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Modeling Tsunamis for Improved Hazard Assessment and Detection

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MODELING TSUNAMIS FOR IMPROVED HAZARD ASSESSMENT AND
DETECTION

BY
MICHAEL R. SHELBY

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ABSTRACT

This body of work uses state of the art numerical models to assess and reduce tsunami hazard. The first manuscript describes the use of these models to explore nonlinear interaction between tide and tsunami in the context of hazard assessment. Inundation due to several probable maximum tsunamis (PMTs) is considered in the Hudson River Estuary (HRE). Of the sources considered, a submarine mass failure (SMF) poses the most significant tsunami threat in this region and across the entire US East Coast. The next manuscript focuses on the how SMF mechanics effect tsunami generation. In addition to inundation, SMF tsunamis are dangerous because of their short or nonexistent warning times. The final manuscript discusses developments to an algorithm which extend the range of tsunami detection by shore-based HF radar, thereby increasing warning times.

Tsunami hazard assessment in the Hudson River Estuary based on dynamic tsunami-tide simulation

The first manuscript is part of a tsunami inundation mapping activity carried out along the US East Coast (USEC) since 2010, under the auspice of the National Tsunami Hazard Mitigation program (NTHMP). Two densely built low-lying regions are situated along this coast: Chesapeake Bay and HRE. HRE is the object of this work, with specific focus on assessing tsunami hazard in Manhattan, the Hudson River and East River areas. In the NTHMP work, inundation maps are computed as envelopes of maximum surface elevation along the coast and inland, by simulating the impact of selected PMTs in the Atlantic Ocean margin and basin. At present, such simulations assume a static reference level near shore equal to the local mean high water (MHW) level. Instead we simulate maximum inundation resulting from dynamic interactions between the incident PMTs and a tide, which is calibrated to achieve MHW at its maximum level. To identify conditions leading

to maximum tsunami inundation, each PMT is simulated at four different phases of the tide and results are compared to those obtained for a static reference level in the HRE. We conclude that changes in inundation resulting from the inclusion of a dynamic tide in the specific case of the HRE, although of scientific interest, are not significant for tsunami hazard assessment and that the standard approach of specifying a static reference level equal to MHW is conservative. However, in other estuaries with similarly complex bathymetry/topography and stronger tidal currents, a simplified static approach might not be appropriate.

Modeling coastal tsunami hazard from submarine mass failures: effect of slide rheology, experimental validation, and case studies off the US East coast

We first validate two models simulating tsunami generation by deforming submarine mass failures (SMFs) against laboratory experiments for SMF made of glass beads moving down a steep slope. These are two-layer models, in which the upper layer is water, simulated with the non-hydrostatic 3D (σ -layer) non-hydrostatic model NHWAVE, and the SMF bottom layer is simulated with depth-integrated equations and represented either as a dense Newtonian fluid or a granular medium.

Using the dense fluid model, we assess model convergence with grid resolution, and sensitivity of slide motion and generated surface elevations to slide parameters. A more limited validation is conducted for the granular slide model. Both models can accurately simulate time series of surface elevations measured at 4 gages, while providing a good simulation of both the geometry and kinematics of the moving slide material.

The viscous slide model, which at present is the only one that can be applied to an arbitrary bottom bathymetry, is then used to simulate the historic Currituck SMF motion, in order to determine relevant viscous slide parameters to simulate SMF tsunamis on the east coast. The same parameters are then applied to simulate

tsunami generation from a possible SMF sited near the Hudson River Canyon.

Simulations are performed for 3 deforming slides with different dissipation parameters and the rigid slump, and results compared; all SMFs have the same initial volume, location, and geometry. Simulations of tsunami propagation are then done for the tsunamis in two levels of nested grids, using the Boussinesq model FUNWAVE-TVD, and maximum surface elevations computed along a 5 m depth contour off of the coast of New Jersey and New York.

At most nearshore locations surface elevations caused by the rigid slump are significantly larger (up to a factor of 2) than those caused by the 3 deforming slides. Hence, the rigid slump provides a conservative estimate of SMF tsunami impact in terms of maximum inundation/runup at the coast, while using a more realistic rheology with some level of SMF deformation, in general, leads to a reduced tsunami impact at the coast. This validates as conservative the tsunami hazard assessment and inundation mapping performed to date as part of NTHMP, on the basis of Currituck SMF proxies simulated as rigid slump.

