south using GPS PRN 16 and 26 (satellites moving north). Detrended vTEC data (10 min high-pass filter) are presented as diamonds with different size depending on amplitude of CID. The data are shown on an oversaturated scale for better visibility.

3.5 (a) Global vTEC map (IONEX) at time of earthquake (12:00:00LT). The ionospheric region above Indian peninsula and Nepal is affected by EIA. (b) The background electron densities map from GEMINI meridional slice (shown on panel a as black line) used in Simulation (1). (c) Comparison of absolute vTEC between IONEX and GEMINI.

3.6 (a) Time-latitude diagram of surface vertical velocities along 85.35° longitude. The color scale is saturated to depict the propagation of surface waves at far distances. (b,c,d) The development of seismic waves at surface for 20, 60 and 180 s after rupture nucleation. (e) Observed and predicted seismograms, where near-field stations (KATNP, NAST, KKN4, CHLM, RMTE, SNDL) waveforms are filtered with 0.07Hz Butterworth lowpass filter; for far-field broadband seismometers (PALK, WMQ, ENH, KNDC), the instrumental response was removed and seismograms were 0.003 – 0.01Hz bandpass filtered.
3.7 (a) Energy spectral density diagram of frequency-altitude profile for the point 10 km to the north from GCMT epicenter. (b) Scaled vertical fluid velocities. ......................................................... 47

3.8 (a-d) Latitude-altitude slices of simulated scaled pressure perturbations for meridional slice along the GCMT epicenter based on Simulation (1). The epicenter latitude is indicated with a red star. Min/Max vertical velocities obtained from this simulation range from -280 to 187 m/s. (e) Oversaturated time-latitude diagram of vertical fluid velocities at altitude of 250 km for epicenter meridional slice. .......................... 50

3.9 Ions field-aligned (positive downward) drift velocities (first column), electron density perturbations (second column) and ions temperature perturbations (third column) for 3 time epochs from Simulation (1). Plasma motion at altitudes <~300km is driven by interaction with neutral species, and at higher altitudes freely propagating ion-acoustic waves can be seen. ................................................................. 53

3.10 (a) Time-Latitude diagram of simulated vTEC perturbations; (b) and (c) time series of vTEC at chosen latitudes and their wavelet power spectra. (d) Fourier transform for plot (c). (e) Oversaturated time-latitude diagram of vTEC. (f-h) Time series of vTEC at chosen latitudes shown on panel e with black dashed lines. ................................. 54
3.11 Panels (a,b,c) show nonlinear AW dynamics at near-epicentral region at fixed altitude. The dislocation of AWs from earthquake hypocenter to east is clearly seen, where there are almost no AWs propagating to west and northwest. (d,e,f) show AW dynamics for same time epochs and 1/100 magnitude of the source. In this linear regime, phase information of leading AWs is preserved.

3.12 (a,b) Vertical fluid velocities from Simulation (3) for 2 time epochs. (c) Time-latitude diagram of vertical fluid velocities.

4.1 a) "Holden-Xu" slip model of the Kaikoura earthquake. b) Comparison of observed and synthetic vertical velocity seismograms, bandpass filtered between 3 and 100 s, for selected near-field strong motion and GPS cites. Vertical displacement field obtained from (c) SAR observations and (d) forward seismic waves propagation simulation based on “Holden-Xu” model with the indication of main rupturing faults and areas. The color scale on plots c and d is oversaturated for better visibility of small features. The red star is positioned at the epicenter.

4.2 (a-b) vTEC observations filtered with 10 s–10 min bandpass Burtterworth filter and their wavelets for 0–10 min period range. (c-f) Absolute vTEC time series. (e) IPP tracks for vTEC observations shown on panels a-f and arrows represent their directions.

4.3 Simulated vertical fluid velocities at 75 and 250 km altitudes for 4 different configurations of the Holden-Xu model. Red star is epicenter position.
4.4 (a-d) Altitude-latitude diagrams of normalized pressure fluctuations for the slice along 173.65°E at 4 time epochs. Time-altitude diagrams of normalized (e) and absolute (f) vertical fluid velocities for the position of PF (42.2°S/173.65°E). Color scales are oversaturated for better visibility of weak signatures. 81

4.5 (a) Field-aligned ion drift velocities, (b) electron density perturbations and (c) ion temperature perturbations in percentage from background from preferred Holden-Xu simulation for the meridional slice along 173.65°E at T=625 s. (d) Latitude-time diagram of simulated vTEC calculated with zenith integration of electron densities. (e) Latitude-time diagram of vTEC perturbations recalculated from sTEC with 50° elevation angle of LOS pointing to south. (d) Latitude-time diagram of vTEC perturbations recalculated from sTEC with 30° elevation angle of LOS pointing to north. Data on plots are oversaturated for better visibility of weak features. (g) Absolute simulated vTEC for the position 173.98°E/40.3°S. (h) Configuration of numerical domains and direction of LOS for latitude-time diagrams shown on panels d-f. 84

4.6 (a-c) Travel-time diagrams of 10s–8min bandpass filtered vTEC to the north from the position of PF with GPS PRN21, 29 and 20. (d) Travel-time diagram of modeled vTEC perturbations for the meridional slice along 173.98°E. (e) Observed (blue) and simulated (red) vTEC perturbations. Both time series are 20s–8min bandpass filtered. 88
4.7 Latitude-altitude diagrams of vertical fluid velocity for the meridional slice along 173.98°E from simulations with models 1, 2, 3 and 5 at (a-d) T=500 s and (e-h) T=620 s. (i-l) Electron density perturbations for the same models as above at T=620 s. m) Scaled pressure signals at 150 and 250 km for the position 42.08°S/173.98°E (along dashed lines though plots a and e). Color scales at plots a-h are oversaturated by 1.5 times for better visibility of weak signatures and absolute maximum velocities are provided for each plot. 91

4.8 (a) Comparison of observed (blue) and simulated TEC perturbations based on Model 1 (red), Model 2 (black), Model 3 (green) for stations to north and northeast from the focal area. (b-d) Comparison of TEC perturbations between models 1 and 5 in 3 LOS geometry of TEC observations. 94

4.9 (a) Wave-field of maximum absolute vertical fluid velocities for the full hour of simulation for the meridional slice along 173.98°E. (b) Speed of shocks relative summed up with speed of sound for each point for the meridional slice along 173.98°E at T=660 s. (c) Field of maximum electron density perturbations calculated for the full hour of the simulation for the meridional slice along 173.98°E. 96

5.1 The snapshots of (a) ocean vertical velocities and (c-f) T’ at 4 altitudes; (g-l) absolute and scaled T’ for meridional and zonal slices shown with dashed lines on panel d. (b) Simulated ocean surface vertical displacements and DART wave gauge data. 107
5.2 Time-distance diagrams of (a,b) ocean surface velocity and (c-i) $T'$ for zonal (left column) and meridional (right column) slices at 4 altitudes. Chosen slices are shown with dashed lines on Figure 5.1,d.

5.3 Power spectral density (PSD) diagrams of ocean surface displacements and $T'$ at 4 altitudes, derived by calculating PSD for each position of the zonal slice shown on Figure 5.1. The data on plots are shown on an oversaturated scales.

5.4 (a) Bathymetry of the numerical domain. (b) The field of maximum simulated vertical ocean surface velocities. (c-f) The fields of maximum $T'$ at 4 altitudes. The data on plots are shown on an oversaturated scales for better visibility of weaker features.

5.5 Fields of maximum horizontal and vertical fluid velocities for (a,b) zonal and (d,e) meridional slices shown with dashed lines on Figure 5.4,b. (c,f) Bathymetry profiles for chosen meridional and zonal slices. The data are shown on an oversaturated scales.

5.6 (a-h) The results from 8 parametric simulations presented with altitude-distances (x-z) slices of $T'$ and corresponding $\lambda_z$ 2D wavelets, travel-time diagrams of $T'$ at 320 km, as well as $\lambda_x$ and period wavelets for slices at 320 km shown with blue lines on travel-time diagrams. Values in travel-time diagrams indicate horizontal apparent phase velocities. (i-j) Fields of maximum horizontal and vertical fluid velocities from the simulation with constantly decreasing bathymetry, (k) Time-distance diagram of $T'$ at 320 km altitude.
5.7 (a-e) Altitude-distances slices of absolute and scaled T’ and travel-time diagrams of absolute temperature perturbations at 85 and 320 km for bathymetry variation cases. Numbers of travel-time diagrams indicate horizontal apparent phase velocities in m/s. (f) Altitude-distance slices of absolute and scaled T’ for simulation with plateau. 123

5.8 The snapshots of sliced horizontally T’ at 4 altitudes from simulation with (a,b) flat bathymetry and (c,d) bathymetry with a shore, shown with dashed vertical lines. (e,f) The fields of maximum T’ at 4 altitudes for (e) flat bathymetry simulation and (f) bathymetry with a shore. 126

5.9 The snapshots of (a) ocean surface velocity and (d-e) sliced horizontally T’ at 4 altitudes from simulation with rise presence. Time epochs of snapshots are indicated on panel a. (e,f) The fields of maximum ocean surface vertical velocity and T’ at 4 altitudes. 127

6.1 (a) Model configuration used for the simulation. The results of vertical fluid velocity are shown on the slide at T=540 s from rupture nucleation using cross-section slices. (b) Background temperature and wind profiles used in simulation. Comparison of vertical fluid velocities in their maxima at the altitude of \( \sim 270 \) km from (c) current simulation and (d) Zettergren et al. [2017] (refer to Section 4). 139

6.2 (a) Map of final vertical displacements. Vertically integrated (b) OH(3,1) and (c) O(\(^1\)S) photon volume emission rate, and (d) OH(3,1) brightness-weighted temperature perturbations 435 sec after rupture nucleation. Dashed lines represent meridional and zonal slices used for keograms. 143
6.3 (a-d) Time-Latitude diagrams of simulated vertical surface velocities (a), integrated OH(3,1) (b) and O(1S) (d) emissions and temperature (c) perturbations along the longitude of ~ highest vertical surface displacements 143.67°E. (e-h) Time-Longitude diagrams of same perturbations as on (a-d), but along the latitude 38.32°N. $T_0$ represents time of rupture nucleation. Plot color scales are oversaturated for better visibility.

6.4 (a-c) Latitude-altitude diagrams of OH(3,1) and O(1S) volume emission rates at 3 moments of time. The plots are compressed horizontally to clearly represent perturbations. (d-e) OH(3,1) and O(1S) volume emission rates time-altitude diagrams.

6.5 Synthetic images of OH(3,1) integrated volume emission rates (IVER) for (a) a zenith pointing wide field (180°) imager and (b) an easward pointing imager with 40° tilt angle of 140° FOV. (c-d) Synthetic images unwarped on a geographic map and shown on an oversaturated scale for better visibility of weaker features. Black circles in plot (c) show observable regions for imagers with 120°, 140° and 160° FOVs, whereas a wide field imager covers the whole region. The yellow point in plots c,d represents the position of the imager.

6.6 (a-c) Simulated ocean surface vertical velocities and OH(3,1) and O(1S) IVER perturbations at 07:26:24 UT. (d-i) Simulated ocean surface vertical velocities and OH(3,1) brightness-weighted temperature perturbations for 3 instances of time. The plots are oversaturated for better visibility of weak signatures.
6.7 (a) Field of (a) absolute maximum vertical ocean surface velocities and (b) absolute maximum OH(3,1) BWT perturbations. The plots are oversaturated for better visibility of weak signatures.

6.8 (a-c) The snapshots of the OH(3,1) BWT perturbations in percent from background for 3 moments of time from Simulation 1. (d-f) Perturbations of OH(3,1) BWT and integrated volume emission rates of the OH(3,1) and the O(^1S) at T_0+505 sec.

6.9 Distance-time diagrams of (a,b) surface vertical velocities, (c,d) the OH(3,1) and (e,f) the O(^1S) IVER perturbations for meridional (top) and zonal (bottom) slices from Simulation 1. The plots are oversaturated for better visibility of weak signatures.

6.10 The snapshots of perturbations in OH(3,1) BTW and OH(3,1) IVER at T_0+300 sec and O(^1S) IVER at T_0+340 sec from 3 simulations discussed in text.

6.11 The snapshots of perturbations in the OH(3,1) BTW and the OH(3,1) and O(^1S) IVER from 3 simulations discussed in text. Time epochs of snapshots are indicated in each panel.
Chapter 1

Introduction

1.1 Natural Hazards-Atmosphere interactions

Natural hazards (NHs) serve as a source of disturbances to the solid (earthquakes, landslides), liquid (tsunamis) or gaseous (tornados, hurricanes, volcanic eruptions) envelopes of the Earth. Through the coupling from the surface to the troposphere, these disturbances can drive acoustic and gravity waves (AGWs). The decrease of atmospheric density with altitude results in exponential growth of AGW amplitudes that can be strong enough to drive detectable neutral and charged particle disturbances in the upper atmosphere and dynamo effects in the ionosphere, resulting in fluctuations in electromagnetic fields.

The pioneering works of Hines [1960] and Hooke [1968] provided an important theoretical basis for the investigations of NH impacts on the upper atmosphere. Presently, disturbances in airglows and ionospheric plasma driven by AGWs are routinely de-
1.2 Detection methods and instrumentation

Systematic observations and investigations of atmospheric and ionospheric dynamics have been made since the 1960s, but only in the last decades have spatial and temporal resolution of data become sufficient to detect small-scale and short-time processes, such as AGWs driven by NHs. Remote ground-based sensing instrumentation, such ionosondes and dynasondes, airglow imagers and LIDAR systems, incoherent backscatter radars have begun to provide an important insight into AGWs dynamics in the upper atmosphere and ionosphere [Chum et al., 2012; Mabie et al., 2016; Occhipinti et al., 2018; Williams, 1989]. In addition, in-situ and remote sensing observations of upper atmosphere made from space-based platforms, for example using Langmuir probes [Croskey et al., 2006], electrometers and magnetometers [Aoyama, 2017; Berthelier et al., 2006], Global Navigational Satellite System (GNSS) radio occultation and reflectometry techniques [Coïsson et al., 2015; Ruf et al., 2016], have also proved their applicability in NH-atmosphere-ionosphere studies.

One of the primary, most accessible and robust sources of information on the ionospheric state and its perturbations, can be extracted from group and phase delays of
GNSS signals that propagate through the dispersive ionospheric medium [Parkinson et al., 1995]. The derived integrated descriptive quantity represents the total electron content (TEC) along the line-of-sight (LOS) between GNSS satellite and receiver and provides fairly accurate measurements of electron density. The ionospheric F layer (∼200-400 km), exhibiting maximum electron density concentration and thus markedly impacting GNSS signals, is a layer of observable perturbations in TEC driven by AGWs from NHs [Zettergren and Snively, 2015]. Thus, for the calculation of the ionospheric pierce point (IPP), that represents the position of the intersection of LOS and ionospheric F layer (that is approximated as a shell layer at a constant altitude of peak electron density) and determines the coordinates of the TEC, the altitude of ∼300 km is often used [Hofmann-Wellenhof et al., 2012].

Networks of ground-based GNSS receivers have increased in density throughout the Globe, especially where there are high risks of NHs; these serve as a source of seismic and geodynamic information. Among them are the Japanese GEONET network of more than 1300 Global Positioning System (GPS) receivers, the Chilean GNSS network (130 stations), and the GNSS networks in the continental United States that operate more than 3000 stations. The deployment of new and the improvement of current GNSS constellations may markedly improve the spatial resolution of data.

In addition to low temporal resolution of measurements of 15-60 sec, providers now often collect 1 and 0.2 sec sampling rate data, which is particularly important for the investigation of short-period dynamics in the ionosphere (with resolvable features ∼10 s period). These data are mostly freely available through global and regional services [Báez et al., 2018; Dow et al., 2009].
A demonstrative event for NH-ionosphere coupling processes is the 2013 Oklahoma EF5 category tornado (TEC responses shown in Figure 1.1). Oscillations of 3-4 minute periods, driven by acoustic waves (AWs) to the south of the tornado, and long-period (~15 min) concentric wave signatures to the north, west and east, generated by gravity waves (GWs), were observed in TEC for at least 8 hours [Nishioka et al., 2013; Snively et al., 2021]. The southward propagation of TEC perturbations from AWs is explained by the dominant plasma mobility along magnetic field lines in the F layer of the ionosphere [Schunk and Nagy, 2009]. For GWs, the motion of fluid has complex dynamics and plasma perturbations should, in general, be observed in all directions. However, concentric perturbations to south were not observed. It is suggested that this restriction may be caused by background wind filtering effects.

Ionospheric perturbations, driven by AGWs, are routinely observed in TEC after large earthquakes and tsunamis [Jin et al., 2015; Komjathy et al., 2016]. One of the most studied cases is the 2011 M9.1 Tohoku-Oki earthquake (Figure 1.2). Three
1.2. DETECTION METHODS AND INSTRUMENTATION

types of disturbances (CID) were discerned: 1) CID of \( \sim 3.5 \) min period and 0.9 km/s horizontal apparent phase velocity \( v_p \) that were driven by AWs from permanent displacements over the focal area, 2) CID of \( \sim 3 \) min period and \( v_p \sim 3.4 \) km/s from AWs generated by seismic waves and 3) long-period CID (\( \sim 15-20 \) min period and \( v_p \sim 0.2 \) km/s) from tsunamigenic acoustic-gravity waves (TAGWs). TEC perturbations over the focal area were detected for at least 4 hours after the earthquake and are proposed to be generated by AWs trapped between the lower thermosphere and ground while tunneling to higher altitudes [Saito et al., 2011]. Quasi-permanent electron density depletion of \( \sim 6 \) TECu was observed over the epicenter and attributed to the nonlinear plasma responses to AWs [Tsugawa et al., 2011].

The coupled processes between NHs and Earth’s geophysical systems can also be studied through observations of the perturbations in the Earth’s geomagnetic field [Aoyama, 2017; Hao et al., 2013; Zettergren and Snively, 2019]. The divergence of perpendicular dynamo currents in the ionosphere, driven by AGWs, leads to the generation of field-aligned currents, that in turn trigger detectable disturbances in geomagnetic field. The fluctuations in TEC, geomagnetic field at the

Figure 1.2: Mapped 30 min high-pass filtered vTEC observations during the 2011 Tohoku-Oki earthquake.
1.2. DETECTION METHODS AND INSTRUMENTATION

ground level, and in seismic data in the vicinity of magnetometer’s station after the 2010 M8.8 Chilean earthquake are shown in Figure 1.3. The disturbances in geomagnetic field have approximately the same periods as registered in TEC, and appeared almost simultaneously, which suggest a common forcing mechanism. Vibrations of the surface, driven by seismic wave propagation, were registered several minutes earlier than observed geomagnetic perturbations and thus could not be their source.

Some sparse satellite observations of upper atmosphere disturbances from co-seismic and tsunamigenic AGWs, as well as meteorological AGWs, have recently been reported. Tulasi Ram et al. [2017] reported the detection of ionospheric disturbances using radio-occultation COSMIC mission data after the 2015 M7.8 Nepal earthquake. Strong tsunami–induced \( \text{O}_2(\text{\textsuperscript{1}}\Sigma_g^+) \) airglow emission perturbations (up to \( \sim \)10\% from the ambient emission) were detected after the 2011 Tohoku-Oki and the 2015 Illapel tsunamis with the use of broadband emission radiometry instrument on board the TIMED satellite [Yang et al., 2017]. Plasma density perturbations were observed over Alaska with Gravity Recovery and Climate Experiment (GRACE)

Figure 1.3: Wavelets and time-series of filtered (a) TEC, (b) Z-component of magnetometer and (c) seismometer observations during the 2010 M8.8 Chilean earthquake.
spacecrafts [Yang et al., 2014] after the 2011 Tohoku-Oki earthquake. Aoyama [2017] performed an analysis of geomagnetic field fluctuations during 54 typhoons using SWARM satellite magnetometer data and showed their applicability for the study of related plasma perturbations. Yue et al. [2019] reported the observations of perturbations in the space-based near-infrared spectral imager over the thunderstorm.

Whereas ground-based TEC observations remain a main source of data for the detection of perturbations driven by NHs [Ducic et al., 2003; Galvan et al., 2012; Reddy et al., 2016], mesospheric and ionospheric airglow imager data could be used as a valuable source of information. Makela et al. [2011] reported ionospheric plasma perturbations over Hawaii using a 630.0 nm thermospheric airglow imager after the 2011 Tohoku-Oki tsunami. Suzuki et al. [2013] demonstrated the detection of AGWs in mesosphere hydroxyl (OH) airglow during a typhoon using an all-sky imager. Smith et al. [2015] reported imaging measurements of thermospheric gravity wave signatures in 630 nm and 557.7 nm nightglow associated with the 2011 Tohoku-Oki tsunami. Simultaneous observations of AGWs in Atmosphere Infrared Sounder 4.3 μm band data in the stratosphere and ground-based Midlatitude Allsky-imaging Network for GeoSpace Observations, generated by a series of convective systems over the midwestern United States, were recently discussed in Heale et al. [2019].

Examples discussed above demonstrate the complexities related to the observations of upper atmosphere responses to AGWs generated by natural hazards. At the same time, these and other recent observational reports indicate their applicability for the investigation of small-scale and short-lived coupled geophysical processes.
1.3 Motivation and Dissertation Organization

Different observational techniques and instruments can be used to detect and study NH-atmosphere-ionosphere coupling processes. However, there are several problems that are challenging to address with the use of available data. Specifics of measurement techniques impose marked limitations for detailed analyses in many cases. For example, the integrated nature of TEC and airglow observations results in a dependence of measured perturbations from mutual geometry of LOS and the direction of CID propagation. The motion of TEC LOS also leads to the measurement of Doppler-shifted values. The success of an airglow observation campaign fully depends on cloud absence. Sparse station coverage and usually low sampling rate of data are also limitations. Satellite measurements provide sparse and short period observations under regions of interest and cannot yet be used for the detailed investigations of small scale AGWs and atmospheric and associated ionospheric disturbances.

In addition, available data uncover only a small part of the dynamics of coseismic/tsunamigenic AGWs at far distances from their sources. For example, the propagation of AWs from the ground can include the formation of shock waves and their confluence below the thermosphere [Chum et al., 2016], whereas GWs can be markedly filtered, reflected, ducted or become evanescent prior reaching the upper atmosphere [Fritts and Lund, 2011]. The insufficiency of observations may suggest marked complications to the interpretation of the characteristics of AGWs and their sources, leading to uncertain conclusions on the nature of coupling. Thus, modeling is an important and necessary step for deeper understanding of atmosphere-ionosphere responses to AGWs, as well as understanding how we can observe them.
Studies reported in this thesis are focused on ground-level NH sources of AGWs, in particular large earthquakes and tsunamis. A schematic representation of the investigated geophysical processes is provided in Figure 1.4. The complex nature of earthquake/tsunami-atmosphere-ionosphere coupling mechanisms require the incorporation of various geophysical processes in a joint analysis of observable phenomena. Both these NHs serve as a source of transient ground-level displacements, that perturb the underlying atmosphere. In addition, permanent surface deformations, generated by earthquakes at regions near their epicenters, can produce substantial transfer of momentum and energy to the atmosphere. AGWs, generated in the atmosphere and exhibiting with periods of fractions to thousands of seconds, propagate upward and can be detected in upper atmosphere with modern instruments.

**Figure 1.4:** Earthquake/tsunami-atmosphere-ionosphere coupling processes. Earthquakes and tsunamis produce crust and ocean surface displacements that serve as a source of AGWs in atmosphere. AGWs reaching upper atmosphere trigger perturbations in airglow, ionospheric plasma and electromagnetic field that can be detected with modern instruments.
1.3. MOTIVATION AND DISSERTATION ORGANIZATION

The scope of this thesis addresses the following focused objectives connected with the investigation of earthquake/tsunami-atmosphere-ionosphere coupling processes including studies of:

1. **Excitation and propagation of acoustic and gravity waves generated by crust and ocean surface displacements that arise from earthquakes and tsunamis;**

2. **Mesospheric airglow and ionospheric plasma responses to realistically-driven AGWs from earthquakes and tsunamis;**

3. **The characterization of earthquake and tsunami sources based on modeling and observations of upper atmosphere responses to AGWs.**

Chapter 2 provides a description of the models and modeling approach utilized. Results from a realistic case study of ionospheric plasma responses to AGWs driven by the 2015 M7.8 Nepal Gorkha earthquake are provided in Chapter 3. It is demonstrated that the incorporation of near-epicentral seismic wave dynamics, based on earthquake finite-fault models, provides marked improvement for the simulation of realistic AGWs and CIDs. The results of the 2016 M7.8 Kaikoura earthquake case study are presented in Chapter 4, where it is shown how TEC observations can supplement seismological studies through the investigation of reproducibility of detected CIDs with different finite-fault models. Chapter 5 includes results from studies of the dynamics of tsunamigenic AGWs, suggesting that their excitation and propagation are markedly affected by heterogeneities in the atmospheric state and nonlinear effects, as well as tsunami evolution under the effect of bathymetry variations. A
modeling investigation of mesopause airglow (MA) responses to AGWs driven by large earthquakes and tsunamis are presented in Chapter 6. It is demonstrated that MA observations can supplement seismological studies and be useful for early-warning tsunami systems. Chapter 7 summarizes main outcomes and contains the scope of future works.
Chapter 2

Modeling and Data Analysis

Methodology

2.1 Previous modeling efforts

Previous modeling efforts have simulated AGW propagation and ionospheric disturbances due to inland and undersea earthquakes and tsunamis. Their results reveal important features that support theoretical predictions and hypotheses. Here, we shortly review previous and current modeling efforts and approaches for simulation of earthquake/tsunami-atmosphere-ionosphere coupled process and point to their advantages and drawbacks.

In the majority of modeling studies of coseismic AGWs and CIDs, ground-level sources of displacements are usually highly simplified. Matsumura et al. [2011a] reported simulation results of plasma perturbations driven by AGWs from an impulsive upward
2.1. PREVIOUS MODELING EFFORTS

surface motion, with the use of two-dimensional numerical model for the solution of Navier-Stokes equations. Rolland et al. [2011] simulated CIDs driven by AWs generated by Rayleigh waves (RW) after the 2008 M7.9 Wenchuan and 2003 M8.3 Tokachi-oki earthquakes, using Earth normal mode modeling and a simplified neutral-plasma coupling approach. Zettergren and Snively [2015] presented modeling and analyses of ionospheric plasma density field responses to different time-dependent near-surface sources of AGWs. Zettergren et al. [2017] reported results of CID simulations, including the formation of plasma depletion regions after the 2011 M9.1 Tohoku-Oki earthquake, using an axisymmetric AGWs source based on initial sea surface displacement data. For the same event, Zettergren and Snively [2019] presented results of TEC and magnetic field fluctuations driven by AGWs with a new 3D multifluid-electrodynamic ionospheric model and a rotated axisymmetric neutral atmosphere AGWs source. Meng et al. [2018] presented results of near-epicentral CID simulations with a point representation of the AGW source based on actual seismic data for the 2011 M9.1 Tohoku-Oki and the 2015 M8.3 Chilean earthquakes.

The simulation of surface displacements near to earthquake epicenters requires the knowledge of spatial and temporal characteristics of rupture propagation, as well as information on Earth’s interior structure [Igel, 2017; Lay and Wallace, 1995]. Assumptions based on maximum surface vertical velocities, and the time and spatial scales of motion, can serve as a good starting point, but cannot provide realistic source frequency response information, which may markedly differ from earthquake to earthquake. This is particularly important for the simulation of near-epicentral AW propagation, which includes the formation of shock waves, effects of thermoviscous dissipation, reflection/refraction and ducted resonances. Point or axisymmetric
2.1. PREVIOUS MODELING EFFORTS

(evenly distributed) sources do not incorporate the comprehensive surface deformation driven by the earthquake faulting processes, as well as seismic wave propagation. Such simplified sources also assume spatially frozen ground surface motion; whereas, in reality, the sources of AGWs include rupture propagation at supersonic speeds. These simplified sources are thus rarely applicable for the simulation of near-field AGWs, especially for strong earthquakes (which usually trigger detectable CIDs), when the faulting processes can take several minutes and be distributed over hundreds of kilometers. As shown in this thesis, realistic representation of AGW sources can improve the spatial and temporal quality of the simulation, and fidelity to observations, and thus be helpful for deeper understanding of coupling processes.

Tsunamis are ocean surface waves from different sources such as earthquakes, landslides, meteorological conditions and man-made activity [Mei et al., 2018] and are a well known mechanism of AGW generation in the atmosphere [Davies and Baker, 1965; Godin et al., 2015; Peltier and Hines, 1976]. The majority of large tsunamis are driven by undersea earthquakes from permanent bottom displacements [Levin and Nosov, 2016a], and the main features of excited by them, tsunamigenic acoustic-gravity waves (TAGWs), are widely investigated [Artru et al., 2005; Huba et al., 2015; Occhipinti et al., 2013; Vadas et al., 2015]. Among them, Hickey et al. [2009] quantitatively described AGW dynamics generated by a tsunami-like source using a linearized, steady-state Navier-Stokes equations model and derived TEC perturbations. Simulation results of TAGWs driven by the 2004 Sumatran and the 2011 Tohoku-Oki tsunamis in a 3D neutral-plasma coupled model, based on the Boussinesq approximation without thermoviscous dissipation, were published in Occhipinti et al. [2006, 2011]. The impact of vertically varying background and tidal wind structures
2.1. PREVIOUS MODELING EFFORTS

On TAGW propagation was investigated with a series of idealized background wind profiles and the use of a 2D nonlinear anelastic neutral atmosphere model [Laughman et al., 2017]. Kherani et al. [2012] performed 3D numerical simulations of the atmospheric and ionospheric perturbations for the 2011 Tohoku-Oki tsunami.

Despite having a long history of research, there remains a lack of comprehensive modeling studies that incorporate realistically varying tsunami waves coupled with compressible and nonlinear atmosphere-ionosphere models, thus spanning from ocean to the upper atmosphere. A simplified TAGW source setup [Laughman et al., 2017; Meng et al., 2015], incomplete or simplified sets of equations in models [Kherani et al., 2012; Occhipinti et al., 2011], the consideration of steady-state [Hickey et al., 2009] or linear [Vadas et al., 2015] dynamics may suggest marked under- or overestimation of TAGW amplitudes, due to insufficient accounting for nonlinearity leading to evolution of their spectra or dissipation. TAGW excitation and propagation are also effected by the evolution of source characteristics resulting from tsunami interaction with bathymetry and shores, interference, dispersion and nonlinear effects. These are not sufficiently captured in the majority of studies related to TAGW dynamics.