Algorithms for tsunami detection by High Frequency Radar : development and case studies for tsunami impact in British Columbia, Canada

To mitigate the tsunami hazard along the shores of Vancouver Island in British Columbia (Canada), Ocean Networks Canada (ONC) has been developing a Tsunami Early Warning System (TEWS), combining instruments (seismometers, pressure sensors) deployed on the sea floor as part of their Neptune Observatory, and a shore-based High-Frequency (HF) radar. This HF radar can remotely sense ocean currents up to a 80 km range, based on the Doppler shift they cause in ocean waves at the radar Bragg frequency. Using this method, however, tsunami detection is limited to shallow water areas where they are sufficiently large due to shoaling and, hence, to the continental shelf.

To extend detection range into deep water, thereby increasing warning time, the authors have proposed a new detection algorithm based on spatial correlations of the raw radar signal at two distant locations along the same wave ray. In a previous work, they validated this algorithm for idealized tsunami wave trains propagating over a simple sea floor geometry in a direction normally incident to shore. In the final manuscript, this algorithm is extended and validated for realistic tsunami case studies conducted for seismic sources and using the bathymetry off of Vancouver Island, BC. Tsunami currents computed with a state-of-the-art long wave model are spatially averaged over the HF radar cells aligned along individual wave rays, obtained by solving geometric optic equations. A model simulating the radar backscattered signal in space and time as a function of the simulated tsunami currents is applied for the characteristics of the WERA HF radar deployed by ONC near Tofino, BC. Finally, numerical experiments show that the proposed algorithm works on realistic tsunami data. This is used to develop relevant correlation thresholds for tsunami detection.

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My advisor, Dr. Stéphan Grilli, has been an endless source of interesting and novel ideas for research. His enthusiasm has sustained a sense of excitement and purpose in the work we do. I cannot find words to express my gratitude for his gentle encouragement through frustrating moments, and for maintaining high expectations even as milestones are reached. I'm sure that the thoughtful feedback he has gifted me during our time together will shape all of my future work. I only hope I can justify the large amount of red ink spent on my behalf.

I would also like to thank the National Tsunami Hazard Mitigation Program (NTHMP) for furnishing the grants which have supported me, and Dr. Grilli for securing these funds.

My wife, Emmy, has been loving, patient and supportive through late nights and changing plans. I am grateful that she reminds me to take care of myself and encourages me to go to the beach once in a while.

Gail Paolino consistently makes even the most convoluted paperwork seem simple. With her in the office, I know that there is always someone looking out for me. It is a relief to know that she is there.

Finally, I am thankful for all of my officemates: Yong-Sung Clark, Rebekka Gieschen, Boma Kresning, Chris O'Reilly, Laruen Schambach, and Scott Hayward. They have been selflessly generous with ideas, wisdom, comfort and humor.

PREFACE

Manuscript 1 was published online in May 2016 by the Journal of Pure and Applied Geophysics (doi: 10.1007/s00024-016-1315-y). Formatting of this manuscript differs from that used for publication in order to meet university guidelines for a manuscript style thesis, but the content is identical. The manuscript references animations which are available online. This work was funded by the National Tsunami Hazard Mitigation Program (NTHMP).

Manuscript 2 will be submitted to Natural Hazards for publication. This research was funded by NTHMP and is part of ongoing collaboration with researchers at the University of Delaware. Using models developed by other authors, M. Shelby set up and ran all models relevant for this manuscript. This required minor debugging of the source code, and the development of a suite of code for data post-processing.

Manuscript 3 has been accepted for inclusion in the Proceedings of the 26th Offshore and Polar Engn. Conf. (ISOPE16, Rodos, Greece. June2016). Publication format has been adjusted to meet thesis guidelines. Listed as the second author of this paper, M. Shelby modeled the propagation of the SEMIDI tsunami and the relevant radial components of velocity; extended the radar simulator developed by Grilli and Grosdider (2016) to 2 dimensions and realistic conditions; developed the suite of code used to calculate wave rays, travel times and correlations; and finally combined these pieces programmatically to create a platform from which the detection algorithm could be tested. All figures in all manuscripts were created by M. Shelby.