Finally, AGW propagation is controlled by spatial heterogeneities of atmospheric state parameters that may result in wave reflection, refraction and tilting [Broutman et al., 2014; Heale and Snively, 2015; Wu et al., 2016], tunneling [Sutherland and Yewchuk, 2004; Yu and Hickey, 2007], ducting [Chimonas and Hines, 1986], damping and interference [Sutherland, 2006; Vadas, 2007]. Nonlinear evolution, breaking, self-acceleration and dissipation can be of particular importance for AGW propagation in atmosphere [Fritts et al., 2015; Heale et al., 2014; Walterscheid and Schubert, 1990].
Thus, the use of stable, nonlinear, neutral atmosphere and ionosphere models, which resolve strong atmospheric disturbances correctly and applicably for simulation of short period AGWs and CIDs, is an important factor in successful reproduction of observed phenomena.

2.2 Proposed modeling approach

In this section, we provide the methodology used for the numerical simulations of earthquake/tsunami-atmosphere-ionosphere coupled processes and an overview of the models. As will be shown through the following studies, numerical simulations are the most direct way to realistically resolve wave propagation, in Earth’s interior or ocean, neutral atmosphere and ionosphere, and subsequently to replicate airglow and ionospheric plasma responses to AGWs.

2.2.1 Simulation of offset and seismic wave propagation

Studies of seismic waves propagation at far-fields from earthquake epicenters are successfully carried out with normal mode summation techniques and point source approximations. However, accurate reproduction of near-field surface displacements, and corresponding AGWs in the atmosphere requires the consideration of finite-fault models [Igel, 2017]. Near-field studies should include coseismic AGW generation from both permanent and transient surface displacements and, in general, require good knowledge of the spatial and temporal characteristics of rupture propagation and slip rates. Accurate resolution of the speeds and amplitudes of the excited seismic waves depends on the knowledge of regional and global crust and mantle structures.
2.2. PROPOSED MODELING APPROACH

These requirements can be met with the use of the SPECFEM3D software package that allow the simulation of near- and far-field surface displacements in forward seismic wave propagation simulations with kinematic or dynamic finite-fault sources [Komatitsch and Tromp, 2002a, b].

The SPECFEM3D model is based on the numerical solution of a linear equation of motion in a weak formulation:

$$\int_{\Omega} \rho w \cdot \partial_t^2 s \, d^3r + \int_{\Omega} (\nabla w : T) \, d^3r - \int_{S} (T \cdot \hat{n}) \cdot w \, d^2r = - \int_{\Omega} M : \nabla w(x_s) S(t) \, d^3r \quad (2.1)$$

where $\rho$ - density, $s$ - displacement field, $T$ - stress tensor, $w$ - any test function, $S(t)$ - source time function and $M$ - moment tensor.

Equation 2.1 is valid for elastic and anelastic earth models with an appropriate expression for $T$, determined for the entire strain history as:

$$T(t) = \int \partial_t c(t - t') : \nabla s(t') \, dt' \quad (2.2)$$

where $c$ - elastic tensor.

The model’s numerical scheme is based on the spectral element method (SEM). SEM is a continuous Galerkin technique and uses Lagrange polynomials of high order (typically 5-10 degree) as basis functions [Gopalakrishnan et al., 2007]. In SPECFEM3D, the interpolation of these functions is done with Lagrange polynomials, defined on Gauss-Lobatto-Legendre (GLL) collocation points. GLL points are $N + 1$ roots of the Legendre polynomial $P_N$ of degree $N$.

Earth’s ellipticity and rotation, topography, ocean effects, gravity load effects, anisotropy and attenuation can be incorporated to the simulations. The model also allows incor-
2.2. PROPOSED MODELING APPROACH

poration of pre-defined or custom 1D, 2D and 3D global and regional crustal, mantle and core models. As its initial state, the Earth is in the hydrostatic balance.

The accuracy of simulation results are determined by the number of grid points per shortest wavelength. We perform the simulations with a parallel Message Passing Interface (MPI) version of the code. The time series of displacements, velocities or accelerations are stored in Seismic Analysis Code (SAC) binary format and can be analysed and processed using the appropriate tools [Goldstein, 2003].

Source setup and simulation result validation

As an input to the forward seismic wave propagation simulations, we use set point sources that represent earthquake kinematics. Each point source is defined by its position (latitude, longitude and depth), time of initiation, duration (rise time) and moment tensor. Kinematic slip models can be obtained through open archives and catalogs, for example provided by the United States Geological Survey (USGS) and Earthquake Source Model Database (http://equake-rc.info/srcmod/), or in an appropriate literature. In the following chapters, we indicate the sources of kinematic slip models used, their derivation and assumptions.

In our studies, the results of seismic waves propagation simulations are validated using different instrumentation: Interferometric synthetic-aperture radar (InSAR), GPS, strong-motion, fault offsets and tsunami data from open archives such as UNAVCO, Incorporated Research Institutions for Seismology (IRIS), United States Geological Survey (USGS) or by the request to regional data providers. The details of the data used are provided in chapters 3 and 4 related to case studies.
2.2. PROPOSED MODELING APPROACH

2.2.2 Simulation of tsunami propagation

The modeling of tsunamis can be performed with different approaches, such as the solution of the compressible ocean fluid dynamics equations \cite{Lotto and Dunham, 2015; Nosov, 1999], Boussinesq-type equations \cite{Pophet et al., 2011; Wei et al., 1995] or shallow water equations \cite{Liu et al., 1995; Titov and González, 1997]. Whereas the first two methods incorporate the propagation of waves in the ocean interior, the third type of the solution is historically used, a sufficient and comparatively computationally inexpensive method for the simulation of long-wavelength tsunami propagation over varying topography and bathymetry \cite{Levin and Nosov, 2016b].

Shallow water equations (SWE) are based on a hydrostatic assumption - the equilibrium of fluid particles in a vertical direction, resulted from the balance of gravity and vertical pressure gradient (vertical acceleration of particles is assumed to be negligible). This is justifiable when the tsunami wavelength is much longer than the depth of the ocean. SWE does not usually take into account the dispersion of waves, though this effect can be important for transoceanic studies, narrow or short-period non-seismic sources and undular bores \cite{Glimsdal et al., 2013; Løvholt et al., 2010].

We use the GeoClaw software \cite{Berger et al., 2011; George, 2008], which numerically solves 2d depth-averaged nonlinear shallow water equations for simulation of the flow over varying bathymetry:

\[
\begin{align*}
    h_t + (hu)_x + (hv)_y &= 0 \\
    (hu)_t + (hu^2 + \frac{1}{2}gh^2)_x + (huv)_y &= -ghB_x - Dhu + \psi_o \\
    (hv)_t + (hv^2 + \frac{1}{2}gh^2)_y + (huv)_x &= -ghB_y - Dhv + \psi_o
\end{align*}
\]
where $h$ - fluid depth, $u$ and $v$ - horizontal vertically averaged fluid velocities, $B$ - seafloor elevation relative to mean sea level, $g$ - gravity constant, $D$ - friction drag coefficient and $\psi_\omega$ - Coriolis force. The bottom friction term is set as:

$$D = \frac{gM^2(u^2 + v^2)^{\frac{3}{2}}}{h^{\frac{5}{2}}} \quad (2.4)$$

where $M$ - Manning Coefficient.

The GeoClaw numerical scheme is based on Clawpack finite volume methods with an augmented Riemann problem solver that allows for the incorporation of the bathymetry term directly to the solution of conservative part of eq. 2.3 [Clawpack Development Team, 2002; George, 2008]. The friction source and Coriolis force terms are solved through time-split numerical schemes, respectively. The code is parallelized with OpenMP and simulations can be run with adaptive mesh refinement (AMR).

**Source setup and simulation result validation**

In order to drive tsunamis in GeoClaw, it is necessary to generate seafloor displacements and the resulting initial tsunami distribution (ITD). SWE presumes that the displacements at the ocean surface are produced instantaneously as at the seafloor, which is a fairly accurate assumption for large undersea earthquake, for which ocean depth and rupturing time are much less than the tsunami propagation velocity. In the tsunami simulations reported in this thesis, the results from the forward seismic wave propagation simulations are filtered with an appropriate low-pass filter (Kajiu-raa filter), to remove short-wavelength components ($k>1/h$) [Kajiura, 1963; Saito and Furumura, 2009] that dissipate in the ocean interior and do not affect the ITD.
2.2. PROPOSED MODELING APPROACH

The validation of tsunami simulation results can be performed using global and regional networks of coastal tide and offshore GPS wave gauges and seafloor bottom pressure recorders. As our interest is concentrated around tsunami propagation in the open ocean, the main source of data for tsunami simulation validation is the National Oceanic and Atmospheric Administration National Data Buys Center (NOAA NDBC) archive. NOAA NDBC provides real-time sea surface height data with 15 and 1 min and 15 sec sampling time interval from Deep-ocean Assessment and Reporting of Tsunami stations. These data are freely available and can be downloaded through NOAA NDBC online services.

2.2.3 Simulation of AGW propagation in neutral atmosphere

Accurate resolution of AGW propagation from the ground to the thermosphere requires the reproduction of the variability of atmospheric constituents with altitude, exponential decrease in density and intensification of thermoviscous dissipative mechanisms and the drastic increase of temperature in the thermosphere. Background winds, that may markedly affect the propagation of AGWs, should also be included in simulations. Capturing nonlinear wave evolution is of particular importance for the simulation of AGW propagation in the upper atmosphere.

For the simulation of propagation of AGWs in neutral atmosphere, we use the Model for Acoustic-Gravity wave Interactions and Coupling (MAGIC). This model was earlier used and validated for the investigation of small-scale AGW dynamics [Burleigh et al., 2018; Heale and Snively, 2015; Snively, 2013; Zettergren et al., 2017].

MAGIC solves the nonlinear compressible Navier-Stokes equations for conservative
2.2. **PROPOSED MODELING APPROACH**

quantities - density, momentum and energy:

\[
\begin{align*}
\frac{\partial}{\partial t} \rho + \nabla \cdot (\rho \mathbf{v}) &= 0 \\
\frac{\partial}{\partial t} \rho \mathbf{v} + \nabla \cdot (\rho \mathbf{v} \mathbf{v} + p \mathbf{I}) &= \rho \mathbf{g} + \nabla \cdot \tau \\
\frac{\partial}{\partial t} E + \nabla \cdot (E \mathbf{v} + p \mathbf{v}) &= \rho \mathbf{g} \cdot \mathbf{v} + (\nabla \cdot \tau) \cdot \mathbf{v} + k \nabla^2 T
\end{align*}
\]  

(2.5)

with the following equation of state:

\[
E = \rho \epsilon + \frac{1}{2} \rho \mathbf{v} \cdot \mathbf{v}
\]  

(2.6)

\[
\epsilon = \frac{p}{(\gamma - 1) \rho}
\]  

(2.7)

where \( \rho \) - density, \( \mathbf{v} \) - fluid velocity vector, \( p \) - pressure, \( E \) - internal plus kinetic energy, \( T \) - temperature, \( k \) - thermal conduction coefficient, \( \mathbf{I} \) - identity matrix, \( \gamma \) - heat capacity ratio, \( \epsilon \) - internal energy. The stress tensor \( \tau \) is defined as

\[
\tau = \mu (\nabla \mathbf{v} + \nabla^T \mathbf{v} - \frac{2}{3} (\nabla \cdot \mathbf{v}) \mathbf{I})
\]  

(2.8)

where \( \mu \) - dynamic viscosity. The model neglects bulk viscosity and the dynamic viscosity coefficient is assumed to vary gently with altitude. Gas kinetics are omitted and chemical composition varies with height based on isentropic expansion factor as:

\[
\gamma = \frac{5[O] + 7([N_2] + [O_2])}{3[O] + 5([N_2] + [O_2])}
\]  

(2.9)

and specific gas constant is specified as:

\[
R_s = R_i \frac{[O] + [N_2] + [O_2]}{16[O] + 28[N_2] + 32[O_2]}
\]  

(2.10)
2.2. PROPOSED MODELING APPROACH

where \( R_i \) - ideal gas constant and \([N_2], [O_2]\) and \([O]\) are individual number densities actively modulated by the dynamics as mass fractions of its major gas (neglecting feedback into the equation of state for these studies in contrast to Piñeyro [2018]).

For mesopause airglow studies, the calculation of photochemistry, as well as modulation of \( \text{OH}(3,1) \) and \( \text{O}(^1S) \), are based on the approaches reported in Snively et al. [2010] using the chemistry of Adler-Golden [1997] and Makhlouf et al. [1995, 1998].

The solution of conservation part is performed with a shock-capturing finite volume scheme based on numerical methods of [Langseth and LeVeque, 2000] and [Bale et al., 2002] in a modified version of Clawpack 4.2 [Clawpack Development Team, 2002; LeVeque, 2002]. An approximate Riemann solver flux-difference method (“f-wave” method) is used [Bale et al., 2002], allowing the incorporation of the gravity term to the solution of the conservation part of eq. 2.5. Viscous dissipation and thermal conductance are solved via a time-split solution with an explicit Euler finite-difference numerical scheme. The code is parallelized with Message Passing Interface (MPI). The HDF5 data model and file formats are used for parallel saving and retrieving of model data inputs and outputs, as well as to enable parallel restarts.

2.2.4 Simulation of ionospheric plasma responses to AGWs

The simulation of ionospheric plasma production and loss, and transport and ionospheric electrodynamics are performed with the use of the Geospace Environment Model of Ion-Neutral Interactions (GEMINI). It is based on the numerical solution of the 5-moment approximation transport equations, assuming negligible stress and
heat flow effects [Schunk and Nagy, 2009; Zettergren and Semeter, 2012]:

\[
\begin{align*}
\frac{\partial}{\partial t} \rho_s + \nabla \cdot (\rho_s \mathbf{v}_s) &= P_s - L_s \rho_s \\
\frac{\partial}{\partial t} \rho_s \mathbf{v}_s + \nabla \cdot (\rho_s \mathbf{v}_s \mathbf{v}_s) &= -\nabla P_s + \rho_s \mathbf{g} + \frac{\rho_s}{m_s} q_s \mathbf{E} + \sum \rho_s \nu_{st} (\mathbf{v}_t - \mathbf{v}_s) \\
\frac{\partial}{\partial t} \rho_s \epsilon_s + \nabla \cdot (\rho_s \epsilon_s \mathbf{v}_s) &= -p_s (\nabla \cdot \mathbf{v}_s) - \nabla \cdot \mathbf{h}_s - \\
&\quad \frac{1}{\gamma_s - 1} \sum \left( \frac{\rho_s}{m_s} \nu_{st} \right) \left[ 2(T_s - T_t) - \frac{2 m_s}{3 k_B} (\mathbf{v}_s - \mathbf{v}_t)^2 \right]
\end{align*}
\]  

(2.11)

where \( \rho \) - density, \( \mathbf{v} \) - drift velocity, \( P \) - chemical production and impact ionization, \( L_s \) - chemical loss, \( p \) - pressure, \( \mathbf{E} \) - electric field, \( \nu \) - collision frequency, \( q \) - species charge, \( \epsilon \) - internal energy, \( \mathbf{h} \) - heat fluxes, \( T \) - temperature, \( k_B \) - Boltzmann constant, \( m \) - mass, \( \mathbf{g} \) - gravity and \( \gamma \) - adiabatic index. Subscripts \( s \) defines ionospheric particles (\( s = O^+, NO^+, N_2^+, O_2^+, N^+, H^+, e \)) and \( t \) additionally includes neutral particles.

The energy equation for electrons also incorporates thermal conduction and thermo-electric effects, as well as inelastic cooling terms and heating by photoelectrons. The momentum conservation equation is solved only along magnetic field lines, resulting in a simplified \( \frac{\rho_s}{m_s} q_s \mathbf{E} \) term. The last term in the momentum equation represent ion-neutral, electron-neutral and ion-ion collisions.

For the time scales of the dynamics considered in this thesis, one may assume that electrons reach steady state momentum balance much faster than ions, making steady-state momentum balance in the parallel direction a sufficient assumption. Assuming that plasma in the ionosphere is in a quasi-neutral state, the density and momentum equations for electrons are solved as:

\[
n_e = \sum_{s \neq e} n_s
\]

(2.12)
2.2. PROPOSED MODELING APPROACH

\[ \sum_{s \neq e} v_s n_s q_s - v_e n_e q_e = J \]  \hspace{1cm} (2.13)

where \( n \) - mass density and \( J \) - electric current.

The model is based on the electrostatic assumption, which is justified as long as electromagnetic wave dynamics are negligible:

\[
\begin{align*}
\nabla \cdot J &= 0 \\
\nabla \times E &= 0 \\
\frac{\partial}{\partial t} D &= 0
\end{align*}
\]  \hspace{1cm} (2.14)

The calculation of a response electric field and currents is based on a steady-state momentum balance of ions drift in a perpendicular direction and electrons drift in a parallel direction. The convection term is omitted, as it produces negligible effects for subsonic motion [Schunk and Nagy, 2009]. After the calculation of the conserved quantities, perpendicular and parallel currents are described by:

\[
\begin{align*}
J_\perp &= \sigma_\perp \cdot E_\perp - \sum_s \mu_{s\perp} \cdot \nabla_\perp p_s + \sum_s n_s m_s \nu_s \mu_{s\perp} \cdot v_{n\perp} \\
J_\parallel &= \sigma_0 E_\parallel - \mu_e 0 \nabla_\parallel p_e
\end{align*}
\]  \hspace{1cm} (2.15)

where \( \sigma \) - electrical conductivity and \( \mu_s \) and \( \mu_e \) - ion and electron mobilities.

The model includes only electrons as parallel electric current drivers, because their mobility along magnetic field lines is of several orders higher than of ions, whereas only ions generate perpendicular currents. This helps to consider non-steady-state ion drift, but at the same time, obtain an expression for parallel currents (along with parallel electric field) based on the current continuity equation that encodes a steady-state momentum balance. The electric field calculation is performed in two steps: for
ambipolar field, and response electric fields. The calculation of the ambipolar electric field is based on a parallel pressure gradient, whereas the perpendicular pressure gradient is neglected:

\[ E_{a,\parallel} = \frac{1}{n_e q_e} \nabla \parallel p_e \]  

(2.16)

The response electric field is derived through the electrostatic potential:

\[ \nabla \perp \cdot (\sigma_\perp \cdot \nabla \perp \Phi) + \nabla \parallel \cdot (\sigma_0 \nabla \parallel \Phi) = \nabla \perp \cdot \sum_s n_s m_s \nu_s \mu_s \perp \cdot \mathbf{v}_s \perp \]  

(2.17)

The conservation parts of eq. 2.11 are solved using explicit finite volume method. Diffusion, stiff and non-stiff source and electric potential equations are solved sequentially in a time-split manner using implicit and explicit finite difference schemes.

**AGW and ionospheric plasma response simulation result validation**

The main source of data used for the validation of simulation results are TEC measurements. GNSS data are freely available in Receiver Independent Exchange Format (RINEX) formats from regional and global GNSS networks operators such as National Oceanic and Atmospheric Administration (NOAA), Scripps Orbit and Permanent Array Center (SOPAC), UNAVCO, International GNSS Service (IGS), Crustal Dynamics Data Information System (CDDIS) etc. Slant TEC (sTEC) and Vertical TEC (vTEC), as well as accompanying parameters, such as IPP positions, are calculated using Jet Propulsion Laboratory software.

To the best of our knowledge, here we present the first simulations of mesopause airglow fluctuations driven by nonlinear AW shocks generated at the near-field region of an earthquake and TAGWs. Thus, no equivalent airglow data for any such event
have been reported to date, and we are not aware of any datasets that would have captured the phenomena predicted here for prior earthquakes. However, we believe that our results will drive additional interest to establish independent validation and potentially conduct observations in the future.

### 2.2.5 Coupling of numerical models and data flow

The summary of models used and the data flow between them is shown schematically on Figure 2.1. The first, Centroid Moment Tensor (CMT) solutions, derived from finite-fault models, are used as an input to the forward seismic wave propagation simulations, performed with the SPECFEM3D software. For tsunami simulations, vertical displacements from SPECFEM3D model are used as an input to the GeoClaw model. Then, vertical velocities from SPECFEM3D and/or GeoClaw model are used in a vertical momentum bottom boundary (that represents ground-level) condition of MAGIC model.

In MAGIC, as the CFL condition is set constant through the whole time of simulation, the bottom boundary condition is linearly interpolated in time during the simulation. Densities of O, N$_2$, O$_2$, fluid velocities and temperature perturbation data are sent from MAGIC to GEMINI time-dependently to simulate ionospheric responses based on chemistry, collisions, drag and production of electric fields by dynamo effects.

Although AGWs that reach surface, having been reflected or excited in the atmosphere, can drive seismic and ocean waves [Fan et al., 2019; Zhang et al., 2008] (whereas ocean waves themselves can be a source of seismic noise [Ardhuin et al., 2011]) their amplitudes are small in comparison with the signals driven by large
2.2. PROPOSED MODELING APPROACH

![Diagram of models and data flow]

**Figure 2.1:** The scheme of models utilized in numerical simulations reported in the thesis. The arrows indicate data flow directions between models.

Earthquakes or tsunamis. In our studies we assume these processes to produce negligible effects, especially for the investigation of the dynamics in the upper atmosphere. Thus, the coupling between SPECFEM3D or GeoClaw and MAGIC is only in one way, as shown on Figure 2.1. In addition, we neglect the dissipative effects of drag by the ionospheric plasma on AGWs. Even at ionospheric F-layer altitudes, where observable TEC perturbations result from AGW propagation, the densities of neutral particles are several orders higher than plasma density, and thus plasma produces fairly small effect on AGWs [Zettergren and Snively, 2015].
Chapter 3

Ionospheric responses to AGWs driven by the 2015 Nepal Mw7.8 Earthquake

We have chosen the 2015 Mw7.8 Gorkha earthquake in Nepal, which has been broadly investigated, as a demonstrative modeling case of coupling processes, the results of which are presented in this Chapter. Simulations reveal amplitudes, periods and velocities of near-field CIDs to compare with observations. We demonstrate the importance of incorporating near-field seismic wave dynamics for reproducing the observable spatial ionospheric non-uniformity of CIDs, as well testing the possibility to describe faulting mechanism of the earthquake based on ionospheric TEC data. Simulation results of far-field CIDs point to AWs driven by long period seismic waves as their sources. However, we find marked inconsistency of simulated far-field CIDs
3.1. EARTHQUAKE CHARACTERISTICS AND OBSERVED COSEISMIC IONOSPHERIC DISTURBANCES

with RWs-related ionospheric disturbances, reported in some observational reports based on TEC data and we discuss these differences in detail.

The Chapter is structured by first providing information about the earthquake and generated surface displacements (Section 3.1). Then, we review CID observations in literature, in addition to our own analyses of CIDs based on available TEC data for better understanding of observed phenomena (Section 3.1). Section 3.2 describes the modeling approach used for simulations of AGWs and CIDs, including information about the models, their configurations and assumptions. Then, we present modeling results, first of the seismic waves propagation simulation (Section 3.3), followed by the results of atmosphere-ionosphere simulations and analyses of AGWs and CIDs at near- and far-field (Sections 3.4 and 3.5) from the epicenter. We discuss results and compare with observations. Section 3.6 contains the Discussion and Conclusions.

The results shown in this chapter were published in [Inchin et al., 2020a].

3.1 Earthquake Characteristics and Observed Co-seismic Ionospheric Disturbances

The Nepal Gorkha earthquake occurred on 25 April 2015 at 11:56:58 NST (06:11:58 UT) and was one of the most devastating in the last hundred years. It occurred at the Main Himalayan Thrust [Bilham et al., 2001], triggering around 4000 landslides and more than 3000 aftershocks within 45 days after the event [Adhikari et al., 2015; Kargel et al., 2016]. According to Global Centroid Moment Tensor Catalog (GCMT) solution [Ekström et al., 2012], the earthquake generated a scalar moment of $8.39 \times 10^{27}$ dyne-cm (7.9Mw); thrust (reverse-fault) motion on the fault [Zhang et al., 2016] with
3.1. **EARTHQUAKE CHARACTERISTICS AND OBSERVED COSEISMIC IONOSPHERIC DISTURBANCES**

297°/6°/96° Strike/Dip/Slip fault orientation parameters and epicenter at 27.91°N, 85.33°E and ~12 km depth (Figure 3.1). Most finite-fault models showed in common that the rupture nucleated near the hypocenter and propagated to the southeast for 140-160 km with 50-60 km cross-strike extent and a velocity of 2.5-3.2 km/s [Fan and Shearer, 2015; Wei et al., 2018; Yue et al., 2016]. It did not reach the surface and the earthquake ruptured only a deep part of the seismogenic zone [Kobayashi et al., 2015; Wang and Fialko, 2015]. Reported slip distributions vary markedly with maximum slips from 3.1 m to 6.5 m [Zhang et al., 2016].

![Line-of-sight deformation map from the ALOS-2 satellite](image)

**Figure 3.1:** Line-of-sight vertical surface deformation after the Nepal 2015 Mw7.8 Gorkha earthquake adapted from Lindsey et al. [2015]. GCMT solution position, epicenter used in Yue et al. [2016] model and the high-rate data GNSS stations positions used for seismic dynamic and CIDs analysis at near-field region are presented. The beach ball is based on the GCMT moment tensor. The moment rate function is adopted from Yue et al. [2016].
3.1. EARTHQUAKE CHARACTERISTICS AND OBSERVED COSEISMIC IONOSPHERIC DISTURBANCES

Registered vertical surface displacements with a trough-to-peak amplitude up to 1.6 m [Lindsey et al., 2015] serve as a useful indicator that this earthquake may generate observable CIDs [Astafyeva et al., 2013]. Vertical velocities found from strong motion accelerometers data and GNSS high-rate measurements were reported up to 64 cm/s [Takai et al., 2016]. The region of uplift was larger and the amplitudes of uplift displacement were higher relative to subsidence zone [Kobayashi et al., 2015]. According to low-frequency backprojection, 3 main stages of the rupture process can be defined: downdip rupture at the nucleation area for the first 20 s, then updip rupture which released most of the radiated energy from 20 to 40 s, and, finally, terminating stage with updip rupture [Fan and Shearer, 2015; Qin and Yao, 2017]. This is fairly comparable with spatial distribution of vertical surface displacements based on ALOS-2 satellite data shown on Figure 3.1 [Lindsey et al., 2015].

Several comprehensive reports have been dedicated to the investigation of CIDs driven by this earthquake based on TEC data. One of the first reports, by Reddy and Seemala [2015], reported CIDs of maximum $\sim1.2$ TECu ($1\text{TECu} = 10^{16} \text{el m}^{-2}$) propagating with 1.18 km/s (called “slow”) and $\sim2.4$ km/s (“fast”) apparent phase velocities. The “slow” CIDs of $\sim4$ mHz were connected with the propagation of AGWs triggered by near-field displacements, whereas the “fast” ones of $\sim2.7$ mHz - with RW AWs. Reddy and Seemala compared detailed maps of RW fundamental group velocities from Acton et al. [2010] with apparent phase velocities of observable RW CIDs and pointed to their consistency in the Indo-Eurasian region. It should be noted that the map shown by Reddy and Seemala [2015] represents group velocities for 10 s period (fundamental mode) RWs. RW group velocity maps for 20–70 s fundamental modes were also provided by Acton et al. [2010] and may serve as a fairly good overview of
3.1. EARTHQUAKE CHARACTERISTICS AND OBSERVED COSEISMIC IONOSPHERIC DISTURBANCES

RW group velocities in this region. From these maps, due to the highly dispersive nature of RWs, their group velocities at Indo-Eurasian region vary between \( \sim 1.8-4.2 \) km/s (generally, longer period RWs exhibit faster group/phase velocities). Reddy and Seemala [2015] also reported the separation of near-field CIDs and RW CIDs based on selected GNSS stations at Andaman arc region and nearby. We find that Ionospheric Pierce Point (IPP) positions for those stations (DGRP, MBDR, HAVE, PBRI, PORT, CARN and CUSV) with GPS satellites in the field of view after the earthquake (GPS PRN 03, 16, 23, 26 and 27) are, on average, at a distance of \( \sim 1400-2300 \) km from the epicenter (see Figure 3.2). Reddy and Seemala [2015] reported the separation of near-field CIDs and RW CIDs at distance of 1400–2300 km from the epicenter. Though this distance varies depending on fault extension, mentioned by Reddy and Seemala [2015] distances seem to be much farther than reported previously, for example 400-600 km reported by Astafyeva et al. [2009]. Our modeling results suggest that these distances can be \( \sim 250 \) km.

Tulasi Ram et al. [2017] reported TEC perturbations of 1.7 TECu (peak-to-peak), being stronger to east and south from epicenter. They also reported far-field CIDs, observed to the south from the epicenter, propagating with phase velocities \( \sim 1.73-2.39 \) km/s, but pointed to the inconsistency of the propagation velocities of these CIDs with RW group velocities calculated from seismometers data, which were reported as 3.4-3.7 km/s. It was concluded that there were no robust evidences of RW CIDs signatures in the available TEC data set.

Catherine et al. [2017] found near-epicentral CIDs propagating to the east with 0.98 km/s velocity, as well as \( \sim 0.65 \) km/s to the west from the epicenter with maximum
amplitudes of 1.5 TECu to east. Comparatively small velocities to the west were proposed to be a result of neutral winds impact and strong nonlinear effects. They reported far-field CIDs propagating with velocities of $\sim 2.6$ km/s and linked them with RWs. Quite sparse set of TEC data were used and far-field CIDs were reported based on one IPP track (Figure 5.a of Catherine et al. [2017]).

Chen et al. [2017], Zhou et al. [2017] and Liu et al. [2017] also analyzed CIDs based on IGS and Crustal Movement Observation Network of China (CMONOC) data. Chen et al. [2017] found CIDs of maximum amplitude of 1.34 TECu propagating with 0.61 and 1.62 km/s and driven by near-field AGWs and RW AWs respectively at near-epicentral region, then RW CIDs with phase velocities of 2.35 km/s at 500-1500 km distances and RW CIDs of 2.74 km/s at distances $> 1500$ km from epicenter. Liu et al. [2017] reported RW CIDs propagating with velocities 2.0-2.3 km/s. It is interesting to
3.1. EARTHQUAKE CHARACTERISTICS AND OBSERVED COSEISMIC
IONOSPHERIC DISTURBANCES

Note that analyzing the same set of data as in two previous reports, Zhou et al. [2017] failed to detect any RW CIDs. Difficulties with obtaining CMONOC GNSS stations data prohibit us from more detailed analysis of RW CIDs in that region.

All reports in common showed lower amplitudes of CIDs to north than to south and connected this with a dominant mobility of ions and electrons along magnetic field lines Otsuka et al. [2006]. Catherine et al. [2017] and Sunil et al. [2017] studied azimuthal asymmetry of near-field CIDs and its relation to east-southeast rupture propagation direction. First appearance of CIDs is shown between 8 and 11 min after the earthquake, which is consistent with time delay of propagation of AGWs to ionospheric heights. Sunil et al. [2017] and Tulasi Ram et al. [2017] investigated coseismic surface deformations and proposed, that crust uplift at focal area led to initial high pressure phases of excited AWs (mostly to the south from strike direction), whereas crust subsidence resulted in the excitation of AWs with initial low pressure phases (mostly to the north from strike direction). Uplift and subsidence zones can be discerned from Figure 3.1. The wide range of characteristics (amplitudes, velocities, frequencies) of reported CIDs and different explanation of observed phenomena, make this case to be of particular interest for modeling. The difficulty of the analysis of CIDs is partially related to sparse coverage of GNSS stations in the considered region.

The earthquake occurred during the local noon. The peak of the northern ionization crest of the equatorial ionization anomaly (EIA) was at the region of Nepal and north of India, resulting in enhancement of electron density up to 72-80 TECu in the proximity of the epicenter (see Figure 3.5). Geomagnetic conditions were relatively quiet with DST index of -15 nT, 126.2 F10.7 solar and ~2 Ap indexes.
3.1. EARTHQUAKE CHARACTERISTICS AND OBSERVED COSEISMIC IONOSPHERIC DISTURBANCES

Figure 3.3: (a) Tracks of IPP positions for several high-rate GNSS stations near the epicenter with GPS satellite PRN16 (except LCK4-PRN27 and NAGP-PRN27 pairs, which are marked separately). Filled circle at the end of every IPP track presents final time tag and white circle along the track - time of CIDs detection. (b) Absolute vTEC time series are shifted one to another (not equally) for better visibility. For PRN16 the tracks are structured from East (top) to West (bottom). Real absolute values scale is valid only for station RMTE and presented for reference.

For the calculation of IPP coordinates, that represent the position of the intersection of satellite–receiver LOS and ionospheric shell layer (which approximates ionosphere as a thin spherical shell), we choose fixed height of ionospheric shell at 350 km.

Our analysis of near-field perturbation using vTEC data reveals first CIDs ~8-9 min after rupture nucleation and maximum amplitudes of ~1.4 TECu (see Figure 3.3). Azimuthal dependence of observable CIDs is presented; stronger CIDs were detected to the east (e.g. KKN4-PRN16, SYBC-PRN16, CHLM-PRN16, NAGP-PRN27, LCK4-PRN27 etc.) from the epicenter and weaker (~0.1-0.2 TECu or non-detectable) to the northwest and west (LCK3-PRN16, GRHI-PRN16, BMCL-PRN16). GPS satel-
3.1. EARTHQUAKE CHARACTERISTICS AND OBSERVED COSEISMIC IONOSPHERIC DISTURBANCES

GPS satellites PRN 3 and 27 are useful for the detection of perturbation to the south, but at distances of \( \sim 300-600 \) km from epicenter with \( \sim 30-50^\circ \) LOS elevation angles. On Figure 3.3, two time series of TEC for pairs LCK4-PRN27 and NAGP-PRN27 are provided to show that sharp enhancements of vTEC were detected to the south. Clear evidence of ionospheric hole formation is presented for TEC observations with IPP positions close to the epicenter, for example for pairs NAST-PRN16, SNDL-PRN16, KKN4-PRN16, SYBC-PRN16. Electron density depleted to \( \sim 0.1-0.3 \) TECU after the passage of AGWs. It is fair to assume that the recovery phase took at least 30 min after the initial decrease in TEC driven by AGWs. The analysis of the recovery duration of electron depletion with TEC is complicated due to the fact that, after sharp decrease in TEC, the diurnal trend of electron density also started to decrease. In addition, the motion of IPP positions away from the focal area excluded the depletion region from observations, as for example for pair DNSG-PRN16.

On Figure 3.4 we present travel-time diagrams for 10 min high-pass filtered vTEC data. The fault strike direction was approximately normal to local meridional direction (297°, GCMT solution), so we group vTEC measurements with IPP positions to the south and north only. For the investigation of near-field CIDs we use measurements with elevation angles higher than \( \sim 50^\circ \) to exclude low-elevation angle data errors and to accurately calculate their propagation velocities. For the analysis of far-field CIDs, especially those in Andaman arc region, where elevation angles of measurements were \( \sim 25-40^\circ \), we present travel-time diagrams for GPS satellites for which IPP positions were moving roughly to the north (PRN 16 and 26) and south (PRN 3 and 23) separately to exclude ambiguities in phase velocity calculations.
Near-field CIDs propagated to north with velocities 0.76-0.96 km/s. Later in time, CIDs are detected propagating with velocities $\sim 0.6$ km/s which may represent superposition of initial CIDs and those driven by reflected and ducted AGWs (see Figure 3.4.a). For IPPs to the south, CIDs propagated with velocities 0.66–1.5 km/s at distances 0-1400 km. Though not shown on plots, the periods of these CIDs are 3-8 min. There is a large gap in TEC measurements and detection of CIDs between $\sim 800–1400$ km from the epicenter, which prohibits detailed analysis at these distances, leading to higher uncertainties in estimation of CID velocities. Further, as it was shown earlier, there is successful detection of ionospheric disturbances (Figure 3.2 and Figure 3.4,b,c). These disturbances are detected at closest distance of $\sim 1400$ km, with higher propagation velocities than those at near-field region.
3.1. EARTHQUAKE CHARACTERISTICS AND OBSERVED COSEISMIC IONOSPHERIC DISTURBANCES

The use of such sparse data on a comparatively small time interval, as well the low sampling rate data of 30 s, might introduce biases for determination of CIDs velocities. The same issue with determination of the velocities of CIDs was shown in Tulasi Ram et al. [2017], where phase velocities were reported as $\sim$1.73-2.39 km/s. Also, it is difficult to address the belonging of some of the registered CIDs to particular phase fronts (Figure 3.4,b,c at near-field). Based on our analyses of data, we determine the velocities of far-field ionospheric disturbances to be $\sim$1.7-2.7 km/s and $\sim$1.2-2.2 km/s for satellites moving to south (panel b) and north (panel c). We assume that the biases may be higher if velocities of IPP positions of vTEC measurements for low-elevation angles (up to $\sim$0.3 km/s) are taken into account. We also find that far-field ionospheric disturbances exhibited comparatively long periods (6-10 min) and had high amplitudes - up to 1.5 TECu peak-to-peak (Figure 3.2,b).

Overall, using available TEC data sets, we find robust signatures of near-field CIDs. The azimuthal dependence of CIDs is presented with almost no detection of CIDs to west and northwest from the epicenter. We find that stations at Andaman arc region and nearby detected some ionospheric disturbances, but their characteristics do not provide clear conclusion that they are driven by RW AWs. Lack of observations does not allow us to unambiguously conclude that RW CIDs were registered at all. We will continue this discussion in Section 3.5, comparing observations with modeling results and reviewing possible causes of differences of mentioned characteristics.
3.2 Model Configuration

Global version of SPECFEM3D model for forward simulations of three-dimensional global seismic wave propagation is used [Komatitsch and Tromp, 2002a]. To simulate realistic near-field seismic wave dynamics a kinematic slip model from [Yue et al., 2016], based on multi-time window approach, is adapted. The summary of model input parameters is presented in Table 3.1. Kinematic slip model parameters are converted to centroid-moment tensor solution formats applicable for the use in the SPECFEM3D-Globe model. Gaussian source-time function is used as discussed in [Komatitsch and Tromp, 2002a]. Vertical surface displacements are calculated for the domain of 4000x720 km along latitude and longitude respectively and GCMT epicenter coordinates as a center of the domain. For Earth’s interior, the three-dimensional anisotropic S362ANI [Kustowski et al., 2008] model with 3D Crust2.0 [Bassin et al., 2000] crustal model are utilized. The time series of displacements in Seismic Analysis Code (SAC) binary format are then differentiated with a standard 2nd order finite-difference scheme and resampled. The number of spectral elements along the surface of one side for every six chunks, which represent the cubed sphere, is set as 336 and. Thus, the simulation is accurate to a shortest period of \( \sim 12.95 \) s (77.2 mHz). The earthquake occurred in a highly varied topographical area that may provide additional effects on the propagation of short period seismic waves. However, considering only comparatively long period seismic wave propagation of the shortest period of 12 sec, the effect of topography is considered to be negligible. In addition, source inversion algorithm for slip model used in Yue et al. [2016] does not include the topography and, thus (as it is inconsistent with the inversion used) its incorporation indeed provides worse agreement with observations.
3.2. MODEL CONFIGURATION

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Hypocenter</td>
<td>28.1473°N, 84.7079°E, 10 km depth</td>
</tr>
<tr>
<td>Seismic moment</td>
<td>$6.40 \times 10^{21}$ Nm (Mw 7.8)</td>
</tr>
<tr>
<td>Fault parametrization</td>
<td>6° Uniform Dip angle</td>
</tr>
<tr>
<td>Subfaults</td>
<td>18x9 along-strike and down-dip, 10x10 km each</td>
</tr>
<tr>
<td>Rupture velocity</td>
<td>3.2 km/s</td>
</tr>
<tr>
<td>Maximum slip</td>
<td>~6 m at ~100 km south-southeast from hypocenter</td>
</tr>
</tbody>
</table>

Several simulations, domain sizes and resolutions of which are presented in Table 3.2, are carried out to investigate different dynamics of AGWs and plasma responses. Simulation (1) is intended to investigate AGW excitation, propagation and dissipation at near-epicentral region and its output is used as a source of neutral perturbations for the ionospheric model GEMINI; Simulation (2), with a numerical domain up to 300 km, is designed to investigate the development of AGW patterns at constant altitude of 250 km where the ion-neutral collisional moment transfer is dominant; Simulation (3), with an extended latitudinal domain, is used for the investigation of AW propagation (and ionospheric perturbations driven by them) from surface displacements driven by seismic waves at far-field. This configuration of simulations is chosen to mitigate computational expenses. After preliminary tests, it is found that resolving waves up to the frequency of 0.08Hz is appropriate for an accurate investigation of the AGW dynamics in upper atmosphere.

For Simulation (1), an wave-absorbing (“sponge”) layer at altitudes 470-500 km is added and the kinematic viscosity and thermal conduction coefficients are set as constant over this altitude range. The assumption to stop increasing the coefficients at high altitudes and to instead damp waves frictionally where the effects of neutrals on plasma perturbations are comparatively small allows saving considerable compu-
3.2. MODEL CONFIGURATION

<table>
<thead>
<tr>
<th>Simulation</th>
<th>Domain (lat., lon., alt.)</th>
<th>Investigation of</th>
</tr>
</thead>
<tbody>
<tr>
<td>(1)</td>
<td>990×600×500 km, 1×2×0.25 km</td>
<td>Near-field AGWs and CIDs</td>
</tr>
<tr>
<td>(2)</td>
<td>600×600×300 km, 1×1×0.25 km</td>
<td>Spatial dynamics of AGWs</td>
</tr>
<tr>
<td>(3)</td>
<td>4000×500 km, 0.2 km uniform</td>
<td>Seismic wave AGWs and CIDs</td>
</tr>
</tbody>
</table>

Simulation time but still obtain robust results. Simulation (2), which is dedicated to study the spatial dependence of AGW patterns from surface dynamics, does not require coupling with the ionospheric model and can be run up to comparatively lower altitudes than Simulation (1) without any additional restrictions. Simulation (3) requires an axisymmetric source in order to eliminate discontinuities at the center ($r = 0$). However, being still interested in simulating AWs propagating at far-fields within the axisymmetric model to preserve geometrical spreading, surface perturbations near the epicenter (+/-300 km to north and south) are smoothly suppressed. By doing so, the near-field perturbations are eliminated, which are investigated through the Simulations (1) and (2), that allowed simulating AWs propagation at far field in 2D up to the altitude of 500 km in polar coordinates with high spatial resolution and without additional restrictions. As the waves never intersect, the requirement for even symmetry is relaxed.

At the top and sides of the MAGIC numerical domain, boundary conditions that allow waves to pass freely through boundaries without reflection are set. The densities of neutrals at 500 km (top MAGIC boundary) are negligible and thus there are no discontinuities in parameters between the models at high altitudes. As a bottom boundary condition, vertical surface velocities calculated in SPECFEM3D-Globe are applied until surface velocities become negligible. After RWs leave the domain, an impermeability condition is set at the bottom boundary to resolve AGW reflec-
3.2. MODEL CONFIGURATION

tions. For the analysis of CIDs, meridional slice at 85.35°E, which is in proximity of the GCMT solution epicenter (see Figure 3.1) is used. Numerical boundaries for the calculation of electric potential are set as “grounded” (zero for all points at the boundaries). The 1D vertical profiles of the ambient neutral atmospheric state for MAGIC and GEMINI models are calculated using the empirical model NRLMSISE00 [Picone et al., 2002]. 1D vertical profiles of both meridional and zonal winds are determined using HWM-14 model [Drob et al., 2015] and incorporated into MAGIC simulations and then encapsulated into GEMINI simulation as a part of meridional fluid velocity. These profiles remain stationary, assuming that their change is negligible over the considered time frame and chosen spatial domain.

The earthquake happened at local noon and in the proximity of the EIA crest region that is important to resolve in the simulation. The EIA markedly affects the resulting electron density perturbations (through the simulation with and without EIA the difference of 0.6 TECu peak-to-peak in maximum perturbations is found). We implement in GEMINI the same approach to model the EIA as described in Huba et al. [2000]. A balanced ionospheric state is achieved by running an additional GEMINI simulation for 24 hours prior to the event to ensure a set of state parameters that are a good representation of the ionosphere. Figure 3.5 shows the results of electron density for GEMINI, as well as comparison of absolute vTEC values for latitudes of earthquake epicenter (27.91°N) calculated in GEMINI and obtained from JPL IONEX map for 12:00:00LT. Fairly good agreement between vertically integrated electron densities from the simulation and vTEC data at near-epicentral region are obtained, but differences can be seen at magnetic equator (Figure 3.5,c). Though neutral atmosphere simulations domains do not reach the magnetic equator region, it
should be noted that IONEX maps represent interpolated vTEC data from a sparse network of stations with 2.5° by 5° resolution (in latitude and longitude directions, respectively) grid, as well sparse in time domain, and cannot serve as a robust reference for comparison. Vertically integrated electron density profiles from the simulation with absolute vTEC values, calculated for satellite-station pairs at near-field region (~72-80 TECu), have fairly good agreement.

Figure 3.5: (a) Global vTEC map (IONEX) at time of earthquake (12:00:00LT). The ionospheric region above Indian peninsula and Nepal is affected by EIA. (b) The background electron densities map from GEMINI meridional slice (shown on panel a as black line) used in Simulation (1). (c) Comparison of absolute vTEC between IONEX and GEMINI.

3.3 Seismic Waves Dynamic Simulation Results

To validate simulation results synthetic seismograms are compared with processed and reviewed data for strong motion accelerometer KATNP maintained by United States Geological Survey, high-rate continuous GPS (cGPS) data, which are applicable for studies of near-field faulting dynamics [Crowell et al., 2012] and far-field
3.3. **SEISMIC WAVES DYNAMIC SIMULATION RESULTS**

GSN broadband seismometers data from IRIS DMC archives. Data from cGPS stations (5 Hz), 4 of which are located at rupture zone (CHLM, KKN4, NAST, SNDL), are obtained through the UNAVCO service. RINEX data were processed using Precise Point Positioning Kinematic service provided by Natural RESources Canada to calculate surface displacements.

Stations KATNP and NAST are placed in the Kathmandu basin, which is located mostly on sediment between igneous rocks leading the whole basin to serve as a resonator for seismic waves [Fujii and Sakai, 2002] with observable 4-5 s period shaking [Galetzka et al., 2015]; this is problematic to resolve in our simulations. However, we do not simulate surface dynamics up to such high frequencies, because their contribution to CIDs is fairly negligible. For comparison of observation and synthetic data we include stations in Kathmandu basin too.

Reasonable agreement between simulation and observations is found (see Figure 3.6). Comparison of RW phase velocities from simulation with regional maps of RW group velocities from Acton et al. [2010] and Ritzwoller and Levshin [1998] reveals their consistency for Indo-Eurasian region and Indian peninsula. Results also show that the simulation preserves the direction of rupture propagation to east-southeast from hypocenter. The main energy lobe of surface waves is directed to southeast, but the directivity is dominant for periods < 80 s. For periods > 100 s, the directivity effect is fairly small. This is in the agreement with Duputel et al. [2016]’s comparison of observed and predicted seismograms based on CMT solution. For long-period RWs, the directivity plays small role with only slightly higher amplitudes to southeast. Thus, the comparison of simulation results with observations shows that such
3.4 Atmosphere-ionosphere simulation results

The coupling of surface displacements with the atmosphere results in the generation of AGWs, which propagate with velocities and amplitudes that depend on background surface dynamic can be used as a realistic source of AGWs generated by the earthquake. Some inaccuracies can be connected with choices of different Earth models for source inversion in the kinematic slip model and that used in SPECFEM3D-Globe and topography, as well as source representation.
3.4. ATMOSPHERE-IONOSPHERE SIMULATION RESULTS

state conditions [Bergmann, 1946; Mikhailov and Martynov, 2017]. They can reach considerably high amplitudes in thermosphere; nonlinear effects of AW propagation in the near-epicentral region leads to the formation of shock waves. Figure 3.7,a represents the energy spectral density of the scaled vertical velocities on an altitude-frequency diagram for the location 10 km to the north from the epicenter (GCMT) from Simulation (1). Scaled instantaneous vertical fluid velocity profiles for 3 time epochs are shown as well on Figure 3.7,b. The generation of AW high frequency spectrum between the ground and mesopause is a consequence of the formation of shock waves [Sabatini et al., 2016]. At higher altitudes, thermoviscous dissipation leads to attenuation of high frequency AWs, along with shifting the AW spectrum to low frequencies due to the lengthening of nonlinear AWs. High frequency content of AWs is markedly attenuated by these damping mechanisms at thermospheric heights corresponding to the ionospheric F layer and higher.

![Figure 3.7](image)

**Figure 3.7:** (a) Energy spectral density diagram of frequency-altitude profile for the point 10 km to the north from GCMT epicenter. (b) Scaled vertical fluid velocities.

A series of snapshots of scaled pressure perturbations $p'/\sqrt{\rho/\rho(0)}$ (where $p'$ - pressure perturbations, $\rho$ - density and 0 represents ground) from Simulation (1) of AGW
propagation is presented on Figure 3.8. Positive (to the south from the epicenter) and negative (to the north) initial phases of leading AWs are clearly seen at altitudes lower than $\sim 80$ km (Figure 3.8,a), because the strike direction was almost perpendicular to the local meridional direction, and plots represent meridional slices along the longitude of GCMT epicenter. This is approximately consistent with the map of surface displacements: uplift to the south and subsidence to the north from the strike direction. Refraction of AWs, both at stratospheric heights ($\sim 50$ km) and at mesopause ($\sim 90$-$120$ km), occurs. Increasing of amplitudes and steepening of AW fronts can be seen from the lower thermosphere, up to $\sim 270$ km.

Starting from the ionospheric F-layer ($\sim 200$ km and higher), leading AWs exhibit a uniform compressional phase front to the south and the north from the epicenter. Although the leading front is not fully meridionally symmetric, and AWs to north are lagged comparatively to AWs to the south (see panels b and c of Figure 3.8), this difference is almost not detectable. This important result shows that the nonlinearity of strong near-epicentral shock AWs, their confluence and interference at near-field region, may change leading phase fronts during AW propagation to the thermosphere and even obscure phase information [Whitham, 1974]. Also, strong AWs with initial compression phase, excited from uplift zone to the south, markedly obscure and overtake AWs excited from subsidence zone to the north. This raises concerns about the possibility to describe faulting mechanism and surface deformations based on ionospheric observations (particularly TEC) where waves become strongly nonlinear. Whereas for altitudes $< 100$ km the uplift/subsidence pattern of excited AWs is still preserved, at thermospheric altitudes the nonlinearity of the propagating AWs, and their interference, may completely change the initial phase structure of AW fronts,
even forming uniform-looking shock fronts. This might be the reason why, for some cases, axisymmetric representation of sources for simulations of near-field disturbances provide quite good agreement with observations [Zettergren et al., 2017]. However, as it is shown here, the dynamics of near-field AWs and the formation of leading phase fronts of AWs, may be much more complicated and lead to false conclusions about AWs propagation from surface to thermosphere. At the same time, simulation shows that AWs driven by seismic waves at farther distances from epicenter propagate in a linear regime and their initial phases are preserved even at ionospheric altitudes.

Panel e of Figure 3.8 represents time-latitude diagram of simulated neutral gas vertical fluid velocities at altitude 250 km along the longitude of the GCMT epicenter. Again, almost fully meridionally-uniform AW fronts can be clearly seen, followed by strong AW rarefaction phases. Three distinct groups of AWs, propagating with velocities \( \sim 8 \text{ km/s} \), \( \sim 4 \text{ km/s} \) and \( \sim 0.9 \text{ km/s} \), are clearly seen. The fastest AWs are generated by surface displacements, that are consistent with phase velocities of P waves, as shown on Figure 3.6. They are of quite small amplitudes (\( \sim 1-3 \text{ m/s peak-to-peak} \)) and cannot be feasibly detected with TEC measurements, particularly due to the need of even lower LOS angle of measurements for their detection than for RW CIDs (see Section ??). They may be observed with, for example, Doppler sounding systems [Chum et al., 2012]. Second source of AWs are RWs and the correspondence between the phase velocities of comparatively long-period RWs at the surface and of these AWs is preserved. Finally, the slowest perturbations of the highest amplitudes are driven by near-field AWs, propagating with phase velocity of \( \sim 0.8-9 \text{ km/s} \).
3.4. ATMOSPHERE-IONOSPHERE SIMULATION RESULTS

Figure 3.8: (a-d) Latitude-altitude slices of simulated scaled pressure perturbations for meridional slice along the GCMT epicenter based on Simulation (1). The epicenter latitude is indicated with a red star. Min/Max vertical velocities obtained from this simulation range from -280 to 187 m/s. (e) Oversaturated time-latitude diagram of vertical fluid velocities at altitude of 250 km for epicenter meridional slice.
3.4. ATMOSPHERE-IONOSPHERE SIMULATION RESULTS

Based on the analysis of CIDs after the Kurile 1994 Mw8.3 earthquake, Astafyeva et al. [2009] demonstrated, that the perturbations driven by near-field AGWs and by RW AWs can be discerned at $\sim 600–700$ km from epicenter. This observation is supported by the fact that near-field seismic wave dynamics are usually very complex, especially for strong earthquakes where faulting process can develop on a region of hundreds of kilometers. It should be noted, that AW wavefronts and LOS geometry of TEC measurements may introduce additional biases here. Our results suggest, that the closest distance where RW AWs can be confidently discerned from near-field AGWs for the Nepal earthquake is $\sim 250–400$ km. This result differs markedly from that reported in Reddy and Seemala [2015] where separation of near-field and far-field CIDs was proposed at distance of $\sim 1400-1500$ km.

Figure 3.9 depicts results of the GEMINI simulation of ionospheric plasma responses to the propagation of AGWs for 3 time epochs. Ion field-aligned drift velocity, plotted in these panels, is a density weighted value of six ion species velocities. Electron density and ion temperature perturbations are shown as the percentage change from the unperturbed state. Steep and nonlinear AWs at altitudes lower than 300-350 km (Figure 3.8), drive plasma responses by transport of charged particles confined predominantly along magnetic field lines toward the magnetic equator. As it was shown in Zettergren and Snively [2015], the plasma motion for altitudes lower than $\sim 300-400$ km is driven mostly by directly-forced perturbations from the neutral gas. At altitudes higher than $\sim 400$ km, plasma velocities become larger in magnitude, due to coupling with nearly freely propagating ion-acoustic waves at higher phase speeds. The same pattern can be seen in our simulation. Latitudinal asymmetry of electron density perturbations is clearly seen on panels b, e and h of Figure 3.9.
Ions in E and F ionospheric layers, due to the heat exchange with neutrals through the collisional interaction and plasma compression and rarefaction driven by AW propagation, exhibit similar pattern of temperature perturbations as for neutral gas shown on Figure 3.8b,c. At higher altitudes, there are freely propagating ion-acoustic waves along magnetic field lines and almost no perturbations driven by AWs.

We calculate vTEC perturbations from the Simulation (1) for a meridional slice along the epicenter (Figure 3.10,a,b). To do so, for every point on a grid, we integrate electron density along the radial direction, that corresponds to 90° elevation angle LOS. The time-latitude diagram shows important nuances of vTEC measurements for the detection of CIDs in the vicinity of the epicenter, where the plasma motion relative to LOS of vTEC is almost perpendicular (zenith). Right above the epicenter (18.1° magnetic latitude), only small electron density perturbations are represented (though perturbations can be clearly seen from time-latitude diagram of vertical fluid velocities on panel e of Figure 3.8). The perturbations can be clearly distinguished in vTEC within the fidelity of our simulation at distances ~50-60 km from the epicenter. This results from several factors: 1) the directivity of rupture propagation that drove the strongest AGWs to southeast, 2) dominant plasma mobility along magnetic field lines, and 3) direction of LOS used (vertical integration of electron density for each point of the numerical domain). Panel a of Figure 3.10 shows that initial enhancement of vTEC to the north from the epicenter is barely seen, whereas, to the south, vTEC enhancement is readily detectable. This is related to meridional plasma response asymmetry, discussed in previous paragraph. Following downward motion of plasma produces quite intense perturbation to north and to south (due to much stronger perturbations: 187 m/s at the leading compression AW phase, ver-
3.4. ATMOSPHERE-IONOSPHERE SIMULATION RESULTS

Figure 3.9: Ions field-aligned (positive downward) drift velocities (first column), electron density perturbations (second column) and ions temperature perturbations (third column) for 3 time epochs from Simulation (1). Plasma motion at altitudes \(<\sim300\text{km}\) is driven by interaction with neutral species, and at higher altitudes freely propagating ion-acoustic waves can be seen.
3.4. ATMOSPHERE-IONOSPHERE SIMULATION RESULTS

sus -281 m/s of the following rarefaction AW phase of vertical fluid velocities). The propagation velocities of CIDs, driven by near-field acoustic and gravity waves, are calculated as 0.92–0.95 km/s and 0.21 km/s, with periods of 200–225 s and 514–720 s respectively. Observed phase velocities of 0.92–0.95 km/s of CIDs are fairly consistent with simulated near-field AGW phase velocities, shown on time-latitude diagram on Figure 3.8,e. Maximum perturbations from presented vTEC latitude-time diagrams are found to be ~2 TECu.

![Image](image1.png)

**Figure 3.10:** (a) Time-Latitude diagram of simulated vTEC perturbations; (b) and (c) time series of vTEC at chosen latitudes and their wavelet power spectra. (d) Fourier transform for plot (c). (e) Oversaturated time-latitude diagram of vTEC. (f-h) Time series of vTEC at chosen latitudes shown on panel e with black dashed lines.
3.4. ATMOSPHERE-IONOSPHERE SIMULATION RESULTS

Electron density depletion can be seen on panel a of Figure 3.10, as a persistent decrease of electron density in the vicinity of the epicenter, which can indicate a nonlinear (semi-permanent) response to the propagation of AWs due to transport of plasma that may also lead to a higher rate of recombination [Zettersgren et al., 2017]. Its magnitude is 0.15–0.17 TECu and comparable with magnitudes presented in past observational reports [Masashi et al., 2015; Tulasi Ram et al., 2017] and with results of our analysis in Section 3.1. The magnitude is small, in comparison, for example, with the 2011 M9.1 Tohoku-Oki earthquake, where AGWs drove a TEC depletion of 4–5 TECu [Saito et al., 2011; Zettergren and Snively, 2019; Zettergren et al., 2017]. We assume that this is due to the much smaller amplitudes of AGWs for the Nepal earthquake. Tulasi Ram et al. [2017] reported the duration of the ionospheric hole’s recovery phase as approximately 4–5 min, but pointed out the difficulty for the estimation, due to the superposition of different CIDs. This difficulty might also be connected with the motion of IPP position of TEC measurements within the spatially constrained region of the ionospheric hole (in simulation it is of only ~200 km latitudinal size), as well as with diurnal electron density variations. Simulation results suggest that the ionospheric hole may exist for a long duration, even 1 hour after the earthquake. The acoustic resonances, generated by trapping of AWs between lower thermosphere and tropopause/surface, can also be seen from panels b and c of Figure 3.10 at times ~06:50 UT, indicating the arrival of the second AWs packet “burst” with similar periods as the first one (~220-225 s).

As shown in Figure 3.3, a sharp decrease of electron density was registered in TEC for satellite-station pairs with IPP to the north from the epicenter, that was linked with AWs with initial rarefaction phases excited at crust subsidence [Sunil et al.,
3.4. ATMOSPHERE-IONOSPHERE SIMULATION RESULTS

2017]. Simulated vTEC time-latitude diagram, two vTEC time series, slightly to the north from epicenter, and one vTEC time series far to the south, are shown on panels e-h of Figure 3.10, respectively. The replication of exact TEC measurements without full 3D ionosphere simulation is not possible. However, the provided time series are comparable with TEC observations for pairs DNSG-PRN16 and LCK4-PRN27, with IPP positions to north and south from epicenter, respectively. Panels f-h of Figure 3.10 show that in the proximity of the epicenter to north, only the depletion phase of vTEC may be registered initially, that could lead to false interpretation of these dynamics to be driven by AWs with initial rarefaction phases, excited at subsidence zone. Observed vTEC enhancement to the south from the epicenter can be seen in our simulation results shown on Figure 3.10,h.

This simple demonstration, based on simulated vTEC, shows that initial decrease in TEC perturbations may not necessarily be connected with AWs excited with initial rarefaction phase from subsidence of the crust. It should be noted that usually low sampling rate and LOS of TEC measurement geometry and direction of the CIDs’ fronts might also lead to misinterpretation of the sequence of phase fronts for observed CIDs driven by different AWs. In addition, the spatial and temporal complexity of the source can complicate data analyses. Meng et al. [2018] addressed this problem based on simulations of the 2011 Mw9.1 Tohoku-Oki and the 2015 Mw8.3 Illapel earthquakes. They found that an axisymmetric AW source driving an ionospheric model (GITM), without any direction preference, provides quite different latitudinal patterns of TEC perturbations. Zettergren and Snively [2019] investigated this latitudinal asymmetry of driven CIDs with the use of an axisymmetric source of AGWs and concluded, that dominant motion of plasma along magnetic field lines markedly
affects the spatial structure of CIDs. The description of faulting mechanism and surface deformations based on TEC measurements still requires additional investigations particularly through comprehensive modeling studies.

Figure 3.11: Panels (a,b,c) show nonlinear AW dynamics at near-epicentral region at fixed altitude. The dislocation of AWs from earthquake hypocenter to east is clearly seen, where there are almost no AWs propagating to west and northwest. (d,e,f) show AW dynamics for same time epochs and 1/100 magnitude of the source. In this linear regime, phase information of leading AWs is preserved.