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MANUSCRIPT 1

**Tsunami hazard assessment in the Hudson River Estuary based on
dynamic tsunami-tide simulations**

by

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Abstract. This work is part of a tsunami inundation mapping activity carried out along the US East Coast since 2010, under the auspice of the National Tsunami Hazard Mitigation program (NTHMP). The US East Coast features two main estuaries with significant tidal forcing, which are bordered by numerous critical facilities (power plants, major harbors,...) as well as densely built low-level areas: Chesapeake Bay and the Hudson River Estuary (HRE). HRE is the object of this work, with specific focus on assessing tsunami hazard in Manhattan, the Hudson and East River areas. In the NTHMP work, inundation maps are computed as envelopes of maximum surface elevation along the coast and inland, by simulating the impact of selected probable maximum tsunamis (PMT) in the Atlantic ocean margin and basin. At present, such simulations assume a static reference level near shore equal to the local mean high water (MHW) level. Here, instead we simulate maximum inundation in the HRE resulting from dynamic interactions between the incident PMTs and a tide, which is calibrated to achieve MHW at its maximum level. To identify conditions leading to maximum tsunami inundation, each PMT is simulated for four different phases of the tide and results are compared to those obtained for a static reference level. We first separately simulate the tide and the three PMTs that were found to be most significant for the HRE. These are caused by: (1) a flank collapse of the Cumbre Vieja Volcano (CVV) in the Canary Islands (with a 80 km^3 volume representing the most likely extreme scenario); (2) an M9 co-seismic source in the Puerto Rico Trench (PRT); and (3) a large submarine mass failure (SMF) in the Hudson River canyon of parameters similar to the 165 km^3 historical Currituck slide (CRT), which is used as a local proxy for the maximum possible SMF. Simulations are performed with the nonlinear and dispersive long wave model FUNWAVE-TVD, in a series of nested grids of increasing resolution towards the coast, by one-way coupling. Four levels of nested

grids are used, from a 1 arc-min spherical coordinate grid in the deep ocean down to a 39 m Cartesian grid in the HRE. Bottom friction coefficients in the finer grids are calibrated for the tide to achieve the local spatially averaged MHW level at high tide in the HRE. Combined tsunami-tide simulations are then performed for four phases of the tide corresponding to each tsunami arriving at Sandy Hook (NJ): 1.5 h ahead, concurrent with, 1.5 h after, and 3 h after the local high tide. These simulations are forced along the offshore boundary of the third-level grid by linearly superposing time series of surface elevation and horizontal currents of the calibrated tide and each tsunami wave train; this is done in deep enough water for a linear superposition to be accurate. Combined tsunami-tide simulations are then performed with FUNWAVE-TVD in this and the finest nested grids. Results show that, for the 3 PMTs, depending on the tide phase, the dynamic simulations lead to no or to a slightly increased inundation in the HRE (by up to 0.15 m depending on location), and to larger currents than for the simulations over a static level; the CRT SMF proxy tsunami is the PMT leading to maximum inundation in the HRE. For all tide phases, nonlinear interactions between tide and tsunami currents modify the elevation, current, and celerity of tsunami wave trains, mostly in the shallower water areas of the HRE where bottom friction dominates, as compared to a linear superposition of wave elevations and currents. We note that, while dynamic simulations predict a slight increase in inundation, this increase may be on the same order as, or even less than sources of uncertainty in the modeling of tsunami sources, such as their initial water elevation, and in bottom friction and bathymetry used in tsunami grids. Nevertheless, results in this paper provide insight into the magnitude and spatial variability of tsunami propagation and impact in the complex inland waterways surrounding New York City, and of their modification by dynamic tidal effects. We conclude that changes

in inundation resulting from the inclusion of a dynamic tide in the specific case of the HRE, although of scientific interest, are not significant for tsunami hazard assessment and that the standard approach of specifying a static reference level equal to MHW is conservative. However, in other estuaries with similarly complex bathymetry/topography and stronger tidal currents, a simplified static approach might not be appropriate.