The last addressed aspect is the longitudinal dependence of AW and CID dynamics from surface displacement dynamics. As it is shown in Section 3.1, during the earthquake TEC perturbations were observed mostly to the south, southeast and northeast from the epicenter, whereas to the west and northwest there were almost
3.4. ATMOSPHERE-IONOSPHERE SIMULATION RESULTS

no observable CIDs. The same spatial pattern is obtained through Simulation (2), see panels a-c of Figure 3.11. The radiated AWs shift approximately from the epicenter position to the east, southeast and northeast. The most intense AWs (Figure 3.11,c) are excited to the southeast. This is consistent with the fact that maximum energy release happen at \(\sim 100\) km from the epicenter position, while the rupture propagates to southeast, close to the longitude of GCMT epicenter. At the same time, there are almost no excited AGWs to west, and especially not to northwest. This points to the advantage of the incorporation of finite-fault models for accurate representation of excited AWs at near-field region. Point or axisymmetric sources, frozen at one location would not be able to reproduce this important feature; in our case, the source of AWs is moving with propagation of rupture to southeast at supersonic speed of 3.2 km/s, and closely replicates surface displacement dynamics in time and space. The dependence of AW initial phases from displacements (uplift or subsidence) is not presented here. However, in order to demonstrate the importance of nonlinear effects on AW propagation, we decrease the amplitudes of surface velocities by a factor 100 and rerun the Simulation (2) with other parameters left unchanged. The results are presented on panels d-f of Figure 3.11. In this case, in linear regime of AW propagation, the preservation of phase information of leading AWs can be clearly seen: rarefaction initial phase of leading AWs to the north and compression initial phase of leading AWs to the south. This again demonstrates the importance of AW nonlinear effects for strong waves excited at the near-epicentral region, and points to the difficulty of the description of surface deformations based on TEC.
3.5 Atmosphere-ionosphere responses at far-field

Transient surface displacements from seismic wave propagation induce AWs in the atmosphere. These AWs may also drive observable ionospheric perturbations even at distances far from the epicenter, although usually of small amplitudes [Chum et al., 2012; Ducic et al., 2003]. Based on Simulation (3) results, we present snapshots of instantaneous vertical fluid velocities for altitude-latitude slice in panels a and b of Figure 3.12. Time-latitude diagram of vertical fluid velocities at constant altitude of 300 km is shown in panel c of Figure 3.12. For this simulation we use meridional slice through GCMT epicenter position.

RWs serve as a source of AWs in the atmosphere that propagate upward with sound speed of $\sim 0.34 \text{ km/s}$ at ground level and a wavefront almost parallel ($\sim 14^\circ$ elevation angle) to the surface as it can be seen from Figure 3.12,a. At ionospheric height, due to the marked increase of sound speed with background temperature, the elevation angle of RW AWs wavefronts reaches $\sim 25^\circ$ (Figure 3.12,b). For P waves the elevation angle at ground level is only $\sim 3^\circ$ and reaches $\sim 8^\circ$ in ionosphere. Simulation results suggest that far from the epicenter the seismic wave-driven AWs propagate in a linear regime. Only long period RW AWs propagate to ionospheric heights and reach distances of 2000 km from epicenter. At distances 750 km from the epicenter and further maximum vertical fluid velocities of RW AWs are $\sim 25 \text{ m/s}$, $\sim 0.02 \text{ TECu}$ with vertical integration and $\sim 0.05 \text{ TECu}$ with 30$^\circ$ elevation angle of electron density integration. Comparatively high frequency content of RW AWs is also found in the simulation with $\sim 2.7-3.0 \text{ km/s}$ apparent (imprinted by their source) horizontal phase velocities and reaching lower thermosphere; however, they almost fully dissipate below 200-250
km; their vertical fluid velocities do not exceed ∼1-3 m/s at 250 km altitude. AWs driven by P waves are present propagating with imprinted horizontal phase velocities of ∼8 km/s that are consistent with their phase velocities at the surface.

Lognonné et al. [1998] demonstrated that the effective coupling of solid modes with atmosphere happens when solid mode eigenfrequencies are close to the frequencies of the atmospheric waveguides, and revealed 2 most energetic atmospheric modes of 3.7 mHz and 4.3 mHz. Rolland et al. [2011]’s modeling and observations of RW CIDs for the 2008 Wenchuan Mw7.9 and the 2003 Tokachi-oki Mw8.3 earthquakes support this. The periods of RW CIDs obtained from Simulation (3) of ∼4 mHz support these results. The inclusion of high frequency spectrum of RW AWs to our simulation does not result in stronger and observable CIDs at far-field. These frequencies are quite higher than frequencies of ionospheric perturbations registered at Andaman arc region (∼2.7 mHz, as reported by Reddy and Seemala [2015]), but this could result from Doppler shift effect due to the motion of IPP position [Savastano, 2018]. It is important to note that phase velocities of RWs with frequencies 3.7-4.3 mHz are >3.5 km/s in the considered region [Acton et al., 2010]; they are much faster than ionospheric disturbances which were related to RWs based on TEC observations by Reddy and Seemala [2015] propagating with ∼2.4 km/s.

Technically, the amplitudes of simulated TEC perturbations driven by RW AWs (0.02–0.05 TECu) can be observed, but we fail to find any such CIDs in the available TEC data set. Probably, they are not resolved spatially within the sparse available measurements, or are masked by other ionospheric dynamics, for example variations in electron density driven by EIA. Simulation assumptions, atmospheric and
ionospheric ambient states and winds, source and models setup may also be insufficient for accurate reproduction of RW CIDs. Planned future investigation of seismic waves-atmosphere-ionosphere coupling processes based on idealized simulations with synthetic sources may clarify discrepancies reported here.

![Figure 3.12](image)

**Figure 3.12:** (a,b) Vertical fluid velocities from Simulation (3) for 2 time epochs. (c) Time-latitude diagram of vertical fluid velocities.

One challenge with the investigation of CIDs based on TEC data is associated with the necessity to use low elevation angle LOS measurements for their detection. Taking into account that plausible geometry for the detection of CIDs is when AWs wavefront and TEC LOS between station and satellite are parallel (due to integration nature of TEC) to each other, practically only low elevation angle TEC measurements are useful for the detection of RW CIDs (approx. <30°). However, these measurements suffer from multipath, numerous cycle slips, noise levels of phase data, and other errors, which affect the accuracy of TEC measurements [Grewal et al., 2007; Komjathy, 1997]. In addition, their IPP velocities cannot be considered as negligible. The velocities of IPP positions are markedly higher for low elevation angles and can be up to \( \sim 0.6 \)
3.5. ATMOSPHERE-IONOSPHERE RESPONSES AT FAR-FIELD

km/s for 10° measurements. Also, the higher altitudes for ionospheric shell layer, the higher velocities of IPP positions. We assume that for the analysis of low elevation angle measurements, this effect or appropriate biases should be accounted.

Source of errors for the analysis of RW CIDs (and CIDs in general) using TEC data may be connected with the choice of ionospheric shell height, or effective height, which is usually linked to peak electron density altitude. The effective height choice results in the determination of IPP position (and subsequently apparent CIDs velocities) and absolute vTEC values and may be highly variable at equatorial latitudes where the non-uniformity of plasma distribution driven by EIA and electrojet provides complex small and medium scale electron density gradients. In equatorial region, a single mean average height cannot be considered to be representative for all LOS of TEC measurements and depends on time of day, the location of observation and the vertical ionization distribution at that location [Rama Rao et al., 2006]. In previous observation reports for the Nepal earthquake, the effective height was used as 300-350 km. Chen et al. [2017] calculated it as 328 km based on IRI 2012 model for April 25, 2015. However, as it was shown by Niranjan et al. [2007] (or, for a general case, by Komjathy [1997]), the median models like IRI do not reproduce the spatial distribution of ionization during times when the region is subject to EIA. Rama Rao et al. [2006] discussed the applicability of such altitudes for low elevation angles measurements for the Indian region. They concluded that the IPP altitude of 350 km may be taken as valid for the Indian peninsula only in cases of TEC elevation angles greater than 50°. Based on monthly mean diurnal variations of $h_mF_2$ from ionosondes data they showed that at local noon the peak density height can reach altitudes of ~400-500 km. This suggests that for the Indian region, where the RW CIDs were detected,
the IPP height, due to EIA effects could be higher than 300-350 km. It also should be noted that far-field CIDs are registered at the proximity of geomagnetic equator, which could markedly affect their propagation direction, velocities and amplitudes retrieved from TEC measurements. At the same time, the choice of IPP position as electron density peak to study ionospheric perturbation is not fully justified because the dominant ion-neutral collisional moment transfer starts at much lower altitudes than the effective height and it plays a main role in plasma disturbances. Thus, for the Nepal earthquake, which happened during local noon, the behavior of plasma could be affected by the presence of EIA.

In summary, based on the analysis of simulation results of far-field ionospheric perturbations for the Nepal 2015 earthquake we find that: 1) In marked contrast to those at near-field, AWs driven by seismic waves at far-field propagate mostly in the linear regime; 2) ionospheric CIDs driven by RWs have similar apparent horizontal phase velocities as RWs at the surface for appropriate periods; 3) thermoviscous dissipative mechanisms markedly attenuate high frequency content of RW AWs, and mostly long-period RW AWs (~4 mHz) can reach ionospheric heights; 4) the registration of TEC perturbations from body wave driven transient surface displacements is unlikely; 5) According to our simulation, the amplitudes of TEC perturbations driven by RW CIDs could be of 0.02–0.05 TECu, but they are not found in the observations.

3.6 Conclusions

This chapter contains case study results of ionospheric responses to atmospheric AGWs after the 2015 Nepal Mw7.8 Gorkha earthquake. Our primary conclusions
3.6. CONCLUSIONS

are: 1) the incorporation of near-field seismic waves dynamic based on finite-fault models provides marked improvement for the simulation of realistic AGWs and CIDs at near-field region; the accuracy of finite-fault model plays a crucial role in the simulation of realistic ionospheric response and should be chosen appropriately. 2) TEC data at near-field region can serve as a complementary source of information about surface dynamics, but the recovery of information about faulting mechanisms or surface deformations based on TEC are complicated by the loss of phase information of AWs due to nonlinear effects, as well as magnetic field effects on plasma motion. In addition, it should be noted that such analysis can be complicated by unfavorable geometry of TEC measurements, lack of data, or the complexity of the surface dynamics during the earthquakes. 3) The electron depletion can be observed even after this inland earthquake, in the same manner as for large undersea earthquakes, e.g., as discussed by Zettergren et al. [2017]. 4) The separation of near-field CIDs and those driven by RWs can be detected already at distances of 250-400 km, but, in general, we assume that this distance varies depending on near-field seismic waves dynamic and fault characteristics. 5) The determination of velocities of CIDs (especially RW CIDs) may be hindered by low density of TEC measurements, impacts of EIA, increased IPP velocities for low elevation angle TEC measurements and geometry of AWs wavefronts and TEC LOS. The regime of propagation of RW AWs for the event considered could be fairly linear.
Chapter 4

Constraining finite-fault kinematics of the 2016 M7.8 Kaikoura Earthquake from ionospheric measurements

Here we present the analysis of AW dynamics and corresponding ionospheric plasma disturbances (CID) generated by the 2016 M7.8 Kaikoura earthquake. This earthquake occurred in the Marlborough fault zone in the northern South Island of New Zealand and resulted in more than dozen surface fault ruptures (e.g., Hamling et al. [2017]). Due to the unusual rupture complexity, some first-order aspects of the rupture evolution and progression remain unclear to date, despite the availability of a wide range of geophysical and field data including Interferometric Synthetic-Aperture
4.1. THE 2016 M7.8 KAIKOURA EARTHQUAKE

Radar (InSAR), Global Positioning System (GPS), strong-motion, fault offsets and tidal and wave gauge data. We address some of the limitations by investigating the ability of different finite-fault models to reproduce the observed TEC perturbations using full physics-based models. We compare the simulation results with observed TEC perturbations and examine the validity of our modeling approach and the sensitivity of the resulting TEC signatures to the finite-fault models.

The Chapter is divided into following sections. Section 4.1 provides the overview of the Kaikoura earthquake and detected CIDs with TEC observations. Our modeling approach, assumptions and parameters are described in Section 4.2. Section 4.3 contains the analysis of AW dynamics and corresponding CIDs and the sensitivity of TEC perturbations to various finite-fault models. The Discussion of the results is provided in Section 4.4, and a Summary of the study in Section 4.5. The results in this chapter form the basis of a manuscript to be submitted for review in consideration for publication; they were initially presented by Inchin et al. [2019].

4.1 The 2016 M7.8 Kaikoura earthquake

4.1.1 Earthquake characteristics

The M7.8 Kaikoura earthquake occurred in the area known as the Marlborough fault system, which is characterized by the tectonic transition from the Hikurangi subduction zone in the northeast to the strike-slip dominating Alpine fault zone in the southwest [Van Dissen and Yeats, 1991; Wallace et al., 2012]. The GCMT solution provides the epicenter of 42.74°S/173.05°E, and the origin time \(T_0\) of 00:02:56 LT 14 November 2016 (11:02:56 UT 2016.11.13). The earthquake is considered as one
of the most complex in-land earthquakes ever recorded \citep{Hamling2017}. Up to $\sim$12-20 ruptured fault segments have been identified from geodetic and field observations, some of which were previously unknown or considered inactive \citep{Xu2018}. The overall faulting area extends to $\sim$180 km from the southwest to the northeast with the vertical surface deformation of up to $\sim$10 m locally \citep{Clark2017, Litchfield2018}. The overall rupture process was fairly slow, as it took $\sim$90 s for the rupture to propagate from the hypocenter, located near its southern end, to the northern end in Cook Strait \citep{Cesca2017, Holden2017}.

Finite-fault models developed in previous studies suggest that the earthquake rupture initiated on the Humps fault, propagated through southern faults including the Leaders, Hundalee, and offshore Point Kean faults, and then onto the northern faults including the Jordan, Kekerengu, Papatea, and Needles faults (Figure 4.1,d) \citep{Ando2018, Bai2017, Cesca2017, Holden2017, Ulrich2019, Wang2018}. The moment magnitude of this event is estimated to be 7.8-7.9. Vertical surface displacement, derived from InSAR data (Figure 4.1,c), suggest three primary zones of large uplifts: (1) around 2-3 m uplift in the vicinity of the Leaders fault (southern faults), (2) $>5$ m uplift in the south of the Papatea fault (PF) and (3) a broader uplift zone in the northwest of the Jordan, Kekerengu, and Needles faults (northern faults).

Due to the unusual rupture complexity, previously published source models show significant disagreement in the details of how the rupture propagated over these faults \citep{Ando2018, Bai2017, Cesca2017, Holden2017, Ulrich2019, Wang2018}. In particular, kinematic finite-fault models show that
4.1. THE 2016 M7.8 KAIKOURA EARTHQUAKE

Maps of final vertical displacements derived from (c) InSAR azimuth and range offsets and (d) “Holden-Xu” Preferred model

Figure 4.1: a) “Holden-Xu” slip model of the Kaikoura earthquake. b) Comparison of observed and synthetic vertical velocity seismograms, bandpass filtered between 3 and 100 s, for selected near-field strong motion and GPS cites. Vertical displacement field obtained from (c) SAR observations and (d) forward seismic waves propagation simulation based on “Holden-Xu” model with the indication of main rupturing faults and areas. The color scale on plots c and d is oversaturated for better visibility of small features. The red star is positioned at the epicenter.
observed strong-motion and high-rate GPS data can be reproduced by rupture model without the Papatea fault (PF) [Holden et al., 2017], while others studies indicate that the Papatea fault ruptured seismically starting at \( \sim 30 \) s after \( T_0 \) [Wang et al., 2018] or simultaneously (\( \sim 60 \) s after \( T_0 \)) with the Kekerengu fault [Archuleta et al., 2018]. The difficulty in identifying and isolating seismic signals generated by the PF is also hinted from the absence of PF imprints in back-projection studies and in strong-motion and high-rate GPS data around the PF [Tan et al., 2019; Zhang et al., 2017]. While some studies [Diederichs et al., 2019] argue that elastic dislocation models fail to explain the observed surface displacement pattern [Tan et al., 2019; Zhang et al., 2017], others attempt to match the displacement data as best as then feasible [Xu et al., 2018]. Note that the earliest satellite image data were acquired a few days after the earthquake [Hamling et al., 2017], which constrains the timing of PF rupture to be within the first few days. Still, it remains unclear when (coseismic or early postseismic) and how (seismically or aseismically) the PF was ruptured.

### 4.1.2 Observable TEC perturbations

Several studies (Li et al. [2018], Bagiya et al. [2018] and Lee et al. [2018]) reported the characteristics of TEC perturbations generated by the Kaikoura earthquake. We independently reanalyze the observed CID, particularly for the purposes of model result validation and to test hypotheses on the PF evolution. In comparison, we use 1 Hz data from 72 GNSS stations on the territory of New Zealand operated by GNS Science. sTEC and vTEC are calculated using Jet Propulsion Laboratory software from raw satellite navigation system data in Receiver Independent Exchange Format (RINEX). For the calculation of IPP positions, we choose ionospheric shell height at
300 km. We use observations from 4 satellites: GPS PRN20 and 29 with elevation angles of line-of-sight between receiver and satellite (LOS) of 70–85°, GPS PRN21 with LOS elevation angles of 30–50° and GPS PRN05 with LOS elevation angles of ~25–50° and IPPs offshore to east from South Island.

Demonstrative vTEC time series are combined in Figure 4.2. vTEC perturbations, shown in vertical panels a and b, are derived by applying 10 s–10 min bandpass Butterworth filter. They are supplemented by wavelet transform evolutions underneath time-series for 0-600 s range of periods. Plots are sorted to show observations to west from the fault area on panel a and to east on the panel b, with observations to the north at the top and observations to south at the bottom. Panels c-f show absolute vTEC time-series for 4 satellite-station pairs to provide an overview of electron density values and gradients above New Zealand during the event. We also depict IPP tracks for the vTEC time series shown on panel g. Track colors are related to the amplitudes of perturbations in order to represent the positions of CID detection. All plots are limited by 11:00-12:00 UT time interval.

During the earthquake, mean absolute vTEC at the region was fairly small due to local nighttime. For the analysis of the mean state, we choose TEC observations from GPS satellites PRN20 and 29 with high elevation angles of LOS (~70–85°) to eliminate possible error from calculation of vTEC values. vTEC above New Zealand was ~7.0–12 TECu (1 TECu = 10\(^{16}\) el m\(^{-2}\)) with average positive gradient from south to north (Figure 4.2,e,f). Zonally, vTEC decreased from west to east, following the nighttime trend (Figure 4.2,a,b). Based on observations with PRN20, we estimate vTEC as 8.5–9.7 TECu in the region around focal area and ~10–11 TECu farther to
4.1. THE 2016 M7.8 KAIKOURA EARTHQUAKE

north (above North Island) with PRN29.

Strong CIDs were detected to the northeast from the focal area (PRN20-AVLN/KORO, as well as in plots on Figure 4.6,e) and reached $\sim 0.45$ TECu peak-to-peak (ptp) or $\sim 2$–$3\%$ from the background. CIDs to northeast and near to northern faults exhibit practically pure N-wave shape (e.g. PRN20-MTQN/KORO). Further from the focal area to the northeast, as shown for pair PRN20-CKID, CIDs are still comparatively strong, but their N-wave shape is markedly distorted. Moving to the west, registered perturbations are weaker – $\sim 0.2$–$0.23$ TECu ptp (PRN20-OKOH/WITH/CMBL/TKHL, PRN29-TRWH/MAHA). TEC perturbations detected farther to west and southwest from the focal area possess stronger depletion phases than enhancements (e.g. PRN20-CMBL/TKHL/YALD).

CIDs of $\sim 0.03$–$0.1$ TECu were detected offshore to the east (PRN05-KAIK) and were proposed to be driven by AWs excited from the south faults Bagiya et al. [2018]. Some perturbations appeared later at $\sim 11:30$–$11:40$ UT and $\sim 12:15$–$12:20$ UT and are present on the majority of TEC observations with GPS PRN05. The first set of perturbations of $\sim 0.1$ TECu and periods $\sim 20$ min appeared later than expected arrival of AWs excited inland, but earlier than CIDs that could be driven by tsunami-generated gravity waves (GWs) of similar periods [Occhipinti et al., 2013]. Based on the analyses of tsunami and gauge waveforms, Heidarzadeh and Satake [2017] revealed a dual-peak tsunami spectrum with energy peaks of $\sim 4.2$ and $\sim 19$ min periods that could be driven by a landslide and earthquake, respectively; an undersea landslide may explain the deficit of high-frequency content in synthetic waveforms in comparison with observations. A recently-reported study, using a globally unique data set of
repeat seafloor measurements and samples, also revealed undersea landslides triggered by this earthquake [Mountjoy et al., 2018]. As landslides, as driven by them tsunami could generate atmospheric acoustic-gravity waves (AGWs). Tsunami modeling points to long-lasting edge waves over shallow areas that could drive atmospheric waves of a broad range of periods [Gusman et al., 2018; Heidarzadeh and Satake, 2017].

Analyzing the available TEC observations, we find that there is a general trend of the strongest CIDs to the northeast with comparatively weak CIDs to the northwest. As expected, comparatively weak CIDs were detected to the south from the focal area due to dominant plasma mobility along magnetic field lines (roughly to north). However, some long-period perturbations are identified even far southeast offshore. The CIDs exhibit expected periods of ~4-7 min.

4.2 Modeling approach and assumptions

We test several finite-fault models and find the one that best-reproduces observed surface vertical displacements and TEC perturbations. Throughout the text, we refer to this preferred model as a “Holden-Xu” model, as it combines the kinematic source model of Holden et al. [2017] (Model A shown in their study) and the Papatea fault model shown by Xu et al. [2018]. We first describe the results of simulations based on the “Holden-Xu” model.

The original Holden model was derived from forward seismic wave propagation simulations using the spectral-element solver SPECFEM3D Cartesian [Komatitsch and Tromp, 2002b; Komatitsch and Vilotte, 1998] with 3D velocity and attenuation models of New Zealand, as well as topography and bathymetry. Initial fault geometry
Figure 4.2: (a-b) vTEC observations filtered with 10 s–10 min bandpass Butterworth filter and their wavelets for 0–10 min period range. (c-f) Absolute vTEC time series. (e) IPP tracks for vTEC observations shown on panels a-f and arrows represent their directions.

and final slip distribution were based on results from Clark et al. [2017]. Due to relatively minor contribution of the PF to local strong motion and GPS waveforms, the original model did not include this fault, despite the large surface displacement seen in the InSAR data (Figure 4.1,d). In the Holden-Xu model, we take non-planar fault
4.2. MODELING APPROACH AND ASSUMPTIONS

ground and slip distribution of the PF, presented in Xu et al. [2018], assume that the rupture on the PF nucleates at 57 s after $T_0$ at the southern end with the average rupture speed of 1.5 km/s, and add this PF model to the original Holden model. The slip distribution of the Holden-Xu model is shown in Figure 4.1,a and the comparison of observed and synthetic vertical velocities for selected near-field strong-motion and GPS stations is presented in Figure 4.1,b. The total moment magnitude of the simulated earthquake is Mw=7.9. The small contribution of PF seismic moment (Mw=7.3) and slow rupture speed lead to small seismic signals from the PF relative to the large moment release on the Jordan-Kekerengu-Needles faults, and hence the simulated waveforms at near-field strong-motion stations are essentially the same as in those shown in the original Holden model (Figure 4.1,b). The map of final vertical surface displacements resulting from the Holden-Xu model is in Figure 4.1,d.

In the Holden-Xu model, the rupture initiates on the Humps fault at $T_0=11:02:56$ UT and propagates predominately toward the northeast at a rupture velocity of ~1.9 km/s on the Hundalee fault. The rupture then moves onto the offshore, Point Kean fault at 31 s after $T_0$, and then continues onto the northern faults including Upper Kowhai, Kekerengu and Needles at a velocity of ~2.0 km/s. As in the original Holden model, the Holden-Xu model incorporates rupture reactivation at the southern segment of the Kekerengu fault, breaking the largest asperity with the rupture propagating at a velocity of ~1.5 km/s. This rupture reactivation, which was supported by other kinematic source inversion approach (Model B from Holden et al. [2017]), helps to explain two dominant seismic signals observed at near-field stations (Figure 4.1,b). As discussed above, the rupture on the PF begins at 57 s and propagates from the south to north. While the Hikurangi subduction interface is included
4.2. MODELING APPROACH AND ASSUMPTIONS

in the Holden-Xu model, there is little (\(\sim 8\%\)) contribution from slip on it to the overall moment. The total rupture process takes \(\sim 90\) s, with the largest seismic moment release on the northern faults, occurring at 60–80 s after \(T_0\).

We include offshore displacements to simulations without considering the propagation of hydroacoustic waves in the ocean interior. For the considered shallow ocean region (\(\sim <2\) km), the water-air interface can be considered as transparent [Godin, 2006; McDonald and Calvo, 2007], resulting in practically the same ocean surface displacements as at the ocean bottom [Levin and Nosov, 2016b; Saito, 2019], taking into account that wavelength of resolved seismic waves is longer than \(\sim 20\) km.

For this investigation, synthetic surface vertical displacements are first obtained from a numerical forward seismic wave propagation simulation as described above, for the domain of 600x600 km around focal area for 200 s from \(T_0\). Then, the displacements are re-gridded, re-interpolated, differentiated and used to impose vertical velocity lower boundary conditions for the 3D MAGIC model. These are mapped over the horizontal domain 900x900 km in latitude and longitude, with an altitude span of 500 km, at resolution of 1 and 2 km in horizontal and 0.25 km in the vertical. After 200 s, we set a reflective boundary condition at the ground surface boundary to enable realistic AW reflections. The ambient profiles of meridional and zonal winds, temperature and major neutral species densities are derived from empirical models HWM-14 and NRLMSISE00, respectively [Drob et al., 2015; Picone et al., 2002], and are actively modulated by the MAGIC dynamics as mass fractions of major gas.

An initial plasma state is calculated based on the GEMINI simulation self-consistently. We obtain similar values, trends and gradients of absolute vTEC as discussed in
previous section. Absolute simulated vTEC values are \(\sim 9\)–\(9.5\) TECu in the region of focal area. The schematic representation of numerical domains is shown on Figure 4.5,h and the profile of simulated absolute vTEC is shown on Figure 4.5,g.

### 4.3 Simulation results

#### 4.3.1 AW dynamics

We examine the nature of the PF rupture and resulting crustal deformation by quantifying its contribution to resulting AW wavefields. Figure 4.3 shows simulated AWs for 4 different finite-fault models: (1) the Holden-Xu (preferred) model, (2) Holden-Xu model without PF contribution (model by Holden et al. [2017]), (3) the Holden-Xu model without northern faults contribution and (4) isolated PF extracted from the Holden-Xu model. This figure shows horizontal plane slices of vertical fluid velocities at 2 altitudes, representing the upper mesosphere (75 km) and ionospheric F-layer (250 km). The rupture propagation direction on the PF in all these cases is assumed to be from the south to north, which is the same as in the preferred model.

Our results suggest that two main areas of surface deformation around the Humps fault in the south and around the Papatea and Kekerengu faults in the north (Figure 4.1,c) contributed to AWs. Bagiya et al. [2018] and Lee et al. [2018] attributed these zones as the source of observed TEC perturbations. Our modeling results are consistent with their findings in that the area of comparatively weak and strong AWs are excited around southern and northern faults, respectively (Figure 4.3,a). In addition, two radiations of AWs originating from the northern faults can be discerned at 75 km altitude. The first one corresponds to AWs of comparatively small amplitudes.
Figure 4.3: Simulated vertical fluid velocities at 75 and 250 km altitudes for 4 different configurations of the Holden-Xu model. Red star is epicenter position.
to the north-northeast by earlier (<55 sec after $T_0$) deformation associated with the Kekerengu fault rupture. The second, strongest AWs are excited by the later (>55 sec after $T_0$) deformation associated with both the PF rupture and the rupture re-activation on the Kekerengu fault (Figure 4.3,a,d). The amplitudes of AW vertical fluid velocities peak at 277 m/s at 250 km, such that nonlinearity – leading to shock formation – is important in their propagation. The simulation with only the southern faults show comparatively weak AWs, though still contributing to AWs of detectable amplitudes to the north and north-east (Figure 4.3,g-i).

Vertical deformation around the PF drives AWs in the atmosphere with vertical fluid velocities up to 130 m/s (Figure 4.3,j-l) and thus can substantially contribute to observed CIDs. At 75 km the amplitudes of arrived AWs to the northeast with and without the PF contribution differ markedly (Figure 4.3,a and Figure 4.3,d, respectively). At ionospheric altitudes, the difference of vertical fluid velocities to the north-northeast is ~50 m/s (Figure 4.3,c,f). The amplitudes of AWs, propagating to the north-northeast, are larger due to the PF contribution. The leading front of the resulting N-wave to the northern latitudes in the Holden-Xu model arrives earlier than in the simulation without the PF.

The spatial asymmetry of AW amplitudes can be seen throughout the whole range of altitudes (Figure 4.3,c,f,i). At ionospheric heights, differentiating between AWs from individual fault segments is complicated, due to the interference of nonlinear AWs, the substantial increase of AW wavelengths, and the thermo-viscous dissipation that markedly damp and smooth shocks in thermosphere. Nevertheless, their overall dynamics can still be discerned. The first emergence of AWs is directed to the northeast...
4.3. SIMULATION RESULTS

and is related to surface deformation around the southern faults (Figure 4.3,b). Later on, there is an arrival of much stronger AWs from the northern faults with a slight rotation of the wave-front of the strongest AWs to the north-northeast (Figure 4.3,c). The direction of rupture propagation plays a role in the AW amplitude asymmetry, even for surface displacements localized in space at the PF (Figure 4.3,j-l). We find that the strongest AWs offshore to the east are driven from the northern faults, which can be seen even in the simulation without the PF.

Our results support findings by Bagiya et al. [2018] who proposed that the sources of offshore CIDs arise from: (1) the southern faults for satellite-station pairs with IPP to southeast (e.g., PRN05-WEST/HOKI), (2) the contributions of both southern and northern faults to CIDs on latitudes of epicenter (PRN05-GLDB/NLSN) and (3) the dominant contributions of northern faults-excited AWs to offshore CIDs registered northeast (PRN05-NPLY). Bagiya et al. [2018] showed no observable CIDs for the pair PRN05-MAHO and this matches our modeling results: the strongest AWs offshore propagate nearly straight to the east from the PF area, and south from the IPP position of PRN05-MAHO, and thus cannot drive strong perturbations. The satellite-station pairs for TEC observations mentioned in this paragraph are not provided in our manuscript, and can be found on Figure 2,d from Bagiya et al. [2018].

Figure 4.3,e,j presents snapshots at early arrival times of AWs. Here, AWs excited by the strong permanent surface deformations at the focal area, as well as by transient surface displacement from seismic waves further from epicenter, can be discerned. The complex pattern of AWs is excited by seismic waves and is related to temporally and spatially varying rupture evolution on different faults.
Figure 4.4,a-d presents scaled pressure perturbations ($\tilde{p}' = p'/\sqrt{\rho/\rho_0}$, where $p'$ - pressure perturbations, $\rho$ - equilibrium density and $\rho_0$ - equilibrium density at the ground level) from the MAGIC Holden-Xu simulation for the meridional slice along 173.65°E at 4 time epochs. The permanent surface deformation drove complex packets of AWs, starting from the southern faults and later at northern faults (Figure 4.4,a). The AWs exhibit nonlinearity as they propagate upward from ground; this includes steepening of shocks, and formation and lengthening of N-waves. The AW propagation is also affected by interference and variations of the atmospheric state with altitude. The N-wave shape can be clearly discerned starting from thermospheric heights, where individual shocks coalesce, resulting in a uniform shock signature reaching ionospheric F-layer ($\sim$200 km and higher). Merging of shocks from northern faults leads to the strongest AWs propagating to north-northeast. Vertical fluid velocities at altitudes 250–300 km peak at +225/-351 m/s (10s of % of local Mach number). The amplitudes of AWs to the south are comparatively small (Figure 4.4,c).

The packets of AWs from seismic waves, propagating away from the focal area with much faster apparent horizontal phase velocities, can also be discerned (Figure 4.4,a-c). In the proximity of the focal area, vertical fluid velocities of these AWs reach amplitudes of $\sim$100 m/s. The direction of the rupture propagation, defining the main lobe of seismic wave energy, also results in stronger AWs to the north of the epicenter. Resolving ground displacements over a comparatively small region, and covering mostly the focal area, a detailed analysis of the AWs induced by seismic wave propagation (e.g. Rayleigh waves) and CIDs driven by them is not provided and would require separate investigation.
4.3. SIMULATION RESULTS

Figure 4.4: (a-d) Altitude-latitude diagrams of normalized pressure fluctuations for the slice along 173.65°E at 4 time epochs. Time-altitude diagrams of normalized (e) and absolute (f) vertical fluid velocities for the position of PF (42.2°S/173.65°E). Color scales are oversaturated for better visibility of weak signatures.

Figure 4.4,e,f include time-altitude diagrams of scaled and absolute vertical fluid velocities, respectively, at the position of the PF. Again, the coalescence of leading shocks to a single N-wave can be seen from ~100 km. Later on, long-lived dynamics, driven by AWs trapped between lower thermosphere and ground, and tunneling to
higher altitudes, is present. AWs leak energy to ionospheric heights practically for the whole shown time period, with stronger amplitudes \(\sim 30-40\) minutes after the earthquake, representing the arrival of second packet of trapped AWs from the ground. AWs refracted from stratospheric heights can be seen (Figure 4.4,e).

The modeling results suggest that the strongest AWs were excited to the north-northeast from the focal area. The contribution from the southern faults is comparatively small, and their corresponding deformation results in weak AWs. At ionospheric heights, the N-wave shock is markedly smoothed by dissipation, and the interference of shocks prevents separation of AWs signals originating from individual faults. The PF markedly contributes to the AWs to the north and northeast.

### 4.3.2 Ionospheric plasma responses to AWs

The GEMINI results from the Holden-Xu simulation for the latitude-altitude slice along 173.65\(^{\circ}\)E are presented in Figure 4.5,a-c. The strongest perturbations in ionospheric plasma are driven to the north, and this asymmetry is connected as with dominant plasma mobility along magnetic field lines [Zettergren and Snively, 2015], as with the propagation of the strongest AWs to north-northeast. At 250 km and higher, the CID\(s\) possess a uniform leading front to the south and to the north. Seismic wave-induced AWs also drive CID\(s\) farther from the focal area, though of considerably smaller amplitudes. The propagation of AWs results in electron density perturbations up to \(\sim 30\)\% from the background unperturbed state, and up to \(\sim 15\)\% in ion temperature. Panel g of Figure 4.5 shows the trend of absolute vTEC for the hour after the event at the position 173.98\(^{\circ}\)E/40.3\(^{\circ}\)S. It can be seen that absolute vTEC values match the observations (see Figure 4.2,c,d,e), and follow the same
4.3. SIMULATION RESULTS

declining nighttime trend.

Stronger (weaker) electron density perturbations are generated by seismic waves to the north (south) (Figure 4.5,a-c). Again, this is driven by two factors: (1) the directivity of rupture propagation that defines the direction of main energy lobe for seismic waves and subsequently strongest AWs, and (2) wave vector that has larger projection along magnetic field lines for seismic wave AWs propagating to north, whereas for AWs propagating to south the wave vector is more perpendicular to magnetic field lines (Figure 4.5,b).

Figure 4.5,d-f presents latitude-time diagrams of simulated vTEC perturbations using 3 geometries of LOS: (d) zenith-looking observations, (e) 50° elevation and 180° azimuth angle observations, and (f) 30° elevation and 0° azimuth angles observations. Azimuth angle is set from the north and the geometries of LOS are shown on Figure 4.5,h for reference. The coordinates are specified for IPP positions. We use ionospheric shell height at 300 km (as for vTEC calculation based on observations) and Single Layer Mapping function to recalculate sTEC to vTEC [Parkinson et al., 1995]. The CIDs exhibit latitudinal asymmetry, with stronger perturbations to the north (Figure 4.5,d). Head shocks to the south do not drive significant observable enhancements in vTEC, whereas following rarefaction phase (possessing higher amplitudes) generates detectable initial decreases in vTEC. This is consistent with the vTEC observations shown on Figure 4.2 for PRN20-METH/YALD. For southward-pointing geometry (with LOS practically perpendicular to plasma motion direction at ionospheric F-layer heights, as shown on Figure 4.5,h), the mapped vTEC perturbation reach ~0.9 TECu peak-to-peak, what is almost twice higher than observed with GPS
PRN20 and 29 with LOS of 70–85° elevation angles from east to west and from south to north, respectively (Figure 4.5,e). The vTEC on panel f is derived from sTEC with LOS practically along plasma motion direction and exhibit amplitudes of ~0.15 TECu. This illustrates phase cancelling effect, as TEC fluctuations tend to be the greatest when TEC LOS lies along CID phase front [Georges and Hooke, 1970].

Figure 4.5: (a) Field-aligned ion drift velocities, (b) electron density perturbations and (c) ion temperature perturbations in percentage from background from preferred Holden-Xu simulation for the meridional slice along 173.65°E at T=625 s. (d) Latitude-time diagram of simulated vTEC calculated with zenith integration of electron densities. (e) Latitude-time diagram of vTEC perturbations recalculated from sTEC with 50° elevation angle of LOS pointing to south. (d) Latitude-time diagram of vTEC perturbations recalculated from sTEC with 30° elevation angle of LOS pointing to north. Data on plots are oversaturated for better visibility of weak features. (g) Absolute simulated vTEC for the position 173.98°E/40.3°S. (h) Configuration of numerical domains and direction of LOS for latitude-time diagrams shown on panels d-f.
4.3. SIMULATION RESULTS

Using 2D ionospheric simulations along meridional slices, we compare synthetic vTEC perturbations with observations. First, the coordinates of IPP positions at the time of leading CID arrivals are found from TEC observations. Data are used from GPS PRN20 and 29 only, for which head shock arrival can be clearly discerned, as for example shown on Figure 4.2,c,d. Then, among this dataset we choose observations that are close to the meridional slices for which we perform ionospheric simulations, representing slices to the west and to the east from the focal area. Finally, to compare we use the same latitudes of IPP positions from simulations as found from TEC observations. In summary, we match IPP positions at moments of CID arrivals from observations and simulations and compare vTEC perturbations. For GEMINI simulation we integrate electron densities vertically (90° elevation angle) for every point on the numerical domain to represent zenith TEC LOS, while in observations we use vTEC recalculated from sTEC values with 75-85° LOS elevation angles.

For the precise comparison of TEC observations with simulation results, we would need 3D ionospheric simulations and track every satellite-station LOS, changing elevation and azimuth angles temporally. Thus, in our comparison, some biases are expected later in time, because we match only positions of initially appearing CIDs with fixed meridional slices from simulations; in reality, IPP positions are constantly moving. Also, the periods of CIDs may be biased from Doppler shift effect (caused by IPP position motion relative to CIDs propagation), while our simulation results are free from this effect due to the assumptions of fixed LOS. Another source of possible inconsistency is the effect of mutual geometry between the CID propagation direction and the LOS that may differ in our simulation and in reality, as discussed in the previous paragraph. This inconsistency is assumed to be small for chosen observations.
4.3. SIMULATION RESULTS

with high elevation angles of LOS, which is supported by the agreement with results. In addition, IPP positions can be offset, as we do not know the exact vertical electron density profiles for each location to calculate ionospheric shell height. However, at ionospheric altitudes AW wavelengths are \( \sim 100-160 \) km and they possess comparatively fast speeds \( \sim 1.1 \) km/s, relative to the slow motion of IPPs for the chosen high elevation LOS observations. Thus, we assume that, for the comparison of leading AWs and CIDs driven by them, our approach is sufficient and, likely, as good as feasible without performing a full sTEC comparison using fully 3D models.

Panels a-d of Figure 4.6 show time-distance diagrams of filtered vTEC observations with GPS PRN20, 21 and 29 and vTEC perturbations from our GEMINI Holden-Xu simulation. For diagrams based on observations we use all measurements from position \( 42.2^\circ \text{S}/173.98^\circ \text{E} \) to \( 433 \) km to north and include TEC data with IPP positions in the range \( 172-176^\circ \)E. Panel d represents simulation results for the slice along \( 173.98^\circ \)E from the same latitude to \( 433 \) km to north. The N-wave shape of CIDs for all 3 satellites observations and simulation can be seen, and at distances of \( \sim 150-250 \) km the tail shock of N-wave is split into two, resulting in a complex shape of CIDs. This shape is present in the majority of time series with GPS PRN20 and PRN29 and IPPs close to the focal area to north. Black lines are added to each plot to the same positions and indicate that the general dynamics of simulated CIDs are fairly close to those in reality. From the travel-time diagram on Figure 4.6,g, the CIDs driven by head shock of N-wave have similar speeds as in observations - \( \sim 1.1-1.3 \) km/s.

The resulting time-series comparisons are presented on Figure 4.6,e. For all plots, we indicate the longitudes of CID arrivals that are estimated from observations (first
4.3. SIMULATION RESULTS

value at bottom-right on each plot) and longitudes of meridional slices used from simulations (second value at bottom-right). Overall, we find good agreement with TEC observations, in terms of the timing of onsets of CIDs appearance and their amplitudes and shapes. The latter indicates also a reasonable reproduction of the nonlinearity, present in the observed wave forms.

Trapped AWs also drive long-lasting CIDs that can be seen from observations and simulation results (e.g., Figure 4.6,a,d). Some of later-emerging CIDs might have come from early aftershocks (as large as M6.2), although their contributions should be relatively small. Our results suggest that these long-lasting CIDs must have been driven by trapped AWs (Figure 4.4,e,f), similar to those reported in Saito et al. [2011]. Note that the recent study by Heidarzadeh et al. [2019] successfully simulated observed run-up of 7 m with the modeling of undersea landslide ~10–20 min after the earthquake, that potentially contributed to observed TEC perturbations.

Finally, simulation results suggest only very limited formation of an ionospheric “hole” - the decrease in electron density persistent over tens of minutes and driven by a higher rate of plasma recombination and quasi-permanent transport of plasma [Mendillo, 1988; Zettergren et al., 2017]. Its duration is ~20 minutes and magnitude ~0.03 TECu. We do not find any signatures in TEC observations that can be connected with ionospheric “hole” with high confidence, particularly because its magnitude is close to the TEC threshold level.
4.3. SIMULATION RESULTS

Figure 4.6: (a-c) Travel-time diagrams of 10s–8min bandpass filtered vTEC to the north from the position of PF with GPS PRN21, 29 and 20. (d) Travel-time diagram of modeled vTEC perturbations for the meridional slice along 173.98°E. (e) Observed (blue) and simulated (red) vTEC perturbations. Both time series are 20s–8min bandpass filtered.
4.3.3 Seismo-ionospheric imagery

As demonstrated earlier, shock waves reaching altitudes of 250 km and higher, driven by complex, spatio-temporal earthquake deformation, coalesce into a single N-wave (Figure 4.2 and 4.4). Such common dynamics resembles the confluence of shocks excited by supersonic aircraft: while shock-wave signatures at close distances depend on the source parameters, individual signals at far distances coalesce into N-wave [Carlson et al., 1966]. In a homogeneous atmosphere, the resulting shape does not depend on the source [Carlson and Maglieri, 1972]. In general, however, the shape of N-waves may be markedly distorted by atmospheric stratification, winds and turbulence [Crow, 1969; Marchiano et al., 2005; Pierce and Maglieri, 1972; Sabatini et al., 2019a], as well as nonlinear effects and focusing [Inoue and Yano, 1997; Sabatini et al., 2019b]. The dynamics is complicated by initial formation of shocks that may evolve at different altitudes due to spatial variability of AW sources. However, the duration, amplitudes and times of arrival of the resulting N-waves at far-field does depend on the source [Howes, 1967]. Thus, the variations in rupture dynamics result in different characteristics of N-wave in ionosphere and subsequent TEC perturbations, and this information must be carefully interpreted on a nonlinear basis – assessing amplitudes and time of arrivals of the nonlinearly-evolved wave field, rather than assuming linearity of signal propagation.

In order to understand the sensitivity of kinematic source parameters on resulting TEC perturbations, we consider various finite-fault models and their configuration, particularly related to the kinematics of the PF. In the finite-fault models of Wang et al. [2018], the rupture on the PF initiates at the northwest of the fault at $\sim$27
4.3. SIMULATION RESULTS

Table 4.1: Finite-fault models and configuration of the PF used in the analysis.

<table>
<thead>
<tr>
<th>Model</th>
<th>PF nucleation time</th>
<th>PF rupture direction</th>
</tr>
</thead>
<tbody>
<tr>
<td>1. Preferred Holden-Xu</td>
<td>$T_0 + 57$ s</td>
<td>From south to north</td>
</tr>
<tr>
<td>2. doi:10.1002/2017GL075301</td>
<td>N/A</td>
<td>N/A</td>
</tr>
<tr>
<td>3. WANG201844</td>
<td>$T_0 + 27$ s</td>
<td>From NW to SE</td>
</tr>
<tr>
<td>4. Holden-Xu</td>
<td>$T_0 + 27$ s</td>
<td>From south to north</td>
</tr>
<tr>
<td>5. Holden-Xu</td>
<td>$T_0 + 27$ s</td>
<td>From NW to SE</td>
</tr>
<tr>
<td>6. Holden-Xu</td>
<td>$T_0 + 57$ s</td>
<td>From NW to SE</td>
</tr>
<tr>
<td>7. Holden-Xu</td>
<td>$T_0 + 57$ s</td>
<td>From north to south</td>
</tr>
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</table>

sec after $T_0$ and propagates toward the southeast. Hence, the timing of the PF rupture in this model is 30 sec earlier with respect to our preferred model described in Section 4.2. To assess whether the timing and rupture direction of the PF can be constrained by the TEC observations, we first compare results of finite-fault models with 5 different PF rupture kinematics (Models 1-5 shown in Table 4.1). In each case, we run seismic wave propagation, atmospheric and ionospheric simulations as described in Section 4.2. Resulting AW and TEC signals are compared for the region of maximum contribution of the PF to the north and northeast, in the same manner as discussed previously.

Figure 4.7 presents snapshots for two time epochs and scaled pressure signals at two altitudes (150 and 250 km), in Models 1,2,3 and 5. The results with other models are presented in the Supporting Information. We find that simulations with models 4 and 5 (i.e, earlier PF rupture initiation) show 20 and 25% lower amplitudes of vertical fluid velocities of AWs than in Model 1, respectively. Simultaneous rupture of the Kekerengu fault and the PF in Model 1 lead to a stronger N-wave before reaching the altitudes where dissipation becomes dominant, resulting in stronger plasma perturbations (Figure 4.7,a and m). However, in the case of earlier rupture initiation on the
4.3. SIMULATION RESULTS

Figure 4.7: Latitude-altitude diagrams of vertical fluid velocity for the meridional slice along 173.98°E from simulations with models 1,2,3 and 5 at (a-d) T=500 s and (e-h) T=620 s. (i-l) Electron density perturbations for the same models as above at T=620 s. m) Scaled pressure signals at 150 and 250 km for the position 42.08°S/173.98°E (along dashes lines though plots a and e). Color scales at plots a-h are oversaturated by 1.5 times for better visibility of weak signatures and absolute maximum velocities are provided for each plot.

PF (i.e., Models 3 and 5), the confluence of shocks does not form a uniform N-wave until ~230 km, where dissipative damping is already substantial, finally resulting in weaker plasma perturbations (panels c,d,g,h and m on Figure 4.7). In Models 3 and
5, the amplitudes of AWs are weaker than those in Model 1, which can be explained by the combined effects of earlier rupture initiation and the difference in the rupture propagation direction. Moreover, Models 3 and 5 show similar AW wavefields and pressure signals at thermospheric heights to the north and northeast. At 150 km, pressure signals exhibit separate leading shocks in both cases. At 250 km, pure N-wave is formed, but with lower amplitudes than in Model 1. Model 2 (i.e., without the PF) show practically uniform N-wave at 150 km, but exhibits smaller amplitudes (~35%). Figure 4.7,i-l, shows electron density perturbations at 620 s, the same as on panels e-h. Here, again, the strongest CIDs are simulated with Model 1, weaker with models 3 and 5 and fairly weak with Model 2, practically to the same percentages. In summary, these results demonstrate that AWs and CIDs are both sensitive to the time of the PF rupture initiation (or to the absence of the PF) and differences vary in the range ~20-35%.

Figure 4.8 presents the comparison of simulated TEC signals based on Models 1,2 and 3 with high elevation angle LOS TEC observations to north and northeast. The results based on Model 3 show on average ~38-40% lower amplitudes of TEC perturbations than from simulation with Model 1 or observations. The absence of the PF in Model 2 shows ~30% lower amplitudes of TEC perturbations and later arrival of CIDs than detected with observations. However, amplitudes of TEC perturbation with Model 5 are practically the same as with Model 1 (Figure 4.8,a). This shows that TEC observations, as an integrated values through wave structures, may be insensitive in particular LOS geometry relative to the plasma motion, though AW amplitudes and plasma density perturbations are ~20–25% weaker in Model 5 than in Model 1.
In order to demonstrate this, we compare sTEC from Models 1 and 5 with different geometries of LOS in Figure 4.8,b-d. Zenith-looking TEC observations show practically no useful difference in perturbations (Figure 4.8,b); however the slant geometry of TEC observation demonstrates differences in time of CID arrivals and CID amplitudes (Figure 4.8,c,d). As it was demonstrated in Section 6.2.4, the IPP position and spatial configuration of LOS relative to the direction of plasma motion, play a crucial role in the detection of perturbations. When CID intersects LOS at altitudes $\sim$200-250 km, as shown on Figure 4.8,c, TEC amplitude differences between two simulations are negligible, but TEC remains sensitive to the time of CID appearance ($\sim$13 s). In case when plasma motion intersects LOS at altitudes $\sim$250-350, the difference in time of CID appearance is negligible, but amplitudes differ by $\sim$20\% (Figure 4.8,d).

We further explore the influence of rupture propagation direction on the resulting AW amplitudes and TEC perturbations. From Models 6 and 7, we find that the PF rupture propagation direction affects AW amplitudes to the north and northeast, but only to $\sim$7–10\% in vertical fluid velocities. Taking into account possible uncertainties in the initial electron density, it is difficult to draw any strong conclusions on the PF rupture propagation direction effects, at least with 2D simulations and proposed comparison approach of simulated vTEC with high-elevation TEC LOS observations. The difficulties are exacerbated by the fact that large vertical surface displacements occurred at two areas over the main PF strand: (1) at its northern end and (2) to southeast, near the shore (Figure 4.1,d) (see e.g., Diederichs et al. [2019]; Langridge et al. [2018]), resulting in a complex pattern of excited AWs in the atmosphere. This should be revisited in 3D simulations with sTEC.
4.4 Discussion

For the 2016 M7.8 Kaikoura earthquake, Bagiya et al. [2018] and Lee et al. [2018] mentioned apparent horizontal phase speeds of CIDs as 1.05–1.4 km/s based on vTEC observations. They pointed that these values exceed the speed of sound in F-layer of the ionosphere. In addition, Lee et al. [2018] concluded that travel-time delays in their synthetic TEC waveforms may be related to the errors in a velocity model used. Although this conclusion may take place, we propose two other reasons for higher
apparent speeds of CIDs. First, the curvature of the AW phase fronts relative to TEC LOS and inhomogeneities in ionospheric F-layer plasma distribution may result in apparent variations of horizontal phase velocities calculated from vTEC observations. Second, this can be explained by nonlinear regime of propagation of AWs. In addition to the substantial increase of the mean atmospheric temperature in the thermosphere, which leads to higher speeds of sound, the faster speed of propagation of head shock of N-wave (and slower speed of propagation of tail shock) leads to the lengthening of the N-wave pulse. Within the weakly nonlinear regime, the speed of the shock relative to the center of N-wave can be found as \( v = \phi c_0 (\gamma + 1)/2\gamma \), where \( \phi = p/p_0 - 1 \), \( \gamma \) - adiabatic index, \( c_0 \) - speed of sound and \( p \) and \( p_0 \) are disturbed and equilibrium pressures, respectively [Beyer, 1974]. For non-negligible pressure changes, the speed of head shock always exceeds the speed of sound, while the tail shock lags, whereas center point of N-wave (point of zero fluid velocity) propagates with the speed of sound. As it was mentioned, our simulated CIDs exhibit practically the same horizontal velocities as in observations. In addition, Figure 4.9,b presents the diagram of shock speeds calculated based on the formula above, summed up with local mean speed of sound for every point of the numerical domain for the meridional slice along 173.98°E at 660 s, when the head shock is in the region of F-layer of ionosphere. As the provided formula neglects terms of order \( \phi^2 + O \) (along with consideration of plane wave propagation in a homogeneous fluid without wind), we still obtain close values of head shock speed as from the MAGIC-GEMINI simulations – 1.141 km/s vs \(~1.1–1.3 \text{ km/s}\).

Figure 4.9,a presents the field of maximum absolute vertical fluid velocities, reached at each point for the considered meridional slice along 173.98°E and Figure 4.9,c - maximum electron density perturbations for the same slice. These maximum values
Figure 4.9: (a) Wave-field of maximum absolute vertical fluid velocities for the full hour of simulation for the meridional slice along 173.98°E. (b) Speed of shocks relative summed up with speed of sound for each point for the meridional slice along 173.98°E at T=660 s. (c) Field of maximum electron density perturbations calculated for the full hour of the simulation for the meridional slice along 173.98°E.

are derived considering the dynamics for the whole hour after the event. The AW wavefield exhibits a meridionally-asymmetric pattern due to the complexity of the source and rupture direction of propagation. The strongest perturbations are driven by AWs to the north, with weaker perturbations to the south and above southern faults. AWs exhibit the strongest fluid velocities at 250-300 km, with a substantial decrease of amplitudes at higher altitudes driven by thermo-viscous dissipation. A similar spatial pattern can be seen in the maximum electron density perturbation field. The strongest perturbations were driven to the north at \( \sim 280-310 \text{ km} \), which supports our choice of an ionospheric shell height at 300 km. Thus, the amplitudes of resulting CIDs depend on both - source and plasma response effects.

The present case study corroborates our previous investigation of the formation of a single N-wave reaching ionospheric F layer with the loss of individual signatures [Inchin et al., 2020a]. At thermospheric heights, larger amplitudes of formed N-wave result in stronger distortions of its shape. Particularly, the curvature of the central
part of the signal is seen from the results with Model 1, where maximum amplitudes of N-wave are simulated. For Model 2, central part of N-wave exhibits only slight distortion, whereas N-wave possesses practically straight central part with Models 3 and 5. This may be due to the strongly nonlinear propagation of AWs, and the importance of higher order effects in $p/p_0$, where $p$ - pressure and $p_0$ - equilibrium pressure [Sabatini et al., 2019b].

Bagiya et al. [2018] proposed that separation between various tectonic sources of AWs at near-field within rupture duration of $\sim 100$ s may not be possible, and explained this by the much shorter time of rupture propagation in comparison with periods of CIDs ($\sim 4$ min). Our results demonstrate the importance of the directivity of rupture propagation to resulted AW wavefield, even at ionospheric heights. Inchin et al. [2020a], based on modeling results of atmospheric and ionospheric responses for the 2015 M7.8 Nepal Gorkha earthquake, showed that the direction of rupture propagation, as a supersonic moving source of AWs in atmosphere, in general defines the vector of propagation of the strongest AWs. Although the Gorkha earthquake exhibited unilateral rupture over a much longer fault length than the Kaikoura earthquake, strong correlation between the direction of rupture propagation and propagation of the strongest AWs is present in both cases. In addition, CIDs of 4-6 minute periods in ionosphere may results not only from AWs of same periods as excited at the surface, but also from nonlinear effects on AW propagation that lead to the formation of N-waves, followed by their lengthening and subsequent smoothing at ionospheric altitudes by thermo-viscous damping. Thus, in addition to complex plasma responses to AWs (magnetic field-seized plasma mobility and plasma distribution), we highlight that source effects also play a vital role. The need to consider finite-faulting processes
for large earthquakes at near-fields is dictated not only by seismological principles, but also by spatial asymmetry of AWs, excited into the atmosphere and propagating through the whole range of altitudes.

The analysis provided in Section 4.3.3 indicates that TEC observations can be used to constrain the finite-fault kinematics of a large earthquake. We found that the absence of the PF results in substantially lower amplitudes of simulated TEC perturbations (~30%), even with zenith LOS. The need for the inclusion of the PF into finite-fault model strongly suggests that the PF was ruptured coseismically, and also shows that TEC can be used for constraining the kinematics of finite-fault models. This finding may be particularly relevant for large undersea earthquakes with lack of geophysical and near-field data. The weakest perturbations in the source model by Wang et al. [2018] are caused by lower amplitudes of surface deformations and differences in the rupture kinematics; although, as we demonstrated, similar AW and CID dynamics to north-northeast can be obtained by re-configuring the PF rupturing process in the Holden-Xu model. The full sensitivity analysis of the rupture kinematics of the PF including its direction of rupture propagation requires full 3D ionospheric simulations. Particularly, with the precise tracking of LOS and the use of low elevation angle slant TEC observations (GPS PRN 05, 21 and 29), it is plausible that further steps toward resolving the PF rupture can be made. In addition to the GPS data used in this study, observations with other constellations (e.g., GLONASS) may markedly improve spatial density of observations for the comparison with modeling results.

Small amplitude of simulated ionospheric “hole” may lead to difficulties with its detection, as it can also be obscured by dynamics from other sources. This also
opens important question of ionospheric “hole” observation applicability for tsunami early-warning systems proposed by Kamogawa et al. [2016b] during nighttime events. Although the uplift and subsidence of the crust are substantial (up to \( \sim 8-10 \) m) and result in strong nonlinear AWs, the simulated ionospheric “hole” is still fairly weak for this case. In addition, nevertheless large tsunami-generating earthquakes are expected to produce TEC perturbations and much deeper depletions [Zettergren and Snively, 2019], recent study by Astafyeva et al. [2014] suggests that strike-slip earthquake may also result in strong TEC perturbations and this question requires further investigation.

4.5 Conclusion and future work

Through numerical simulations, we investigated the dynamics of infrasonic acoustic waves (\( \sim 0.17 \) mHz - 0.1 Hz) and disturbances in ionospheric plasma generated by the 2016 M7.8 Kaikoura earthquake. We simulated time-dependent surface deformation using forward seismic wave propagation code, which were used to drive the 3D nonlinear compressible neutral atmosphere model MAGIC. Plasma responses excited by the earthquake-induced atmospheric infrasonic acoustic waves were then simulated with the 2D nonlinear multi-fluid ionospheric model GEMINI.

First, we investigated the contribution of the Papatea Fault (PF) and southern and northern faults to atmospheric and ionospheric signals through the use of different configurations of a finite-fault model. We found that our preferred finite-fault model reproduces the amplitudes, shapes and time epochs of appearance of detected TEC perturbations most accurately. The absence of the PF showed \( \sim 30\% \) lower amplitudes
of TEC perturbations, suggesting that the PF was ruptured within the coseismic phase of the earthquake. We then varied the timing of rupture initiation on the PF and its rupture propagation direction, and demonstrated that these parameters result in differences of amplitudes of AWs reaching ionospheric heights and corresponding electron density perturbations. However, we found that for chosen high elevation angle observations, useful for the comparison with 2D ionospheric simulation results, vTEC is not sensitive to these kinematic source parameters. At the same time, we showed that in case of the 3D ionospheric simulations and precise tracking of LOS of slant TEC perturbations, the timing and the direction of the PF rupture would be constrained. Thus, we found substantial justification for future fully-3D slant TEC (sTEC) investigations of seismically generated infrasonic acoustic wave fields towards reconstructing earthquake processes.

We highlight the following primary conclusions:

1. The regime of propagation of AWs, driven by large earthquakes, can be weakly to strongly nonlinear, leading to substantially different dynamics in comparison with linear assumptions. In this case, direct numerical simulations are the most comprehensive way to resolve AW propagation through the whole range of altitudes and to subsequently reproduce accurate CIDs.

2. For the simulation of AW excitation, and subsequent CIDs from the earthquake’s source region, finite-fault models should be considered, particularly for complex events. Rupture propagation (and its direction) plays an important role in the spatial asymmetry of observed CIDs, and can be assessed separately from observational biases imposed by the geomagnetic field. Simple point or axisymmetric sources lead to marked biases in spatial and temporal dynamics of AWs that in turn may lead
to misinterpretation of the physics underlying earthquake-atmosphere-ionosphere processes.

3. Trapped (resonant) AWs occur between thermosphere and ground, leaking some part of their energy into ionosphere, can drive long-appearing CIDs, observable in TEC even an hour after the event. These have been widely reported in prior investigations, e.g., by Matsumura et al. [2011b].

4. TEC observations may supplement seismological studies through the investigation of different finite-fault models and their ability to reproduce detected ionospheric perturbations. However, TEC observations (as they are integrated spatially along specific line of sights) are affected by the geometry of LOS and plasma motion direction. Thus, TEC, in a particular geometry of LOS, may be insufficiently sensitive to the differences in amplitudes of electron density perturbation. For such studies the analysis of vTEC perturbations presented here can be improved by tracking TEC LOS precisely, and comparing time of CID arrival and their amplitudes based on slant TEC observations.