1.1 Introduction

Tides and tsunamis are both long waves, whose propagation can accurately be modeled by a long wave theory [1], such as linear Stokes theory in deep water or Saint Venant (a.k.a., Nonlinear Shallow Water equations; NSW) or Boussinesq equations in shallow water, depending on the relative magnitude of nonlinearity and dispersive effects. In deep water, tsunamis are not significantly affected by tides, because both the tidal range is small with respect to depth and tide-induced currents are very weak; hence, tsunami phase speed and shoaling are not significantly affected by the small change in water depth caused by tides. This also applies to shallow coastal water areas that have a simple bathymetry and a fairly straight coastline, as is the case for most of the ocean-exposed US east coast (USEC), from Florida to Massachusetts. There, while tide-induced currents may become larger and tidal range be more significant as compared to the local depth, dynamic tidal effects are still small as compared to those in tsunamis, and tsunami inundation and runup can still be accurately assessed by modeling tsunami propagation over a static high antecedent water level (typically the 10% exceedence tide or the mean high water (MHW) level). This was for instance the approach followed for performing tsunami inundation mapping in Ocean City, MD due to tsunamis caused by submarine mass failures (SMF) along the upper USEC [2].

When assuming a static increase of the mean water level (MWL) in model

simulations, both tsunami phase speed and elevation will be affected by the increased depth, yielding larger inundation further onshore. However, in coastal regions where tidal range is large and/or bathymetry is complex (e.g., creating funneling effects), tide-induced currents may become both large and significantly varying in space, leading to both stronger and more dynamic tsunami-tide interactions. In such cases, earlier work (e.g., [3]) indicates that one needs to more carefully and accurately model tide-tsunami interactions to achieve a conservative coastal hazard assessment. This requires, in particular, evaluating whether non-linear and dynamic interactions between tide and tsunami currents and elevations may lead to more hazardous conditions than with the standard maximum static level approach. Along the USEC, significant tsunami-tide interactions could occur around the mouth of a few large and complex estuaries, which are also highly populated areas having numerous critical infrastructures (such as major harbors and power plants), with prominent examples being New York, NY in the Hudson River estuary and Norfolk, VA near the mouth of the James River estuary in the Chesapeake Bay, where the largest US naval facility is located.

Since 2010, under the auspices of the US National Tsunami Hazard Mitigation Program (NTHMP; <http://nthmp.tsunami.gov/index.html>), the authors and colleagues from the University of Delaware have been developing tsunami inundation maps for the USEC (e.g., [5]) by modeling coastal tsunami hazard from the Probable Maximum Tsunamis (PMTs) in the Atlantic Ocean basin. These PMTs included (Fig. 1.1 and 1.2; see also [6, 7, 8]): (1) near-field submarine mass failures (SMFs) on or near the continental shelf break, represented along the upper east coast by four large SMFs sited at selected locations, with the characteristics of the historical 165 km³ Currituck (CRT) underwater landslide [4, 2] (thus referred to as CRT SMF proxies; details will be provided later; Fig. 1.2); (2) an extreme

Figure 1.1. Footprint of the 1 arc-min resolution grid G4 in the Atlantic Ocean basin (Table 2.1), with marked locations of the three far-field PMT sources used in NTHMP work: MTR, CVV and PRT. The “HRE” label marks the location of the Hudson River Estuary mouth. Color scale is bathymetry (< 0) and topography (> 0) in meters, from ETOPO-1 data.

hypothetical M9 seismic event occurring in the Puerto Rico Trench [9, 10]; (3) a repeat of the historical 1755 M9 earthquake occurring in the Azores convergence zone (MTR; Madera Tore Rise; [11, 11, 12]); (4) an extreme flank collapse (80 km³ and 450 km³ volume scenarios) of the Cumbre Vieja Volcano (CVV) in the Canary Islands [13, 14, 15]. To develop tsunami inundation maps, simulations were performed using the fully nonlinear and dispersive model FUNWAVE-TVD [16, 17], by one-way coupling, in a series of coarse to finer nested grids. According to the standard methodology, in these simulations, the reference level in the coastal grids was set to a high tide value (such as the Mean High Water (MHW) level). Hence, potential dynamic interactions between tide- and tsunami-induced flows were neglected.

It should be noted that in this past work Abadie *et al.* [14] and Tehranirad *et al.* [15] simulated two flank collapse scenarios for CVV, one with a 450 km³ volume and geometry similar to Ward and Days [13] original scenario and one, deemed the

Figure 1.2. Zoom-in of Fig. 1.1 on upper US East Coast region encompassing two large estuaries modeled in the NTHMP tsunami-tide interaction work (Chesapeake Bay (CB) and Hudson River Estuaries (HRE)); the figure covers the area where SMF tsunamis were simulated (from the Carolinas to Cape Cod). *Red boxes* mark locations of the historical Currituck slide site and of four areas (1-4) where large tsunamigenic SMF sources are sited [4, 2] and parametrized as Currituck SMF proxies [2]. Depth is in meters, in the color scale and bathymetric contours.