It is anticipated that modeling case studies based on full 3D coupled atmosphere-ionosphere model can provide additional insights on earthquake-atmosphere-ionosphere processes. Such simulations, though currently computationally intensive, may provide a significant path forward to constrain finite-fault models of large earthquakes, especially for complex events.
Chapter 5

The dynamics of tsunamigenic acoustic-gravity waves

In this Chapter we report simulation results of TAGW dynamics based on the 2011 Japan Tohoku-Oki tsunami case study, that incorporate tsunami evolution over realistic bathymetry. Separately, we investigate source and bathymetry effects on TAGWs through parametric 2D and 3D simulations with simplified and demonstrative ITDs and ocean depth variations. Inclusion of accurate tsunami evolutions is important to understand the longer-period evolutions of ionospheric responses in vicinity of undersea earthquakes [Galvan et al., 2012]. The results in this chapter form the basis of a manuscript to be submitted for review in consideration for publication.

Our results demonstrate that TAGW propagation is affected by the atmospheric state and nonlinear evolution. Substantial amplitudes of TAGWs in the thermosphere can lead to instabilities, followed by the excitation of secondary acoustic and gravity
waves, spanning a broad range of periods. Bathymetry variations play a crucial role on TAGW characteristics. Particularly, ocean depth changes result in TAGW amplitude increase or decrease at different altitudes, as the whole TAGW packet tilts from the variations of the intrinsic frequency, exhibiting different dissipation altitudes. Focusing of tsunami waves over rises and their interaction with seamounts and islands can lead to the substantial enhancement of generated TAGWs. Long-period TAGWs that propagate ahead of the tsunami wavefront may generate early-detectable perturbations in the mesosphere and thermosphere, whereas TAGW dissipation also leads to the excitation of secondary AGWs. Our modeling results suggest that TAGWs can drive detectable and quantifiable perturbations in the upper atmosphere, under a wide range of scenarios, but also uncover new challenges and opportunities for their observations. The results of the 2011 Tohoku-Oki tsunami case study are presented in Section 5.2 and of parametric studies in Section 5.3. The discussion and summary of these investigations are provided in Section 5.4.

5.1 Numerical simulation approach

As a realistic scenario, we study the large tsunami that was generated by the 2011 M9.1 Tohoku-Oki subduction megathrust earthquake near the east coast of Honshu Island, Japan (epicenter at 38.297°N/142.373°E, 05:46:24.120 UT, USGS). The earthquake rupture area was estimated as ~400 km along-strike and ~220 km across the width [Lay, 2018]. Source inversion studies point to the slips on a fault of 30-50 m during ~100 s. On the shore of Japan, waves reached ~15-20 m in height with a run-up of more than 20 m [Fritz et al., 2012; Mori et al., 2011]. An unprecedented
amount of TAGW-driven Earth’s magnetic field, airglow and ionospheric plasma disturbance data were collected at near-epicentral region [Galvan et al., 2012; Liu et al., 2011; Maruyama and Shinagawa, 2014; Saito et al., 2011] and far-fields [Azeem et al., 2017; Hao et al., 2013; Makela et al., 2011; Yang et al., 2014, 2017].

The complexity of the earthquake resulted in studies leading to a wide range of proposed finite-fault models that produce different peak vertical displacements at the seafloor of 7–22 m [Lay, 2018]. Through simulations of the Tohoku-Oki tsunami, MacInnes et al. [2013] investigated the reproduction of wave gauge data with 10 different ITDs. Based on the analysis of their and our own modeling results, here we specify temporally-varying ITD, calculated in a forward seismic wave propagation simulation based on finite-fault model by Shao et al. [2011].

We specify 500 m Gridded Bathymetry Data J-EGG500 with 30' USGS GTOPO topography near the Japan Islands area and 1' ETOPO1 bathymetry from the National Environmental Satellite, Data and Information Service (NESDIS) archive for open ocean. Bottom friction is incorporated with a Manning roughness coefficient of 0.025. To avoid boundary discontinuities, we apply a smooth taper at the edges of the ITD domain prior to incorporating it into GeoClaw. ITD dynamics are resolved for 12 minutes after earthquake’s rupture initiation for the region of 1200x1200 km around the epicenter. The tsunami evolution is computed for 7 hours, until direct waves reach the farthest boundary of the numerical domain.

TAGW dynamics are resolved for 7 hours in a domain of 6000x5490x500 km in latitude, longitude and altitude directions, respectively. The horizontal and vertical resolutions are chosen as 5 and 1 km, respectively, resulting in 0.6588B grid points of
5.2. TOHOKU-OKI TSUNAMI CASE STUDY

the numerical domain. We apply sponge layers near the edges of MAGIC numerical domain to avoid boundary wave reflections. Meridional and zonal winds and temperature profiles for altitudes up to 55 km, were used from the MERRA-2 database. For higher altitudes, empirical models NRLMSISE00 [Picone et al., 2002] and HWM-14 [Drob et al., 2015] were specified. MERRA-2 data are selected for 15:00:00 LT, 14 minutes after rupture initiation. The configurations of models for parametric 2D and 3D studies are provided in Section 5.3.

5.2 Tohoku-Oki tsunami case study

This section contains 5 figures that demonstrate TAGW characteristics and evolution from different perspectives and through the whole range of altitudes, distances and times. Figure 5.1 presents snapshots from the simulated ocean surface vertical velocities (panel a), and absolute major gas temperature perturbations (T’), sliced horizontally at 4 altitudes (panels c-f). The snapshots of T’ for chosen meridional and zonal slices, shown with dashed lines in Figure 5.1,d, are provided in Figure 5.1,g-l. Time-distance diagrams of ocean surface vertical velocities and T’ at 50, 150, 250 and 350 km altitudes for zonal and meridional slices are presented in Figure 5.2. Power spectral density (PSD) diagrams for the zonal slice of the ocean surface vertical displacements and T’ at 4 altitudes are provided in Figure 5.3. The bathymetry of the numerical domain, obtained through NESDIS services, and the field of maximum simulated tsunami amplitudes are presented in panels a and b of Figure 5.4, respectively. The fields of maximum T’ from horizontal slices at 50, 150, 250 and 350 km are shown in Figure 5.4,c-f. The fields of maximum horizontal and vertical
fluid velocities for meridional and zonal slices are depicted in Figure 5.5 along with corresponding to these slices bathymetry profiles. All maximum perturbation fields are calculated incorporating the dynamics during 7 hours of simulation.

Tsunami evolution is influenced by bathymetry variations, resulting in refraction, reflection, focusing and branching of waves [Mofjeld, 2000; Satake, 1988]. The strongest waves are simulated to the southeast of the focal area (Figure 5.4,b). The propagation of the tsunami to the north and northeast is markedly affected by shallow bathymetry near the shore of Japan and the Kuril Islands. Long-lived dynamics, driven by trapped and deflected waves, is discernible near the coast of Japan (Figure 5.2,a,b). Along with the shores, notable tsunami wave reflections result from the Hawaiian-Emperor Seamount Chain (HESC) to the east, Izu-Bonin-Mariana Arc (IBMA) and Ogasawara Plateau to the south and the Mid Pacific Mountains (MPM) to the southeast. Apparent focusing of waves is driven by Shatsky and Hess rises. Tsunamis exhibit a broad spectrum of periods in the range of \( \sim 7-60 \) min with a dominant peak at \( \sim 19-20 \) min, and horizontal wavelength \( (\lambda_x) \) varying between 140-400 km (Figure 5.3,a). The average apparent phase speed of the direct waves is \( \sim 231 \) m/s, though its marked change results from bathymetry variations, for example over Shatsky Rise or IBMA. The comparison of synthetic and DART wave gauge data is provided on Figure 5.1,b.

The strongest perturbations in the atmosphere are generated over the focal area by strongly nonlinear AWs from intense ocean surface displacements (Figure 5.5). In this area, the perturbations are present even 5 hours after the earthquake (Figures 5.1,f and 5.2,i,j) and exhibit periods of \( \sim 4.2 \) min (Figure 5.3). They result from
Figure 5.1: The snapshots of (a) ocean vertical velocities and (c-f) T' at 4 altitudes: (g-l) absolute and scaled T' for meridional and zonal slices shown with dashed lines on panel d. (b) Simulated ocean surface vertical displacements and DART wave gauge data.
5.2. TOHOKU-OKI TSUNAMI CASE STUDY

acoustic waves (AWs) that are trapped between the ground and lower thermosphere and tunneling into the upper layers (Figures 5.1,f,g,i and 5.2,h). Similar long-lived dynamics were observed for at least 4 hours after the earthquake in TEC data [Saito et al., 2011; Tsugawa et al., 2011]. With time, the simulated AWs are slightly shifted to the east from the focal area, that is explained by the dominant eastward wind from ground to \(~110\) km height (wind profiles are provided on Figure 5.5,a,d). It should be noted that the chosen resolution of the numerical grid is suitable primarily for TAGW dynamics to meet computational expenses, whereas short-period acoustic waves can be under-resolved. AW dynamics and generated mesopause airglow perturbations, driven by this ITD, are discussed in Chapter 6.

TAGW packet structure is preserved for the whole range of distances and altitudes (Figure 5.1,g-l). Notable leading phases in the TAGW packet exhibit periods of \(~25-45\) min and propagate ahead of the tsunami (Figures 5.1,c-j and 5.2,d,g,h,j). Possessing substantial amplitudes from the ground to lower thermosphere, they experience marked damping above \(~150\) km, as having comparatively small vertical wavelengths \((\lambda_z)\) of \(~150-200\) km \((\lambda_z \sim 250-500\) km), leading to their dissipation at lower altitudes [Vadas, 2007]. The leading phases are substantially reflected in the stratosphere (Figure 5.1,h,j). Dominant TAGW phase, slightly trailing behind the tsunami wavefront and locked with a dominant tsunami period of \(~19\) min (at thermospheric heights \(\lambda_x \sim 260\) km, \(\lambda_z \sim 260\) km and \(v_z \sim 228\) m/s), in general exhibits stronger amplitudes above lower thermosphere than leading phases (Figure 5.2,g-j). From the ground to the mesosphere, its amplitude is usually lower than of the leading phases, though marked variations are present at all altitudes (Figure 5.4,g,i). The next trailing phase has comparatively short \(\lambda_x \sim 190\) km, but large \(\lambda_z \sim 300-330\) km in the thermosphere.
5.2. TOHOKU-OKI TSUNAMI CASE STUDY

This phase front may exhibit substantial amplitudes at ionospheric heights, but does not produce any notable perturbation below thermosphere. The following train of TAGWs includes ducted and resonating waves, as well as TAGWs from refracted and reflected tsunami waves, spanning a broad range of periods, though with some components of comparatively small amplitudes.

Thermo-viscous molecular dissipation of TAGWs at thermospheric heights, and transience within the tsunami-driven wave field, are causes for the local generation of secondary AGWs (SAGWs). They propagate ahead of the TAGW packet with an apparent horizontal phase velocity \( (v_x) \) of \( \sim 600 \) m/s and exhibit amplitudes up to \( \sim 50 \) K (Figure 5.1,f,g). A particularly strong SAGW is driven from the near-epicentral region and propagates practically through the whole domain. SAGWs have large \( \lambda_x \) of \( \sim 600-700 \) km and periods of \( \sim 14-18 \) min. Our results demonstrate an agreement with a previous study by Kherani et al. [2015], who reported similar characteristics of these SAGWs. Interesting to note the local generation of SAGWs from the transience within the tsunami-driven wave field as TAGWs packet tilts horizontally from the change of the intrinsic frequency. Indeed, as it can be seen from Figure 5.2,g,i, in the majority of cases, the horizontal tilt of TAGW packet is followed by the excitation of SAGWs.
Figure 5.2: Time-distance diagrams of (a,b) ocean surface velocity and (c-i) T' for zonal (left column) and meridional (right column) slices at 4 altitudes. Chosen slices are shown with dashed lines on Figure 5.1,d.
At thermospheric heights, TAGWs exhibit temperature perturbations of $\sim 150$–$250$ K in average, though some local peaks reach $600$–$700$ K (Figure 5.4,b). Vertical fluid velocities, starting from lower thermosphere, vary in a range $\sim 150$–$200$ m/s, whereas horizontal fluid velocities are $\sim 150$ m/s on average. The phase speed of the tsunami and TAGWs at altitudes lower than $\sim 100$ km is practically the same of $\sim 231$ m/s (Figure 5.2,c) and TAGWs exhibit similar amplitude variations as the tsunami (Figure 5.4,c). At thermospheric heights, there is strong variability of TAGW phase speeds (Figure 5.2,g-j) and amplitudes (Figure 5.4,d-f), which result from bathymetry variations and nonlinear effects. Although the dispersive nature of the TAGW packet leads to its spreading [Laughman et al., 2017], this effect takes comparatively long distances prior to becoming clearly discernible.

The gap in the fields of maximum perturbations of $\sim 300$-500 km from epicenter, where
5.2. TOHOKU-OKI TSUNAMI CASE STUDY

practically no perturbations are observed in the upper atmosphere, is explained by the distances that need to be covered by TAGWs before they reach upper layers [Occhipinti et al., 2013]. Then, the strongest TAGWs propagate at distances 500-800 km to the southeast from the focal area and are driven by the substantial amplitudes of the tsunami at near-epicentral region (Figure 5.4,b,g). The leading phases exhibit comparatively strong amplitudes at ~153°E and ~130-150 km altitudes. The following trailing phases appear in lower thermosphere already with practically vertically-aligned phase fronts and apparent instabilities arise right after them (Figure 5.1,k). Leading phases induce mean flow and increase local \( v_x \) (Figure 5.5,d), which results in the horizontal acceleration of lagging phases in the direction of TAGW packet propagation, that start exhibiting larger \( v_z \) and phase distortions. As the speed of the mean flow exceeds phase velocities of TAGWs, breaking of waves occurs. Further to the southeast, apparent breaking at the tail of the TAGW packet occurs, but at altitudes of ~200 km (Figure 5.1,l). Again, prior to breaking, trailing phases start tilting vertically as they are shifted to higher intrinsic frequencies and larger \( \lambda_z \). At these altitudes, the amplitudes of leading phases are already comparatively small, and we attribute instabilities to the induced mean flow by the dominant phase. The dynamics strongly support that the simulated instabilities arise via self-acceleration effects from strong mean flow induced by the leading phases at the head of the TAGW packet [Dong et al., 2020; Fritts and Lund, 2011; Fritts et al., 2015]. Along with the generation of long-period SAGWs, TAGW breaking leads to the generation of AWs (e.g., Snively [2017], and references therein) that drive long-lived dynamics (Figures 5.3,d and 5.4,f and 5.5,d,e). The TAGW breaking may be in part enhanced from the interaction of TAGWs excited by direct tsunami waves and TAGWs from tsunami
waves that are first reflected from the coast of Japan and then closely follow direct tsunami waves (Figure 5.1,a). As TAGWs from the direct tsunami slows down over Shatsky Rise, they can more readily interact with the TAGWs from the reflected tsunami.

The spreading of ocean waves over the passage of Shatsky Rise leads to the decrease of their amplitudes, but focusing intensifies the tsunami to the southeast (Figure 5.4,b). At altitudes lower than ~100 km, TAGWs practically mimic the tsunami wavefront (Figure 5.1,c). They exhibit similar initial decrease and further enhancement of amplitudes, and two intensified region of T’ to the southeast and east can be discerned (Figure 5.4,c). In the lower thermosphere, the Shatsky Rise effect produces similar effects, although less prominent. At 250 and 350 km, there is no elongated southeastward enhancement of TAGW amplitudes, but instead the pattern of perturbations practically mimics the shape of Shatsky Rise: from Tamu Massif at the southwest to Shirshov Massif at the northeast (Figure 5.4,a,e,f). We attribute this enhancement in the perturbation to TAGW self-acceleration and following instabilities from tsunami passage over Shatsky Rise, as it is accompanied by substantial increase of horizontal fluid velocities in these regions (Figure 5.5,d).

In a north-south direction and meridional slice shown with a dashed line on Figure 5.1,d, TAGWs possess comparatively smaller amplitudes (~150-200 K) and do not display nonlinearity, in contrast to east and southeast, and no apparent TAGW instabilities occur. Even at 150 km altitude, the leading phases exhibit the strongest amplitudes (Figure 5.2,d,f). At higher altitudes, the dominant phase exhibits the strongest amplitudes and the lagging phase is practically absent at ionospheric F-
Figure 5.4: (a) Bathymetry of the numerical domain. (b) The field of maximum simulated vertical ocean surface velocities. (c-f) The fields of maximum $T'$ at 4 altitudes. The data on plots are shown on an oversaturated scales for better visibility of weaker features.
layer (Figure 5.2,h,j). After the passage of Ogasawara Plateau, the dominant phase amplitude is decreasing locally, while leading phases enhance (Figure 5.2,j). Over shallower bathymetry the tsunami speed decreases, followed by the diminishing of $\lambda_z$ of TAGWs. The whole TAGW packet tilts horizontally and the leading phases intensify (though are still damped at ionospheric heights drastically), whereas trailing phases start dissipating at lower altitudes. This can be seen from the increase of the meridional fluid velocities (from the intensified leading phases) at altitudes lower than $\sim 200$ km and the diminishing of vertical fluid velocities and $T'$ above $\sim 250$ km (Figure 5.5,a,b). Notable decrease of vertical and the enhancement of horizontal fluid velocities can also be seen farther to the south, as the tsunami runs into a shallower bathymetry of IBMA and the TAGW packet tilts horizontally.

The reflections of the tsunami from the HESC result in ocean surface waves and TAGWs of notable amplitudes as they return to the Japan coast (Figure 5.2,a,g). This is supported by observed TEC perturbations from these reflected waves [Tang et al., 2016]. The arrival of the simulated and observed reflected TAGWs from the HESC to the shore of Japan is $\sim 6$ hours after the earthquake. Simulated $T'$ of these TAGWs reach $\sim 100$–$140$ K at $250$ km. Notable tsunami reflections to the north from the focal area result in a second TAGW packet that follows the TAGW packet generated by direct tsunami waves propagating to south (Figure 5.1,g). These two packets are spatially separated and do not interact. The reflection and refraction of tsunami waves from the HESC leads to the excitation of short-scale waves in the ocean and subsequently short-scale TAGWs in atmosphere (Figure 5.3), which transit to longer periods away from HESC.
The interaction of the tsunami with seamounts and islands leads to the generation of compact, but markedly intensified short-scale TAGWs (Figure 5.4,d). These TAGWs are particularly apparent to the east of the IBMA in a quadrant 15–30°N and 150–165°E (Figure 5.4,b). However, they are markedly attenuated at higher altitudes and are practically not present at 250 km (Figure 5.4,e). In the southeast quadrant of the domain, the perturbations are absent at almost all altitudes, which results from tsunami damping from the interaction with the MPM. Notable perturbations at 50 and 150 km follow the same way to the north from the MPM as the tsunami (Figure 5.4,c,d), but this evolution is practically not present at 250 and 350 km.

Finally, the evolution of the tsunami with distance leads to the filtering of small-scale ocean waves and TAGWs, as can be seen in a zonal direction between 142-160°E in Figure 5.3. Over the inland part, in the absence of a quasi-continuous source (tsunami), the TAGW energy peak is shifting from shorter to longer periods and horizontal scales. Dissipative filtering drastically affects phases with longer $\lambda_z$ Heale et al. [2014], but even 10° to the west, the power of the TAGW is still substantial at thermospheric heights (Figure 5.3,d,e). This explains the detection of perturbation in TEC at far inland distances from shores and similar dynamics were observed in TEC data over the continental U.S., where initial $\lambda_z$ of $\sim$150-250 increased to $\sim$250-400 km with distance Azeem et al. [2017].
5.3. EFFECT OF TSUNAMI SCATTERING ON TAGWS

The 2011 Tohoku-Oki tsunami case study shows that bathymetry variations may markedly affect the amplitudes and characteristics of TAGWs. In this section, we continue this analysis based on simulations with simplified and demonstrative bathymetry variations and ITDs.

\textbf{Figure 5.5:} Fields of maximum horizontal and vertical fluid velocities for (a,b) zonal and (d,e) meridional slices shown with dashed lines on Figure 5.4.b. (c,f) Bathymetry profiles for chosen meridional and zonal slices. The data are shown on an oversaturated scales.

5.3 Effect of tsunami scattering on TAGWs

The 2011 Tohoku-Oki tsunami case study shows that bathymetry variations may markedly affect the amplitudes and characteristics of TAGWs. In this section, we continue this analysis based on simulations with simplified and demonstrative bathymetry variations and ITDs.
5.3. **EFFECT OF TSUNAMI SCATTERING ON TAGWS**

We use a single 1D or 2D circular Gaussian model as the ITD:

\[
ITD(x_i, y_j) = A \cdot e^{-\left(x_i-x_c^2\right)/(2\sigma^2)} \cdot e^{-\left(y_j-y_c^2\right)/(2\sigma^2)}
\]  

(5.1)

where the ITD is set as instantaneous, \(x_c, y_c\) - position of the center of the peak, \(\sigma\) - standard deviation and A - amplitude, set as 0.6 m. Tsunami dynamics for the 2D TAGW simulations are calculated in the Cartesian 1D version of GeoClaw. The species densities and temperature profiles are utilized from the Tohoku-Oki case study. Winds are not included in these idealized simulations to avoid complexity.

We start with the investigation of the effect of ITD size on TAGWs. Figure 5.6,a-d shows the simulation results of TAGW dynamics generated by 4 different tsunamis with \(\lambda_x=107, 154, 214\) and 297 km and periods 9, 13, 18 and 25 min, respectively. The bathymetry is set as flat at 4 km depth and the amplitudes of ITDs are decreased by a factor of 100, in order to exclude nonlinear effects. For each case, we provide a x-z snapshot of T’ and a corresponding 2D wavelet of \(\lambda_z\), time-distance diagram of T’ at 320 km, and \(\lambda_x\) and period wavelets for slices shown with blue lines on the time-distance diagram. The data on the time-distance diagrams are shown with oversaturated scales for better visibility of weaker features and the maximum and minimum values are indicated. As expected, tsunamis of different periods result in different dominant intrinsic frequencies and thus phase velocities of the TAGW packet. The amplitudes of the long-period leading phases that reach 320 km become notable with the increase of the tsunami period, while trailing phases are apparent with the decrease of the tsunami period. In common, shorter period tsunamis result in smaller amplitude TAGWs in thermosphere, though tsunamis with periods of 13 and 25 minutes show fairly similar amplitudes in thermosphere. The dominant \(\lambda_z\) in
5.3. EFFECT OF TSUNAMI SCATTERING ON TAGWS

thermosphere of these TAGWs varies in the range \(\sim 147-156\) km, whereas \(\lambda_x\) increases markedly with the increase in tsunami period, from \(206\) km for a 9 min period tsunami, to \(264\) km for a 25 min period tsunami. Thus, tsunami waves and their coupling with AGWs clearly establishes the observable spectrum in the atmosphere.

Figure 5.6,c,e,g,h shows simulation results of TAGW dynamics generated by a tsunami with a dominant period of 18 min that propagates over flat bathymetry of 2, 4, 6 and 8 km. Shallower bathymetry results in shorter tsunami’s \(\lambda_x\), and a shortening of dominant \(\lambda_z\) of generated TAGWs, from \(236\) km in case of 8 km depth to \(92\) km for 2 km depth. Shallow bathymetry are causes for a notable presence of leading TAGWs, even at 320 km, that exhibit comparable amplitudes to the dominant phase (Figure 5.6,e). In case of deep bathymetry, leading phases are practically not presented, but lagging phases appear. The maximum TAGW amplitudes at thermospheric heights are 2.28, 4.84, 5.91 and 1.141 K for 2, 4, 6 and 8 km depth bathymetry, respectively. In case of 8 km bathymetry, practically all phases are evanescent starting from the stratosphere (scaled \(T'\) snapshot is provided in Figure 5.6,h). Although at thermospheric heights some of these phases are freely propagating, they exhibit small amplitudes. The dominant \(\lambda_x\) at 320 km varies from \(\sim 270\) km for 2 km depth to \(\sim 230\) km for 6 km depth bathymetry.

In Figure 5.6,i-k, we provide simulation results with gradually increasing ocean depth from 2 km to 8 km over 5500 km distance. The increase of ocean depth results in a constant change of the relative amplitudes of the horizontal and vertical fluid velocities of TAGWs, with different phases dominating for different ocean depth. This can also be seen in the time-distance diagram of \(T'\) at 320 km shown in Figure 5.6,k.
5.3. EFFECT OF TSUNAMI SCATTERING ON TAGWS

Figure 5.6: (a-h) The results from 8 parametric simulations presented with altitude-distances (x-z) slices of $T'$ and corresponding $\lambda_x$, 2D wavelets, travel-time diagrams of $T'$ at 320 km, as well as $\lambda_x$ and period wavelets for slices at 320 km shown with blue lines on travel-time diagrams. Values in travel-time diagrams indicate horizontal apparent phase velocities. (i-j) Fields of maximum horizontal and vertical fluid velocities from the simulation with constantly decreasing bathymetry, (k) Time-distance diagram of $T'$ at 320 km altitude.

At some time epochs, two phases may exhibit similar amplitudes at one altitude. The whole TAGW packet tilts gradually vertically with increasing ocean depth and the
rise of the intrinsic frequency. In general, TAGWs drive stronger perturbations at higher altitudes as the ocean depth increases. However, it is not the rule; for example, fluid velocities decrease at ~250-300 km altitudes at distances of ~2500 and ~5500 km. Over some depth, the constantly tilting TAGW packet may provide a fairly small signal at these heights when its phases exhibit lower dissipation altitudes (the altitude where thermo-viscous dissipation starts dominating wave amplitude growth from the decrease of atmospheric mean density). As the bathymetry reaches a depth of ~7 km or deeper, TAGW in the thermosphere start to lose intensity, as most of the generated TAGWs are evanescent at altitudes lower than 100 km.

Next, we address the bathymetry effect based on 2D simulations with simplified ocean depth variations. The snapshots of absolute and scaled T’ are presented in Figure 5.7. Distance-altitude plots present the TAGW dynamics after the passage of bathymetry feature (variation) that is located at x=0 km. Except bathymetry features, the depth of the ocean is set as 4 km in all simulations. The ITD generates a tsunami with λx ~214 km, as shown in panel c of Figure 5.6, but here its amplitude peaks at 0.6 m height. Figure 5.7,a includes simulation results with flat bathymetry for reference, and in this case T’ reaches ~191 K in the thermosphere.

Arcs and chains of seamounts cause a marked change of tsunami characteristics, reflection and trapping of waves. We present simulation results with different shapes of arcs. A narrow arc is represented with a Gaussian function with σ=70 km and a wide arc, comparable with the Tamu Massif of Shatsky Rise, is represented by a Gaussian function with σ=300 km. A plateau is represented by a bathymetry depth change from 4 to 0.5 km with an extension of ~267 km and an uplift and downlift of 31
km using a tapered cosine function (useful for the representation of the plateau):

\[
ITD(x) = \begin{cases} 
\frac{1}{2} \left\{ 1 + \cos\left(\frac{2\pi}{r} [x - \frac{r}{2}] \right) \right\}, & 0 \leq x < \frac{r}{2} \\
1, & \frac{r}{2} \leq x < 1 - \frac{r}{2} \\
\frac{1}{2} \left\{ 1 + \cos\left(\frac{2\pi}{r} [x - 1 + \frac{r}{2}] \right) \right\}, & 1 - \frac{r}{2} \leq x < 1 
\end{cases}
\] (5.2)

where \( r \) - cosine fraction and set as 0.25 (Figure 5.7,f). The narrow arc does not markedly affect TAGW propagation, with only localized enhancement of the leading phases and diminishing of amplitudes of the dominant phase (Figure 5.7,b). However, in case of wide arc, initial TAGW packet is markedly dissipated prior to a newly formed TAGW packet (formed from waves after arc passage) appearing at later time epochs and distances (Figure 5.7,c). After the formation of the new TAGW packet, slowly dissipating leading TAGWs from the initial packet may still generate notable perturbations over \( \sim 100-1500 \) km distances.

The propagation over the wide arc also results in a local generation of TAGWs that can be discerned at 85 km, but they are weak at thermospheric altitudes. In the case of tsunami propagation over the flat plateau, part of the tsunami waves are reflected back from the plateau uplift and generate TAGWs propagating toward the ITD. Trapped and propagating over the plateau tsunami waves generate short-scale TAGWs that are practically all dissipate below lower thermosphere.

In panels d and e of Figure 5.7, we show the result of simulations with uplift and downlift escarpments with 2 km depth variation over 75 km distance. Such escarpments are present, for example, at IBMA, Tonga Ridge, Ryukyu Arc, shelves etc. The propagation of the tsunami over the escarpment results in a change of its \( \lambda_x \) and \( v_x \).
5.3. EFFECT OF TSUNAMI SCATTERING ON TAGWS

Figure 5.7: (a-e) Altitude-distances slices of absolute and scaled $T'$ and travel-time diagrams of absolute temperature perturbations at 85 and 320 km for bathymetry variation cases. Numbers of travel-time diagrams indicate horizontal apparent phase velocities in m/s. (f) Altitude-distance slices of absolute and scaled $T'$ for simulation with plateau.
5.3. EFFECT OF TSUNAMI SCATTERING ON TAGWS

The TAGW packet, driven by the tsunami prior to reaching the escarpment, continues propagating in the same direction. Without continuous forcing by the tsunami, TAGWs disperse and dissipate, transitioning toward larger dominant $\lambda_x$. The newly formed TAGW packet appear $\sim 400$ and $\sim 600$ ahead of the escarpment at 85 and 320 km altitudes, respectively. As the packet tilts horizontally (exhibiting lower intrinsic frequency) over shallow bathymetry, the maximum TAGW amplitudes at 85 km height enhance from 0.55 to 0.8 K, whereas at 320 km TAGW amplitudes fall drastically from 40 K to 9.8 K. The reverse situation is present after the passage of the downlift escarpment, where TAGW amplitudes become smaller at 85 km (from 0.51 to 0.46 K) and larger at 320 km height (from 6.46 to 22.78 K).

Next, we present simulation results of continued TAGW propagation over land, as the tsunami reaches the coast. ITD generates a tsunami with a dominant period of $\sim 18$ min and $\lambda_x \sim 214$ km over flat bathymetry of 4 km depth. For reference, we run the 3D simulation with flat bathymetry of 4 km depth and the results are presented in Figure 5.8,a,b,e. TAGWs generated over flat bathymetry propagate concentrically away from the source. The generation of SAGWs in thermosphere, propagating ahead of the TAGW packet, can be seen. Again, the leading phases are more prominent at lower altitudes, whereas phases with larger $\lambda_z$ dominate in the thermosphere. The fields of maximum perturbations are not uniform, as TAGWs disperse spatially and SAGWs are excited that propagate separately (Figure 5.8,e). A straight-line shore is set as a transition of ocean depth from 4 to 0 km over 45 km extension. The shore is shown in Figure 5.8,c,d,f with a dashed line. The distance between ITD and the shore is set as 2000 km in a straight direction. The propagation of TAGWs over the inland part leads to comparatively fast filtering of the dominant and lagging
5.3. EFFECT OF TSUNAMI SCATTERING ON TAGWS

phases that possess longer $\lambda_z$ (Figure 5.8,c,d). As it was demonstrated by Heale et al. [2014], this filtering is driven by faster dissipation of phases with larger $v_z$, that reach thermospheric heights earlier and dissipate first. Far inland, the leading phases of the TAGW packet start exhibiting the largest amplitudes. In the vicinity to the shore, packets of small-scale TAGWs are excited and are driven by shortening of the tsunami wavelength as it reaches shore, though they dissipate below $\sim 200$ km altitude.

As the tsunami approaches to the shore obliquely, reflected and direct tsunami waves interfere at the vicinity of the shore, producing an obliquely intensified region. The same pattern can be seen in the TAGW fields of maximum perturbations at distances 100-500 km to inland part, where constructive and destructive interference of TAGWs leads to regions of enhanced and decreased perturbations. At 85 km altitude, TAGW amplitudes decrease rapidly as they propagate over inland part of the domain, exhibiting $\sim 50\%$ lower amplitudes 500 km and $\sim 15\text{-}18\%$ at distances 1000 km on shore. TAGWs in the thermosphere dissipate with a much slower rate. At 320 km altitude and distances 500 km on shore, TAGWs still exhibit practically the same amplitudes as near the shore, and 1000 km on shore they fall to $\sim 30\text{-}50\%$.

Finally, based on the 3D simulation, we demonstrate the effect of tsunami focusing and tsunami shore approach on TAGW evolutions. We set a rise using a single circular Gaussian model of 300 km diameter, which is roughly represents the Tamu Massif of Shatsky Rise. The peak of the rise is at 2 km depth, whereas the area around is set as flat with 4 km depth. The snapshots in Figure 5.9,b-e present the simulated $T'$, sliced horizontally at 4 altitudes. Tsunami evolution for the same time epochs are provided in Figure 5.9,a and the maximum tsunami vertical velocities and maximum
5.3. EFFECT OF TSUNAMI SCATTERING ON TAGWS

Figure 5.8: The snapshots of sliced horizontally $T'$ at 4 altitudes from simulation with (a,b) flat bathymetry and (c,d) bathymetry with a shore, shown with dashed vertical lines. (e,f) The fields of maximum $T'$ at 4 altitudes for (e) flat bathymetry simulation and (f) bathymetry with a shore.

$T'$ at 85, 150, 250 and 320 km are provided in Figure 5.9,k-o.

As they pass over the rise, the tsunami waves become superposed in the vicinity of a cusped caustic (Figure 5.9,g). Such evolution is discussed by Berry [2007], and we find similar dynamics in the case of the Tohoku-Oki tsunami focusing from Shatsky Rise. TAGWs also exhibit correlated dynamics at higher altitudes, e.g., at 85 and 150 km where they are superposed, leading to amplification and interaction. Initially, the central part of the TAGW front bends while propagating over the rise (Figure 5.9,b,c); then, enhanced TAGWs arise inside the caustic from the superposition of
5.3. EFFECT OF TSUNAMI SCATTERING ON TAGWS

Figure 5.9: The snapshots of (a) ocean surface velocity and (d-e) sliced horizontally T’ at 4 altitudes from simulation with rise presence. Time epochs of snapshots are indicated on panel a. (e,f) The fields of maximum ocean surface vertical velocity and T’ at 4 altitudes.

waves (Figure 5.9.g,h). At altitudes of 250 and 320 km, the bending of the TAGW front is followed by an initial marked decrease of TAGW amplitudes (from the tilting of TAGW packet horizontally due to the decrease of intrinsic frequency) and at some distances, the signal is practically absent (Figure 5.9.e). Later on, focusing of waves can be seen inside the caustic (Figure 5.9.i,j). Leading components of the TAGW packet, formed prior to the passage of the rise can be discerned, propagating ahead of the newly formed TAGW packet (Figure 5.9.j).

Tsunami focusing results in wave amplification and subsequent intensification of generated TAGWs (Figure 5.9,k). At 150 km, TAGW amplitudes rocket more than twice
the original value and reach 88.5 K in direct distances after the passage of rise (Figure 5.9,m). At higher altitudes, TAGWs enhance to 1.2-1.5 times their initial amplitude and reach 70.3 K at 250 km and 27.3 K at 320 km (Figure 5.9,n,o). Again, as in the case of the Tohoku-Oki tsunami, we see that TAGWs mimic the evolution of the tsunami at 85 km, whereas at higher altitudes, the amplitude variations exhibit more complex patterns. Although we demonstrate focusing of a comparatively small tsunami, such strong amplification of TAGWs clearly indicate that in case of large tsunamis, this amplification can readily lead to localized nonlinear effects.

5.4 Discussion and conclusion

Through numerical simulations, we investigated the dynamics of acoustic and gravity wave dynamics generated by tsunamis. Simulated nonlinear ocean surface wave evolutions are used as a source of AGWs in the three-dimensional compressible and nonlinear neutral atmosphere model. We performed a case study of the 2011 Tohoku-Oki tsunami, as well as parametric numerical studies with demonstrative bathymetry variations and ITDs. This section summarizes main outcome of these studies.

The TAGW packet has discernible structure, with phase variations from long-period (but long $\lambda_x$) phases at the head of the packet to short-period (but short $\lambda_x$) phases in its tail. The dominant phase is locked to the dominant tsunami phase velocity. This supports earlier findings by Vadas et al. [2015] on TAGW packet excitation as a mix of discrete and continuum spectral components, both above and below the fundamental mode. The tsunami continuously generates TAGWs in the atmosphere, representing a quasi steady-state forcing (analogous to a moving, evolving mountain).
Thus, phases in the tail of TAGW packet with larger $\lambda_z$ do not dissipate earlier, as in the case of source-free propagating AGWs where trailing short-period phases reach higher altitudes and dissipate earlier [Heale et al., 2014]. Phases in the tail of the TAGW packet usually exhibit stronger amplitudes at thermospheric heights than the leading phases, whereas at lower altitudes the leading phases may dominate.

Slower (faster) tsunamis or their deceleration (acceleration) over bathymetry features leads to a horizontal (vertical) tilt of the whole TAGW packet as its intrinsic frequency varies. This lead to the change of dissipation altitudes of components in the packet, which in turn leads to the variations of phase fronts that drive the strongest perturbations at different altitudes; the dominant phase, locked with a dominant tsunami period, does not necessarily provide the strongest perturbations. For example, in a shallow bathymetry case (or long period tsunami) the dominant phase amplitude can be comparable with amplitudes of the leading phases even at ionospheric altitudes, whereas deep bathymetry result in the absence of the leading phases above lower thermosphere. Although very deep bathymetry drives higher intrinsic frequency AGWs into the atmosphere (as the tsunami propagates faster), the resulting TAGWs become evanescent in the stratosphere. With 8 km bathymetry, practically all phases from the TAGW packet are evanescent, in some cases tunneling upward and becoming propagating in the thermosphere (e.g., Snively and Pasko [2008], and references therein). As it was demonstrated by Wu et al. [2016], there is a barrier of $2NH < V < c$ (N - Brunt-Vaisalla frequency, H - scale height, V - tsunami speed, c - speed of sound) that serves as a filter to TAGWs in the atmosphere.

The intensification of thermo-viscous dissipation at thermospheric heights filters small-
scale TAGWs, along with tilting of phase fronts vertically as they shift to higher intrinsic frequencies [Hines, 1968; Vadas, 2007]. Leading phases of the TAGW packet and thermospherically-generated SAGWs may drive earlier than tsunami arrival perturbations, which is consistent with previous finding [Bagiya et al., 2017; Kherani et al., 2015; Vadas et al., 2015]. For example, the coherent structure of TEC perturbations and ionospheric airglow that are detected ahead of the tsunami over Hawaiian Islands by Makela et al. [2011] ($\lambda_x = 290 \pm 12.5 km, v_x = 184.5 \pm 33.8 m/s$ and $\lambda_x = 189.9 \pm 4.9 km, v_x = 222.9 \pm 52.4 m/s$), points to leading phases as their source, whereas SAGWs propagate with much faster $v_x$. However, further investigation is needed to assess the sources of all apparent secondary and primary fast AGWs.

Nonlinear evolution of TAGWs, generated by large tsunamis (particularly along a main lobe of tsunami energy), can lead to substantially different dynamics in comparison with linear assumptions. Self-acceleration effects cause the distortion of TAGW phases, which are followed by instabilities. In addition to recent modeling results on self-acceleration effects and instabilities in lower thermosphere and below [Dong et al., 2020; Fritts and Lund, 2011], our results suggest they can develop in the lower regions of the F-layer of the ionosphere. TAGW instabilities lead to the generation of acoustic and gravity waves, spanning broad range of periods, many that are radiated downward to form ducted waves that persist after the TAGW’s passage.

Undersea seamounts and plateaus, rises and escarpments cause reflection, refraction and trapping of ocean surface waves, as well as their acceleration or slowdown. We demonstrated that these dynamics may also markedly affect TAGW characteristics and amplitudes. Thus, highly varying bathymetry in the West Pacific Ocean (Shatsky
and Hess rises, HESC, MPM, IBMA etc.) drastically affects TAGW characteristics. This also seems to be true for the Indian Ocean, where marked undersea features (e.g., Ninetyeast Ridge, Diamantina Fracture Zone) can result in a substantial variability of tsunamis and subsequent TAGWs. Focusing and branching of tsunami waves cause their amplification up to an order in magnitude [Berry, 2007; Degueldre et al., 2016] and TAGWs can experience similar variations of amplitudes. Bathymetry variations may also lead to the superposition of TAGWs phases inside the packet. Below the thermosphere, TAGW mimic the tsunami wave evolution, exhibiting the same amplitude variations. At ionospheric altitudes, only large undersea scale massifs (such as Shatsky Rise) results in notable change of TAGW characteristics. Finally, TAGWs may propagate inland and still drive comparable perturbations even $\sim$1000-1500 km away from the shore, filtering toward larger $\lambda_x$. These outcomes are also supported by previous observational studies, for example by Azeem et al. [2017], finding travelling ionospheric disturbances (TIDs) driven by the Tohoku-Oki tsunami based on TEC observations as far inland as western Colorado, while horizontal wavelength of TEC disturbances increased with distance.

It seems plausible that tsunami heights can be retrieved from upper atmosphere observations and can be useful for future applications and tsunami early-warning systems [Rakoto et al., 2018; Savastano et al., 2017]. However, it is not yet clear how accurately the ocean response can be retrieved while incorporating nonlinearity in the atmosphere and bathymetry effects in the ocean. Further studies can be directed toward the investigation of mesopause and ionospheric airglow signatures, as well as ionospheric plasma responses to TAGWs and characteristics of detected signals with an incorporation of realistic bathymetry and ITDs. Studies may also address
the dispersive nature of tsunamis and full 3D ocean-atmosphere coupling, that will provide deep insight into atmosphere responses to undersea earthquakes and tsunamis generated by them.
Chapter 6

Mesopause airglow responses to AGWs from earthquake and tsunamis

6.1 Introduction

Here, we report the results of the investigations of mesopause airglow (MA) perturbations generated by AGWs during large earthquakes and tsunamis. First, we present hypothetical simulation results for a nighttime equivalent of the 2011 M9.1 Tohoku-Oki earthquake, driven by surface deformations and the resulting tsunami, and then for the 2016 M7.8 Kaikoura earthquake in New Zealand, which occurred at night but for which no data were successfully collected. The results shown in Chapter 6.2 were published in [Inchin et al., 2020b].
6.2 Mesopause airglow disturbances driven by nonlinear infrasonic acoustic waves generated by large earthquakes

6.2.1 Modeling approach and assumptions

The 2011 M9.1 Tohoku-Oki earthquake in Japan occurred at 14:46 LT (JST) 11th of March 2011 and was one of the largest recent megathrust earthquakes, causing a devastating tsunami. The hypocenter was located at ~70 km off the coast of Japan (38.103°N, 142.861°E) at the depth of ~24 km (Japan Meteorological Agency, JMA). An unprecedented amount of TEC data were collected during the earthquake and tsunami in Japan [Galvan et al., 2012; Tsugawa et al., 2011] and, for example, on the Hawaiian islands [Makela et al., 2011] and the continental part of the USA Azeem et al. [2017]. Important aspects of AGW dynamics and coseismic ionospheric disturbances were described, including body and surface wave driven AW perturbations in the ionospheric plasma, acoustic resonances and plasma depletion [Liu et al., 2011; Maruyama and Shinagawa, 2014; Saito et al., 2011]. Yang et al. [2017] observed tsunami-induced airglow emission perturbations retrieved from the Sounding of the Atmosphere using Broad-band Emission Radiometry (SABER) instrument on board the Thermosphere-Ionosphere-Mesosphere Energetics Dynamics (TIMED) spacecraft. Makela et al. [2011] reported observations and simulation of 630.0 nm ionosphere airglow emission perturbations, driven by the tsunami over the Hawaiian Islands.

Numerous studies were dedicated to earthquake/tsunami-atmosphere-ionosphere coupling process modeling for this event. Recent investigations have confirmed the non-
6.2. MESOPAUSE AIRGLOW DISTURBANCES DRIVEN BY NONLINEAR INFRASONIC ACOUSTIC WAVES GENERATED BY LARGE EARTHQUAKES

linear acoustic waves above the source in detail. Among them, Shinagawa et al. [2013] reported simulation results of driven plasma disturbances using nonlinear non-hydrostatic compressible fluid equations based on a neutral gas model coupled with a single-fluid collisional plasma model in 2D. Matsumura et al. [2011a] simulated near-field AGW dynamics based on a 2D nonlinear compressible neutral gas model. Zettergren et al. [2017] presented modeling results of ionospheric responses to a ground level axisymmetric source of AGWs in a 2D nonlinear atmosphere model, and electron density depletions ("ionospheric holes" formation) in TEC signatures were simulated with a 2D multi-fluid plasma model. Meng et al. [2018] simulated ionospheric responses in 3D, using a point source representation of excited AGWs at ground level, and assuming a linear regime of their propagation at altitudes lower than ~100 km. Zettergren and Snively [2019] simulated AGWs and plasma perturbations using a 3D multi-fluid ionospheric model coupled with a 2D axisymmetric nonlinear compressible atmospheric model, and investigated the latitudinal asymmetry of plasma responses, dynamics of driven electric currents and geomagnetic field perturbations. These studies point to the excitation of strong near-epicentral AGWs, causing substantial ionospheric responses, in agreement with TEC observations. However, in these studies, ground-level AGW sources are highly simplified, omitting time and spatial dynamics of rupture, which extended for hundreds of km and lasted more than a minute [Satake et al., 2013]. In addition to simplified modeling assumptions, this may result in marked inconsistency with real coupling mechanisms and preclude comparisons of the resulting acoustic wave fields with high-resolution data than TEC.

Realistic surface displacements, driven by a model description of the 2011 Tohoku-Oki earthquake, are used to excite AWs at the ground-level boundary of the 3D
6.2. **MESOPAUSE AIRGLOW DISTURBANCES DRIVEN BY NONLINEAR INFRASONIC ACOUSTIC WAVES GENERATED BY LARGE EARTHQUAKES**

MAGIC. Whereas the previously mentioned studies focused on ionospheric responses, the aim of the present study is to investigate and demonstrate the possibility to detect perturbations in MA, triggered by AWs at the region near the epicenter. We also insure constraint of wave field amplitudes, in comparison to prior simulations by Zettergren et al. [2017] and Zettergren and Snively [2019]. Although the earthquake happened during the day and MA observations are most feasible at night-time, the amplitudes of simulated AW perturbations in the day-time background atmosphere are similar to those in a night-time atmosphere, so they can be readily compared. This case study thus demonstrates the hypothetical observability of a Tohoku-Oki-like earthquake under night-time airglow chemistry assumptions, for the hydroxyl OH(3,1) band and OI 557.7 nm green line emission. Our results suggest that, although they may be limited to night-time scenarios, MA observations can be a valuable source of information on earthquake-induced AWs, in particular for large earthquakes. They can provide an additional high-resolution data source for seismological studies, and enable the detection of waves minutes prior to their arrival in the ionosphere.

Surface displacements were simulated with the SPECFEM3D_GLOBE 3D forward seismic wave propagation model [Komatitsch and Tromp, 2002a; Komatitsch et al., 2009] and then used to drive MAGIC at its bottom boundary, which represents the Earth’s surface. MAGIC solves numerically compressible, nonlinear Navier-Stokes equations using a shock-capturing finite volume scheme [Snively, 2013], based on the methods from Langseth and LeVeque [2000] and Bale et al. [2002] in a modified version of Clawpack 4.2 [Clawpack Development Team, 2002; LeVeque, 2002]. It simulates AGW dynamics in neutral atmospheres and includes mesospheric airglow chemistry [Snively, 2013; Snively et al., 2010; Zettergren and Snively, 2015].
6.2. MESOPAUSE AIRGLOW DISTURBANCES DRIVEN BY NONLINEAR INFRASONIC ACOUSTIC WAVES GENERATED BY LARGE EARTHQUAKES

The offshore component of the earthquake source generates compressional hydro-acoustic and tsunami waves, and both may result in a sea-surface height change. In our simulation, we consider atmospheric dynamics driven by seismic waves of periods longer than 23 sec. Also, the wavelengths \( \lambda \) of Rayleigh waves, resolved in our simulation, are not shorter than \( \sim 70-80 \) km and the ocean depth \( H \) is 1-5 km in the region considered. Ocean acoustic waves are longitudinal waves with periods of usually lower than \( \sim 10 \) sec [Nosov et al., 2015] with sets of horizontally propagating modes. In the case of \( \lambda \gg H \), seismic wave dynamics at the bottom result in practically equivalent vibrational motion of the ocean surface [Levin and Nosov, 2016b; Saito, 2019]. For \( \lambda \ll H \), the solid-liquid interface supports Scholte mode [Kessel, 1996], otherwise surface waves generated at the crust-water interface are RWs and behave as in the absence of the ocean [Kennett, 2009; Kessel, 1996].

In general, ocean compressibility has a negligible effect on the initial tsunami distribution, and it is a usual approach to consider the ocean surface height change to mimic the displacements at the ocean bottom [Kozdon and Dunham, 2014; Lotto and Dunham, 2015; Satake, 1987]. This is particularly true for large size and long duration sources, as in the case of the 2011 Tohoku-Oki earthquake [Lay, 2018]. Thus, in the simulation, we assume that permanent and transient (driven by seismic waves with periods > 23 sec) ocean bottom displacements result in the same instantaneous sea-surface height change, and use them directly as a source of AGWs in the atmosphere in MAGIC model without considering dynamics within the ocean. This relaxes the need for detailed modeling of ocean responses to the undersea earthquake, but may not be appropriate in case of the incorporation of high frequency content of seismic waves, rapid faulting processes or small fault sizes [Kajiura, 1963; Saito
6.2. MESOPAUSE AIRGLOW DISTURBANCES DRIVEN BY NONLINEAR INFRASONIC ACOUSTIC WAVES GENERATED BY LARGE EARTHQUAKES

and Furumura, 2009]. Also, we do not simulate the propagation of the tsunami and related atmospheric dynamics, concentrating our attention on MA responses to infrasonic AWs driven by vertical displacements. Vertical group velocities of internal atmospheric gravity waves generated by tsunami are \( \sim 50 \text{ m/s} \) [Occhipinti et al., 2011], that are substantially slower than the speed of sound in the atmosphere (\( \sim 280-340 \text{ m/s} \) in the considered region). Tsunamigenic gravity waves reach MA altitudes \( \sim 25-30 \text{ min} \) after the earthquake and thus are not considered in our simulation of atmospheric dynamics during 12.5 min after rupture nucleation.

Extensive observations from regional and global seismic, geodetic and ocean instrumentation networks provide a unique opportunity for the investigation of earthquake faulting mechanisms. However, the complexity and variety of models, based on different algorithms, data sets and assumptions makes it challenging to select a single source representation that comprehensively explains all phenomena of interest [Lay, 2018]. After reviewing available kinematic slip models, we have chosen one by Shao et al. [2011] (Model III), which is found to be appropriate for forward seismic wave propagation simulation, as well as for tsunami simulation [MacInnes et al., 2013; Ren et al., 2013]. Source inversion results show satisfying agreement with near-field final displacements, based on GPS measurements and tsunami observations.

The 3D forward seismic wave propagation simulation is resolved up to \( \sim 0.043 \text{ Hz} \). S362ANI and CRUST2.0 Earth interior models are used [Bassin et al., 2000; Kustowski et al., 2008]. Ocean and gravity effects are incorporated, as discussed in Komatitsch and Tromp [2002b]. The MAGIC simulation is resolved with a uniform 500 m resolution for the domain 1200x1200x200 km in meridional, zonal and vertical di-
6.2. MESOPAUSE AIRGLOW DISTURBANCES DRIVEN BY NONLINEAR INFRASONIC ACOUSTIC WAVES GENERATED BY LARGE EARTHQUAKES

reactions, respectively (note that 250 m resolution test cases found equivalent results, albeit at very high cost, confirming reasonable convergence). At the bottom boundary, we initially impose vertical velocities, and enforce a closed boundary condition after 415 s to incorporate possible AW reflections at the surface.

Figure 6.1: (a) Model configuration used for the simulation. The results of vertical fluid velocity are shown on the slide at T=540 s from rupture nucleation using cross-section slices. (b) Background temperature and wind profiles used in simulation. Comparison of vertical fluid velocities in their maxima at the altitude of ~270 km from (c) current simulation and (d) Zettergren et al. [2017] (refer to Section 4).
6.2. MESOPAUSE AIRGLOW DISTURBANCES DRIVEN BY NONLINEAR INFRASONIC ACOUSTIC WAVES GENERATED BY LARGE EARTQUAKES

Top and side boundaries of the MAGIC numerical domain are effectively open. Meridional and zonal winds, as well as temperature profiles for altitudes up to 55 km, were used from the MERRA-2 database. For higher altitudes, empirical models NRLMSISE00 [Picone et al., 2002] and HWM-14 [Drob et al., 2015] were specified (Figure 6.1,b). MERRA-2 data are selected for 15:00:00 LT, 14 minutes after rupture initiation. The background profiles remain stationary, and their change is found to be insignificant for the short simulation time frame across the spatial domain. The calculation of photochemistry for OH(3,1) and O(1S) modeling is based on the approach reported in [Snively et al., 2010] using the chemistry of [Adler-Golden, 1997] and [Makhlouf et al., 1995, 1998]. As we investigate only relative perturbations to the emitting layers, the results are insensitive to the specific rate coefficients used. The earthquake occurred during comparatively active geomagnetic time (Dst index $\sim$-61 nT). We did not include the topography/bathymetry to SPECFEM3D_GLOBE simulation, because their effect on long-period seismic wave propagation is small.

6.2.2 Model Simulation Results

For validation purposes we compare our results with simulations reported by Zettergren et al. [2017] and Zettergren and Snively [2019]. They investigated ionospheric responses to AGWs excited near the epicenter and found good agreement of simulation results with TEC observations, as well as confirming the formation of an ionospheric depletion. We compare neutral vertical fluid velocities at altitudes of $\sim$270 km (see Figure 6.1,c-d) and find good agreement between our 3D and their original 2D axisymmetric simulations. Our simulation provides more comprehensive dynamics of AWs driven by the offset near the epicenter and AWs driven by transient
6.2. MESOPAUSE AIRGLOW DISTURBANCES DRIVEN BY NONLINEAR INFRASONIC ACOUSTIC WAVES GENERATED BY LARGE EARTHQUAKES

displacements at further distances from seismic wave propagation.

Simulated perturbations of vertically integrated airglow OH(3,1) and O(^1S) emission rates and OH(3,1) brightness-weighted temperature, which are calculated as discussed in Section 3.1.3 in Snively et al. [2010] and represent the simplified observational geometry of ground-based imagers, are presented in Figure 6.8. The snapshots correspond to the time 435 sec after Tohoku-Oki earthquake rupture nucleation (defined by kinematic slip model at $T_0=14:46:23$ LT). The map of final displacements is provided for reference in Figure 6.8,a. Surface displacement dynamics developed to the east from approximate JMA epicenter position, and resulted in maximum uplift of $\sim 13$ m and subsidence of $\sim 5$ m near Japan Trench ($\sim 38^\circ$N/143.3$^\circ$E). The final displacements obtained in our simulation are comparable with results based on other finite-source rupture models [MacInnes et al., 2013]. The details of the kinematic slip model and rupture dynamics can be found in Shao et al. [2011]. Simulation results show that AWs at altitudes of 87 and 95 km (at the peaks of the OH(3,1) and O(^1S) layers), excited from near-epicentral displacements, may exhibit vertical fluid velocities from -200 to 120 m/s and from -260 to 160 m/s, respectively. Major gas temperature perturbations reach values from -50 to 35 K at altitude 87 km and from -100 to 60 K at 95 km. This points to stronger perturbations in the O(^1S) layer ($\sim 95$ km), due to its higher altitude, in comparison with perturbations in the OH(3,1) layer ($\sim 87$ km). Amplitudes of vertical fluid velocities of acoustic waves driven by RW at mesopause altitudes are in the range -40 to 30 m/s and also exhibit nonlinear effects.

The perturbations in OH(3,1) emission rate and temperature appear at $T_0+300$ sec and at $T_0+340$ sec for O(^1S) integrated emission rates. Perturbations (the departure
6.2. MESOPAUSE AIRGLOW DISTURBANCES DRIVEN BY NONLINEAR INFRASONIC ACOUSTIC WAVES GENERATED BY LARGE EARTHQUAKES

from the initial state) peak at ~ 50% and ~ 70% in OH(3,1) and O(^1S) emission rates. OH(3,1) emission rate perturbations appear earlier due to the fact that the OH(3,1) layer is situated at lower altitudes than O(^1S) (see Figure 6.4). OH(3,1) brightness-weighted temperature perturbations peak at ~ 15% and reach values -25/+37 K. The propagation of AWs is followed by the decrease in OH(3,1) and O(^1S) emission rates that is persistent over multiple oscillations from the propagation of AWs and concluded by a comparatively long-period recovery phase (depletion). The spatial extent of the depletion is ~160x250 km and resembles the shape, orientation and size of the region of largest seafloor displacements. Our simulation shows perturbations of 1-5% in OH(3,1) and O(^1S) integrated volume emission rates driven by RW AWs.

Time evolutions of simulated surface vertical velocities, OH(3,1), O(^1S) integrated emission rates and brightness-weighted temperature are shown in Figure 6.3. MA layers exhibit strong perturbations, driven by AWs from near-field displacements. MA perturbations driven by RW-excited AWs are also visible in the simulations, even at distances 600 km from epicenter, though much weaker. Apparent phase velocities of excited MA perturbations from RW AWs correspond to phase velocities of RWs at the ground and are ~4 km/s. Note, that seismic and atmospheric acoustic waves of periods > 23 sec are resolved in the simulation and, thus, airglow perturbations reveal responses to only low frequency RWs, whereas higher frequency content of acoustic waves excited by RWs may be still present at these altitudes. Though not clearly detectable on time evolution diagrams, there are also MA perturbations driven by AWs generated by body waves, but their amplitudes are negligible (likely undetectable), in comparison with near-field and RW’s driven AWs.
6.2. MESOPAUSE AIRGLOW DISTURBANCES DRIVEN BY NONLINEAR INFRASONIC ACOUSTIC WAVES GENERATED BY LARGE EARTHQUAKES

Figure 6.2: (a) Map of final vertical displacements. Vertically integrated (b) OH(3,1) and (c) O(1S) photon volume emission rate, and (d) OH(3,1) brightness-weighted temperature perturbations 435 sec after rupture nucleation. Dashed lines represent meridional and zonal slices used for keograms.

Whereas there is no marked depletion in OH(3,1) integrated emission rate, it can be seen for O(1S) after the propagation of acoustic shock waves, due to their impacts on the oxygen layer and taking into account that AW amplitudes of velocities at altitudes...
6.2. MESOPAUSE AIRGLOW DISTURBANCES DRIVEN BY NONLINEAR INFRASONIC ACOUSTIC WAVES GENERATED BY LARGE EARTHQUAKES

of O(^1S) emission layer are substantially higher than those at OH(3,1) layer, due to exponential decrease of atmospheric density with altitude [Bergmann, 1946].

Figure 6.3: (a-d) Time-Latitude diagrams of simulated vertical surface velocities (a), integrated OH(3,1) (b) and O(^1S) (d) emissions and temperature (c) perturbations along the longitude of ~ highest vertical surface displacements 143.67°E. (e-h) Time-Longitude diagrams of same perturbations as on (a-d), but along the latitude 38.32°N. T_0 represents time of rupture nucleation. Plot color scales are oversaturated for better visibility.

Figure 6.4,a-c shows the series of snapshots of OH(3,1) and O(^1S) volume emission rates for meridional slice along 143.62°E - the same as used in Figure 6.3. Calculated emission rates for both species are overlaid on the same plots. The perturbations in both emitting layers coincides with the dynamics of AW propagation. Leading AWs result in steep and substantial displacement of layers upward and outward (Figure 6.4,a) and are particularly strong in the O(^1S) layer. Further, nonlinear oscillations of MA follow downward motion of the initial layers (Figure 6.4,b), and by further up-
6.2. MESOPAUSE AIRGLOW DISTURBANCES DRIVEN BY NONLINEAR INFRASONIC ACOUSTIC WAVES GENERATED BY LARGE EARTHQUAKES

ward displacements, etc. Figure 6.4,d,e presents time-altitude diagrams for location 38.4°N/143.62°E (0 km on Figure 6.4,a-c). Partially, this depletion is driven by the advection, leading to the displacements of minor species layers and reduced emission rates. At ~400 s after rupture nucleation, a substantial “hole” in O(^1S) volume emission rate is detectable (see Figure 6.3,d,h and Figure 6.4,c), whereas for OH(3,1) the layer exhibits smaller depletion of the emission; these differences are largely explained by different amplitudes of AWs in each layer.

6.2.3 Discussion

The results shown are based on the finite-fault model of Shao et al. [2011] (Model III); we also considered the kinematic slip model by Melgar and Bock [2015], which is based on near-field data and may provide better spatial and time control of the source. The results of MA perturbations based on this model are up to ~ 36% and ~ 70% in OH(3,1) and O(^1S) integrated volume emission rates respectively and ~ 9% in OH(3,1) temperature. Though a bit weaker, the amplitudes of the perturbations based on this model support our conclusion: The nonlinear AWs generated at the near-field region during large earthquakes can drive substantial perturbations in MA, which can potentially be observable with modern imagers that are sensitive to perturbations to both emissions considered here, e.g., with sensitivity to 1% intensity and ~1 K temperature for hydroxyl [Pautet et al., 2014; Schmidt et al., 2013].

It should be noted, that the fault model by [Melgar and Bock, 2015] cannot be adapted for forward seismic waves propagation simulation in SPECFEM3D_GLOBE without extensive new development. Thus, for this simulation we considered only temporal and spatial offset dynamics without directly simulating seismic waves propagation.
that may result in lower amplitudes of AWs. The use of comparatively large subfaults in this model, which is appropriate for the main purpose of the original investigation of Melgar and Bock [2015] emphasizing tsunami formation, provides discontinuities at the edges of subfaults that drive surface deformations and subsequently AWs in atmosphere. Also, wide range of peak vertical displacements reported based on different models, may lead to under/overestimation of simulated MA responses that is not possible to validate; direct observations of offset or MA perturbations are needed to quantify precisely the amplitudes of AWs and perturbations in MA. This points to needs and opportunities for observational instrumentation deployments and further modeling studies, which may lead to important applications, e.g., the estimation of ocean bottom displacements amplitudes based on MA observations.

Astafyeva and Heki [2009] and Occhipinti et al. [2018] reported that ionospheric observation, such as high frequency Doppler, TEC and radar measurements, can be a useful seismological data source. GNSS TEC represents the observations of plasma responses to AGWs roughly at the altitudes 250-300 km, where the coupling of neutral gas with plasma is effective. The amplitudes of AGWs, affected by strong dissipation at these altitudes and marked enhancement of nonlinear effects, should also be taken into account [Heale et al., 2014; Sabatini et al., 2019b]. In addition, because the plasma in the F-region of the ionosphere moves most easily along geomagnetic field lines and experiences ambient drift, its motion is not a direct representation of AW effects (e.g., Hooke [1970]; Nishioka et al. [2013]; Zettergren and Snively [2019]); MA perturbations, driven by the neutral gas, do not suffer from this issue. For instance, for the analysis of RW propagation speeds, MA measurements may be used for the detection and analyses of RW excited acoustic waves, which are not yet affected
by strong dissipation at mesopause altitudes. Although detectable for the example shown, which represents a very large earthquake, it is important to note that the characteristics of RW AWs, and subsequently the presence of RW signatures, depend on the earthquake’s magnitude and faulting mechanism, as well as properties of the seismic waves propagation media [Udías et al., 2014]. The investigation of time and spatial faulting dynamics, based on observations of MA perturbations may also be possible for certain events and these possibilities remain to be verified through detailed observational and/or modeling studies.

Figure 6.4: (a-c) Latitude-altitude diagrams of OH(3,1) and O(1S) volume emission rates at 3 moments of time. The plots are compressed horizontally to clearly represent perturbations. (d-e) OH(3,1) and O(1S) volume emission rates time-altitude diagrams.

It is reasonable to assume that amplitudes of upper atmosphere responses are approximately correlated with amplitudes of ocean surface displacements (e.g., where initial tsunami disturbances provide a source of AWs in the atmosphere) and thus can be
used for the characterization of initial tsunami distribution. Recently, the possibility of early tsunami warning systems, based on TEC observations, have attracted considerable attention [Kamogawa et al., 2016a; Occhipinti et al., 2013; Savastano et al., 2017]. Our results suggest that MA observations may supplement such systems in particular at night-time. The Tohoku-Oki earthquake occurred in the vicinity of the Japanese coast (\( \sim 70 \) km offshore) and wide field imagers (i.e., with fields of view of 120-180°) installed on the coast or even inland, as well as imagers aimed off-zenith over the ocean, could have readily detected the simulated MA perturbations arising from AWs excited by a night-time equivalent of this earthquake. In order to demonstrate this, we synthesised images that can be obtained with such an airglow imager; the results are presented in Figure 6.5. The images of the integrated vertical emission rates (IVER) of OH(3,1) are created by integrating through the line-of-sight (LOS) of each 512 x 512 pixel through the 3D OH(3,1) airglow layer volume emission rate data. An imager is first defined from its lens profile and field-of-view (FOV) and then rotated to be aligned with north or pointing at desired tilt and azimuth angle, at any given surface height, latitude or longitude; in this case the imager is placed at the east-most point along Japan’s coast roughly at the same latitude of the strongest perturbations in OH(3,1) and it is looking into zenith direction (Figure 6.5a,) and at 40° tilt and eastward 90° azimuth (Figure 6.5b) which is pointing at the center of the data at 87 km height. The algorithm solves for the intersection of each pixel’s LOS and an ellipsoid that represents an atmospheric thin layer at each height; this ellipsoid is determined by taking the ellipsoid that represents the surface of the Earth and expanding it by an ellipsoidal height that matches the vertical resolution. The LOS is set with 1 km resolution for every pixel (to capture scales of several kilome-
6.2. MESOPAUSE AIRGLOW DISTURBANCES DRIVEN BY NONLINEAR INFRASONIC ACOUSTIC WAVES GENERATED BY LARGE EARTHQUAKES

ters) and this effectively maps a thick airglow layer on a 2D pixel surface, such as a CCD sensor. This is done for both perturbation and background data to obtain relative perturbations (Figure 6.8). The resulting synthetic image is then unwarped at the mean peak OH(3,1) airglow height (87 km) by using the same LOS-ellipsoid intersection equations and subsequently plotted on a projected map (Figure 6.5c,d). The imager was placed at the location 38.3220°N/141.5219°E.

Simulation results suggest that substantial perturbations in MA would have been clearly imaged ~6 min after rupture initiation. First arrivals of tsunami to the shore of Japan were detected ~15-20 minutes after the earthquake [Mori et al., 2011]. Thus, the time between possible detection of airglow perturbations, indicating ocean displacements, and tsunami arrival is sufficiently short to be useful for near-shore tsunami early-warning systems, in addition to other techniques and instrumentation.

According to our simulation and observations reports, clear detection of perturbations in MA can be registered ~3 min earlier than in TEC for the Tohoku-Oki earthquake, 6 min and 9-10 min after the earthquake, respectively. The synthetic MA data also include AW features with smaller scales and shorter periods than those captured by TEC, potentially revealing more information about their source processes. Also, low electron density in the ionosphere during the night may result in fairly small (less detectable) TEC perturbations, even if AWs have marked amplitudes, whereas MA observations do not suffer from this problem and instead provide a complementary capability that favors night-time detections. We want to note that the precise estimation of initial tsunami distribution may be complicated by the effective directivity of the acoustic waves radiated by the source, as well as their interference, nonlinear evolution and dissipation. In addition, strike-slip earthquakes with comparatively small
vertical surface displacements may also result in upper atmosphere perturbations comparable in amplitudes with tsunamigenic earthquakes (with large vertical surface displacements), as it was demonstrated based on TEC observations in Astafyeva et al. [2014], and this requires further investigation.

**Figure 6.5:** Synthetic images of OH(3,1) integrated volume emission rates (IVER) for (a) a zenith pointing wide field (180°) imager and (b) an eastward pointing imager with 40° tilt angle of 140° FOV. (c-d) Synthetic images unwarped on a geographic map and shown on an oversaturated scale for better visibility of weaker features. Black circles in plot (c) show observable regions for imagers with 120°, 140° and 160° FOVs, whereas a wide field imager covers the whole region. The yellow point in plots c,d represents the position of the imager.
6.2.4 Conclusions

To the best of our knowledge, our report presents the first simulations of MA emission perturbations driven by nonlinear AW shocks generated at the near-field region of an earthquake. Results support the hypotheses of Bittner et al. [2010] as to the potential value of airglow data for sensing hazard-generated AWs, and also demonstrate the importance of nonlinearity, where signatures may persist following the passage of initial AWs. No equivalent airglow data for such events have been reported to date, and we are not aware of any datasets that would have captured the nonlinear phenomena predicted here for prior earthquakes. Thus, more detailed investigations, including model simulations of case studies based on prior earthquakes, as well as new airglow observations in seismically active regions, are needed. Similar (observable) signatures are anticipated for weaker earthquakes, too, especially when large vertical displacements occur, but may be closer to the threshold of feasible detectability.

Simulation results reported here demonstrate that AWs following large earthquakes (e.g., comparable to the Tohoku-Oki earthquake) are sufficiently intense to drive large perturbations in the OH(3,1) and O(1S) mesospheric night-time airglow emissions. The airglow signatures of the nonlinear AWs would have been readily measurable from ground or space by contemporary imagers. The case reported here can be considered a night-time analog to the 2011 Tohoku-Oki earthquake; airglow data may be able to provide actionable information, if measured, with strong shock responses confirming the presence of a very large ocean surface displacement as early as ~6 minutes following a similar type of event. New targeted observations above regions at high risk of seismic hazards may thus provide an additional source of data for tsunami
6.2. MESOPAUSE AIRGLOW DISTURBANCES DRIVEN BY NONLINEAR INFRASONIC ACOUSTIC WAVES GENERATED BY LARGE EARTHQUAKES

early-warning systems, as well as new diagnostics of surface displacements (including amplitude and horizontal spatial characteristics) for seismological studies.

Three types of coseismic MA perturbations are discussed in this report and their temporal and spatial characteristics can provide guidance on what airglow imager can be appropriate for their detection. First, large-amplitude MA perturbations, up to 50% and 70% in OH(3,1) and O(\(^{1}\)S) IVER and up to 15% in OH(3,1) BWT, are generated over the focal area. Their sources are strongly nonlinear shock waves, that exhibit dominant period of \(\sim 60\) sec and vertical wavelength of 17 km at mesopause altitudes. Secondly, a quasi-permanent depletion in VER, that is particularly strong in O(\(^{1}\)S), can be observed over the focal area. As it was mentioned, the duration of MA depletion recovery depends on large-scale dynamics that are not included into the simulation. However, as large-scale gravity and planetary waves exhibit on much longer time intervals and periods, we speculate that MA depletion can be detected for at least several minutes, e.g., a significant fraction of a buoyancy period, after the arrival of shock waves to mesopause altitudes, as simulated here. The area of MA perturbations and the depleted region are \(\sim 160 \times 250\) km with a center located \(\sim 70\) km off the coast of Japan to east. Finally, MA perturbations of 1-5% in OH(3,1) and O(1S) IVER and periods of \(\sim 50\) sec are driven by RW AWs and propagate with apparent horizontal phase velocities of \(\sim 4\) km/s. Their small amplitudes and rapid passage through the field of view suggests that they could only be observed under very certain conditions and with special observing systems. In addition, RW AW wavefronts have low elevation angles (\(\sim 20-25^\circ\)) and their registration (as in the case of RW AWs observations in TEC) requires appropriate LOS directionality [Inchin et al., 2020a]. As the 2011 Tohoku-Oki earthquake was one of the strongest earthquakes in a
recent time, we caution the reader that the characteristics of signals described above are applicable only to this or similarly-scaled events.

6.3 Mesopause airglow responses to tsunamigenic acoustic-gravity waves

In this section, we present the simulation results of MA disturbances generated by TAGWs during a nighttime equivalent of the 2011 Tohoku-Oki tsunami. These results are based on the same simulation as reported in Chapter 5, in which we incorporated MA dynamics.

The snapshots of vertical ocean surface velocities and OH(3,1) and O(\(^1S\)) IVER perturbations are presented in Figure 6.6,a-c. IVER perturbations are shown as the percentage change from the unperturbed state. On the average, the amplitudes of OH(3,1) and O(\(^1S\)) IVER perturbations are \(\sim 15-20\%\) along main energy lobe of tsunami propagation to the southeast, albeit locally. For example in the region of tsunami focusing after the passage of Shatsky Rise, the IVER perturbations reach a level of \(\sim 40\%\). OH(3,1) BWT perturbations peak at \(\sim 7\%\) at the near-epicentral region and to the southeast of Shatsky Rise.

OH(3,1) BWT perturbations, along with ocean surface vertical velocities, are presented in Figure 6.6,d-i for 3 time epochs. As demonstrated earlier, strong shock acoustic waves from the focal area lead to the formation of quasi-permanent depletion in the mesopause airglow volume emission rates in both species under consideration. With time, the position of the depletion is shifted in accordance to the local
6.3. MESOPAUSE AIRGLOW RESPONSES TO TSUNAMIGENIC ACOUTIC-GRAVITY WAVES

Figure 6.6: (a-c) Simulated ocean surface vertical velocities and OH(3,1) and O(1S) IVER perturbations at 07:26:24 UT. (d-i) Simulated ocean surface vertical velocities and OH(3,1) brightness-weighted temperature perturbations for 3 instances of time. The plots are over-saturated for better visibility of weak signatures.
meridional and zonal winds. In addition, we find the formation of mesopause airglow depletion in the region of TAGW instabilities; however the amplitudes are comparatively smaller (~20-40%). The MA responses are practically linear with distance along the main lobe of the tsunami, as well as to the south and to the north.

![Field of maximum ocean surface vertical velocities](image1.png) ![Field of maximum OH(3,1) BWT perturbations](image2.png)

**Figure 6.7:** (a) Field of (a) absolute maximum vertical ocean surface velocities and (b) absolute maximum OH(3,1) BWT perturbations. The plots are oversaturated for better visibility of weak signatures.

The fields of maximum ocean surface vertical velocity and maximum OH(3,1) BWT perturbations are presented in Figure 6.7. As discussed in Chapter 5, the propagation of the tsunami is markedly affected by spatially varying bathymetry and topography. Dynamics similar to the tsunami evolution can be seen in the patterns of modulated MA perturbations. The most prominent bathymetry effect results over the region of tsunami focusing from Shatsky Rise to the southeast. First, we find the decrease of amplitudes, resulting from tsunami spreading over the Rise in the ocean surface velocities and MA perturbations. After Rise passage, focused tsunami waves cause marked increase in amplitudes of MA perturbations to the southeast. Enhancement of MA perturbation amplitudes can also be seen over Hess Rise to the east, and MPM to the southeast.
6.4 Earthquake kinematics constraint based on mesopause airglow response observations

The results clearly point to the detectability of MA perturbations after large earthquake with both satellite and ground-based imagers. The tracking of MA perturbations, although only feasible at nighttime, may help to supplement tsunami early-warning systems. In comparison with TAGW dynamics in the thermosphere that can be affected by nonlinear effects, we find that TAGWs have comparatively smaller amplitudes and thus can provide readily applicable information about their source. However, wave reflection, filtering and tunneling of TAGW can lead to challenges in the estimation of tsunami amplitudes and this must be addressed in the future.

The goal of this section is to present modeling results that demonstrate the applicability of MA observations for the investigation of rupturing processes of crustal earthquakes. As a demonstrative example, we perform a case study of the 2016 M7.8 Kaikoura earthquake. We report simulation results based on 3 finite-fault models discussed in details in Chapter 4: model 1 with PF rupture nucleation at $T_0+57$ sec and propagation from the south to north, model 2 with PF rupture nucleation at $T_0+30$ sec and propagation from south to north and model 3 with PF rupture nucleation at $T_0+30$ sec and propagation from the northwest to southeast, as proposed by Wang et al. [2018]. Here $T_0$ represents earthquake origin time.

At least 4 distinct areas of AW sources can be identified based on simulated MA responses. Figure 6.8,a-c show the results of perturbations in the OH(3,1) BWT at three time epochs based on simulation 1. First disturbances appear at $T_0+240$ sec, which
Figure 6.8: (a-c) The snapshots of the OH(3,1) BWT perturbations in percent from background for 3 moments of time from Simulation 1. (d-f) Perturbations of OH(3,1) BWT and integrated volume emission rates of the OH(3,1) and the O(1S) at $T_0+505$ sec.

is roughly consistent with the time of AW arrival from the ground to the altitudes of the OH layer (~87 km). This first group of AWs propagates to the northeast from the area of Humps fault (Figure 6.8,a) and generates MA perturbations of ~1.6% (1.95 K ptp) in the OH(3,1) BWT, 6% ptp in the OH(3,1) and 3% ptp in the O(1S) IVER. IVER is calculated by vertically integrating volume emission rates for each point of
the numerical domain with 90° elevation angle of LOS. Then, the group of AWs of lower amplitudes drives MA perturbations of ~1.2% (2.36 K ptp) OH(3,1) BWT, 5% ptp in OH(3,1) and 3.4% ptp in O(1S) IVER and their pattern is rotated slightly to east, pointing to Hundalee and Point Kean fault displacements as AW sources (Figure 6.8,b). These groups of AWs indicate the direction of rupture propagation, roughly from the southwest to the northeast. The third group of MA perturbations is modulated to the northeast and reproduce the rupturing at Kekerengu and Needles faults (Figure 6.8,c). The last group of AWs drive the strongest MA perturbations (~3.2% or 3.7 K ptp OH(3,1) BWT, 14% ptp OH(3,1) and 12% ptp O(1S) IVER), above the PF and the southern part of Kekerengu fault (Figure 6.8,c), where maximum vertical surface displacements are simulated. Quasi-permanent depletions in the OH(3,1) and the O(1S) IVER are generated. It is difficult to discern perturbations from the PF and the southern segment of Kekerengu fault, as they produce vertical displacements close in time and space. We present the OH(3,1) BWT, OH(3,1) and O(1S) IVER perturbations at $T_0+505$ sec in Figure 6.8,d-f. Even later in time, the southern (around Humps fault) and northern areas (around the PF and Kekerengu fault) of AW generation can be distinguished.

The pattern of MA perturbations reproduces the evolution of surface vertical displacements, producing weaker or stronger perturbations accordingly. Figure 6.9 represents time-distance diagrams of vertical surface velocities and the OH(3,1) and the O(1S) IVER perturbations for meridional (along 173.31°S) and zonal (along 42.06°E) slices. MA perturbations with much faster apparent horizontal phase velocities of ~3.5 km/s far from the focal area, are driven by AWs from seismic wave. However, it can be difficult to detect these perturbations with modern airglow imagers due to
6.4. EARTHQUAKE KINEMATICS CONSTRAINT BASED ON MESOPAUSE AIRGLOW RESPONSE OBSERVATIONS

Figure 6.9: Distance-time diagrams of (a,b) surface vertical velocities, (c,d) the OH(3,1) and (e,f) the O(1S) IVER perturbations for meridional (top) and zonal (bottom) slices from Simulation 1. The plots are oversaturated for better visibility of weak signatures.

their small amplitudes (<1% OH(3,1) IVER). After the propagation of shock waves, the depletion in O(1S) IVER of ~12–14% is formed.

Our modeling results suggest that the time of rupture nucleation on the PF can be identified based on MA observation. We present the snapshots of MA perturbations from 3 simulations mentioned above at $T_0+300$ sec for the OH(3,1) IVER and BWT and $T_0+315$ sec for the O(1S) in Figure 6.10. At these time epochs, MA perturbations in simulation 1 are driven only by AWs from Humps fault, whereas MA perturbations driven by AWs from the PF already appear to the northeast in simulations 2 and 3 (Figure 6.10,e,h).
Figure 6.10: The snapshots of perturbations in \( \text{OH}(3,1) \) BTW and \( \text{OH}(3,1) \) IVER at \( T_0 + 300 \) sec and \( \text{O}^{(1)}\text{S} \) IVER at \( T_0 + 340 \) sec from 3 simulations discussed in text.

From simulation 2, where the PF rupture propagation is specified from the south to north, the pattern of MA perturbations from the PF is focused from the south-
6.4. EARTHQUAKE KINEMATICS CONSTRAINT BASED ON MESOPAUSE AIRGLOW RESPONSE OBSERVATIONS

west to northeast (Figure 6.11,d-f). In case of simulation 3 with the PF rupture propagation specification from the northwest to southeast, the pattern of MA perturbations is focused from the PF from the northeast to south-southwest (Figure 6.11,g-i). The direction of generated MA perturbations does not fully reproduce the direction of rupture propagation at the PF. This is connected with the spatial elongation of resulted surface vertical displacements at the PF in the southwest-northeast direction, defining similar focusing of generated AWs. Thus, the direction of rupture propagation on the PF can be identified with certain accuracy, in particular defining whether rupture propagates from the south or from the north. The amplitudes of MA perturbations from simulation 2 are comparable with those in simulation 1, though producing smaller depletion (3.27 K ptp OH(3,1) BTW and up to 10.31% ptp OH(3,1) and 9.67% ptp O(1S) IVER). The strongest depletion in O(1S) IVER of 18.35% is generated in simulation 3, which is connected with constructive interference of AWs from Humps, Hundalee and Point Kean faults and the PF. The OH(3,1) IVER and BWT from simulation 3 are up to 11.49% and 3.7 K ptp, respectively.

6.4.1 Conclusions

Simulation results show that nighttime MA observations can provide valuable information and improve the understanding of earthquake-atmosphere processes. They can be used for the investigation of rupture propagation and timing of AW excitation and could potentially provide additional constraints on earthquake kinematics of the Kaikoura earthquake or similar crustal earthquake.
6.4. EARTHQUAKE KINEMATICS CONSTRAINT BASED ON MESOPAUSE AIRGLOW RESPONSE OBSERVATIONS

Figure 6.11: The snapshots of perturbations in the OH(3,1) BTW and the OH(3,1) and O(^1S) IVER from 3 simulations discussed in text. Time epochs of snapshots are indicated in each panel.
Chapter 7

Conclusions and future work

7.1 Main outcomes

This thesis presents four distinct pieces of work as laid out in chapters 3-6. We developed and validated the modeling approach for the simulation of coseismic and tsunamigenic acoustic and gravity waves and the resulting disturbances in mesopause airglow and ionospheric plasma. Chapters 3 and 4 are devoted to the case studies of two inland earthquakes - the 2015 M7.8 Gorkha earthquake in Nepal and the 2016 M7.8 Kaikoura earthquake in New Zealand. Chapter 5 includes the results of the investigation of tsunamigenic acoustic-gravity wave dynamics, based on a case study of the 2011 Tohoku-oki tsunami and generalized parametric studies with simplified tsunami sources and bathymetry. In Chapter 6, we report the results of mesopause airglow responses to coseismic and tsunamigenic acoustic and gravity waves. Finally, in conclusion, we use the subsequent sections in this chapter to address the questions
that we posed at the beginning of this dissertation in Chapter 1. Anticipated and proposed future works are summarized at the end of this Chapter.

**Excitation and propagation of acoustic and gravity waves generated by crust and ocean surface displacements from earthquakes and tsunamis**

For the simulation of AGW excitation from the earthquakes’ near-epicentral regions, and resulted airglow and plasma disturbances, finite-fault models should be considered, in particular for complex and/or large earthquakes. Point or axisymmetric sources may lead to marked biases in spatial and temporal dynamics of AWs that, in turn, can cause the misinterpretation of the physics underlying earthquake-atmosphere-ionosphere processes. The regime of propagation of AWs, driven by large earthquakes, can be weakly to strongly nonlinear, leading to their substantially different evolution in comparison with linear assumptions. In this case, direct numerical simulations is the most comprehensive way to resolve AWs propagation through the whole range of altitudes and to subsequently reproduce accurate CIDs. The regime of propagation of seismic wave-driven AWs at fields far from the epicenter is linear.

The propagation of TAGWs is markedly affected by the variations in the atmospheric state and nonlinear effects arising from interactions of the waves with their environment. The latter can lead to TAGW instabilities at the thermospheric heights, followed by the excitation of secondary AGWs that span a broad range of frequencies. Bathymetry variations play a crucial role on TAGW characteristics; ocean depth changes result in the increase or decrease of TAGW amplitudes at different altitudes, as the whole TAGW packet tilts, exhibiting different dissipation altitudes. Focusing
of tsunami waves, and tsunami interactions with seamounts and islands can lead to substantial enhancements of TAGWs. Long-period AGWs, also conveyed via viscosity, that propagate ahead of the tsunami can drive detectable perturbations in the mesosphere and the thermosphere; their dissipation, as the TAGWs are subject to transience, leads to the excitation of SAGWs.

**Mesospheric airglow and ionospheric plasma responses to realistically-driven AGWs from earthquakes and tsunamis**

Rupture propagation (and its direction) plays an important role in the spatial asymmetry of observed CIDs, and can be assessed separately from observational biases imposed by the geomagnetic field. AWs trapped between the ground and the lower thermosphere, tunneling into upper layers, can drive long-lived CIDs, observable in TEC even an hour after the earthquake. The electron depletion can be observed after inland earthquakes, in the same manner as for large undersea earthquakes. However, the depletion may be insufficiently intense to be detected in spite of strong vertical displacements generated by the earthquake. AWs, driven by long-period RWs, can be strong enough in the thermosphere to drive detectable TEC perturbations.

TAGWs can drive detectable and quantifiable mesospheric nighttime airglow emission perturbations under a wide range of scenarios. Ground- and satellite-based observations of these perturbations may provide additional information on tsunami evolution and be incorporated to tsunami early-warning systems.

AWs following large earthquakes can be sufficiently intense to drive large perturbations in the OH(3,1) and O(1S) mesospheric nighttime airglow emissions that are readily detectable with modern imagers. Strongly nonlinear AWs can drive quasi-
7.1. MAIN OUTCOMES

permanent depletions in the OH(3,1) and O(^1S) VER and be detected for minutes after large earthquake over focal areas. MA perturbations are also driven by RW AWs, though their small amplitudes and rapid passage through the field of view suggest that they can only be observed under very certain conditions.

The characterization of earthquake and tsunami sources based on modeling and observations of upper atmosphere responses to AGWs

TEC observations may supplement seismological studies through the investigation of different finite-fault models and their ability to reproduce detected ionospheric perturbations. However, the investigation of AGW dynamics based on TEC observations may be hindered by low spatial and/or temporal density of observations, gradients in electron density fields (e.g., impacts of EIA) and geomagnetic field dependencies. In addition, TEC observations (as they are integrated spatially along specific line of sights) are affected by the geometry of LOS and plasma motion direction. Thus, they may be insufficiently sensitive to differences in signal amplitudes, even though AW and plasma perturbations clearly include differences. For such studies, the tracking of TEC LOS and the comparison of time of CID arrival and their amplitudes based on slant TEC observations can be performed. The recovery of information about faulting mechanisms or surface deformations based on TEC are complicated by the loss of phase information of AWs due to nonlinear effects, as well as magnetic field effects on plasma motion. In addition, such analysis can be complicated by unfavorable geometry of TEC measurements, lack of data, or the complexity of the surface dynamics during the earthquakes.

The airglow signatures of AGWs generated by large earthquakes and tsunamis are
readily measurable from ground or space by contemporary imagers. Airglow data can provide actionable information, if measured, with strong shock responses confirming the presence of a very large ocean surface displacement as early as \( \sim 6 \) minutes following the event. New targeted observations above regions at high risk of seismic hazards may thus provide an additional source of data for tsunami early-warning systems, as well as new diagnostics of surface displacements (including amplitude and spatial characteristics) for seismological studies.

\section*{7.2 Future work}

Although not presented in the thesis, some of our recent efforts address the coupling of the 3D versions of MAGIC and GEMINI models. Though practical numerical realization is still considerably computationally intensive, it will allow investigating specific cases, where the detailed information on AGW and CID dynamics and their distribution are required; for example, precise comparison of modeled and observed TEC along LOS. Two-way coupled MAGIC and GEMINI models will help to get better insights into neutral-charge particle interactions, but are likely of secondary important in the study of AGW-ionosphere coupling.

The investigation of the generation of electric currents from AGWs in the ionosphere and subsequent Earth’s magnetic disturbances is potentially important for the understanding of ionospheric responses to the sources from the ground and the troposphere. Some of our preliminary observational and modeling studies suggest that magnetic field observations may potentially be incorporated for the NH-atmosphere-ionosphere studies \cite{Zettergren and Snively, 2019}.
Future works will also be directed to the investigation of the correlation between mesopause and ionospheric airglow and ionospheric plasma density responses (and formation of depletions) and amplitudes of their sources. For example, for now we have deferred the diagnosing the duration of the mesopause airglow depletion’s recovery, since it requires a proper incorporation of realistic atmospheric dynamics driven by planetary, tidal and gravity waves, ambient turbulence and varying atmospheric states and winds, as well as more comprehensive chemistry.

As an ambitious distant goal, we intend to explore fully coupled Earth’s interior/ocean-atmosphere-ionosphere problems. This will help to investigate atmospheric responses not only to tsunamis, but also to hydroacoustic waves, providing realistic dynamics of earthquake/tsunami-atmosphere-ionosphere coupled processes.
Bibliography


Aoyama, I. T. , N. K., T. (2017), Magnetic ripples observed by swarm satellites...


Astafyeva, E., S. Shalimov, E. Olshanskaya, and P. Lognonné (2013), Ionospheric response to earthquakes of different magnitudes: Larger quakes perturb the


Catherine, J., D. U. Maheshwari, V. Gahalaut, P. Roy, P. Khan, and N. Puviarasan (2017), Ionospheric disturbances triggered by the 25 April, 2015 M7.8 Gorkha earth-


Clawpack Development Team (2002), Clawpack software, version 4.2.


Croskey, C., J. Mitchell, M. Friedrich, F. Schmidlin, and R. Goldberg (2006), In-situ electron and ion measurements and observed gravity wave effects in the polar mesosphere during the macwave program.


Ekström, G., M. Nettles, and A. Dziewoński (2012), The global CMT project 2004-


Kennett, B. (2009), Seismic wave propagation in stratified media, *ANU Press*.


Kobayashi, T., Y. Morishita, and H. Yarai (2015), Detailed crustal deformation and fault rupture of the 2015 Gorkha earthquake, Nepal, revealed from ScanSAR-based


Litchfield, N., P. Villamor, and e. a. Russell Van Dissen (2018), Surface Rupture of


Maruyama, T., and H. Shinagawa (2014), Infrasonic sounds excited by seismic waves


BIBLIOGRAPHY


Mofjeld, H. O. (2000), Analytic theory of tsunami wave scattering in the open ocean with application to the north pacific.


Occhipinti, G., L. Rolland, P. Lognonné, and S. Watada (2013), From Sumatra 2004
BIBLIOGRAPHY


Picone, J., A. E. Hedin, D. P. Drob, and A. C. Aikin (2002), Nrlmsise-00 empirical


Piñeyro, B. (2018), Nonlinear acoustic waves generated by surface disturbances and their effect on lower thermospheric composition, Master dissertation, Embry-Riddle Aeronautical University”.


BIBLIOGRAPHY


Satake, K. (1987), Inversion of tsunami waveforms for the estimation of a fault hetero-


Shao, G., X. Li, C. Ji, and T. Maeda (2011), Focal mechanism and slip history of the 2011 Mw 9.1 off the Pacific coast of Tohoku Earthquake, constrained with tele-


BIBLIOGRAPHY


Zettergren, M., J. Snively, A. Komjathy, and O. Verkhoglyadova (2017), Nonlinear iono-