most probable extreme scenario based on a slope stability analysis, with a 80 km^3 volume. In the tsunami hazard assessment work done for NTHMP, considering that the 450 km^3 CVV scenario is believed to have a return period of more than 100,000 years, it was decided to use the smaller, but likelier, 80 km^3 scenario, since its return period may be closer to that of the other extreme tsunami scenarios considered in the inundation mapping study (PRT, MTR, CRT), which have return periods on the order of hundreds to a few thousand years. Consistent with this rationale, the 80 km^3 CVV scenario was also used in the present tsunami-tide interaction study, which was conducted as part of NTHMP work. In their detailed modeling and comparison of the farfield impact of both CVV sources on the US East coast, Tehranirad *et al.* [15] showed that for the 450 km^3 scenario very large incident waves, over 1025 m high, were simulated off of the upper East Coast shelf (on the order of those also found by Ward and Day), but then bottom friction on the shelf reduced wave elevations significantly and the coastal inundation was predicted to be on the order of 36 m along the coast. They showed that, for the 80 km^3 scenario, the coastal inundation along the USEC was approximately reduced by a factor of 3, to 12 m. Ward and Day [13] had predicted a much larger coastal impact along the USEC for their extreme CVV scenario, but their modeling approach had important differences with that of Abadie *et al.* [14] and Tehranirad *et al.* [15]. Abadie *et al.* simulated the subaerial landslide resulting from the CVV flank collapse with a multi-fluid 3D full Navier Stokes model, and then propagated the tsunami with the fully nonlinear and dispersive long wave model FUNWAVE, with the inclusion of dissipation from bottom friction and breaking. By contrast, Ward and Day [13] used a simplified slide model and a propagation model based on linear theory without dissipation; the latter model led to significantly overestimating coastal wave impact.

To date, interactions between tide and tsunami waves have only rarely been studied. Kowalik *et al.* [18] first hypothesized that for strong tidal flows significant effects due to tsunami-tide interactions should be observed in the tidal and tsunami currents. Kowalik and Proshutinsky [19] then modeled tsunami-tide interactions in Cook Inlet (Alaska), which has one of the largest tidal ranges in North America. They showed that results significantly differed from a simple linear superposition of separate simulations of tide and tsunami, and that maximum elevations depended on the tide amplitude and phase; with tsunami being intensified or damped, depending on mean basin depth, which is regulated by tides. They concluded that, in their simulations, the main effects of the tide were to change water depth, thus affecting tsunami phase speed, propagation and amplification, and dissipation by bottom friction. These, however, were site-specific conclusions and, tsunami-tide interactions effects cannot a priori be predicted without performing dynamical simulations combining tide and tsunami forcing. Zhang *et al.* [20] performed high resolution simulations of the impact of the 1964 Prince William Sound tsunami on the US Pacific Northwest coast, with and without dynamic tide effects, and evaluated tidal influence on wave elevation, velocity, and inundation. As could be expected, results showed that the tide had minimal effects near the open coast, but significantly affected both wave runup and inundation near the mouth of and within estuaries and rivers. On this basis, they concluded that dynamic tsunami-tide interactions should be considered in estuaries, as these could account for 50% of the observed runup and up to 100% of the inundation in some cases. To better understand the observed 100-km upstream propagation of the Tohoku 2011 tsunami in the Columbia River (Oregon) Yeh *et al.* [21] and Tolkova [3] modeled tsunami-tide interactions. Tolkova [3] found that tsunami waves propagated further on a rising tide in the lower portion of the river; however, upstream the tsunami propagated

further at the maximum high tide. Results also showed a potential amplification of tsunami waves directly after high tide. Tolkova [3] concluded that the interaction of the two long waves is completely dependent on the specific environment in which the interaction occurs, which justifies performing site-specific studies. More recently, performing similar studies based on data from a river in Japan, Tolkova *et al.* [22] showed that the Tohoku 2011 tsunami had caused increased surface elevations in the river by hindering drainage; this translated into increased tsunami inundation during tidal ebb. In the same geographic area, Nakada *et al.* [23] performed high-resolution simulations of the propagation in Osaka Bay of a large tsunami generated in the Nankai Trough. To quantify tide effects they run many cases in which tsunami propagation started every hour, through two tidal cycles. They concluded that strong flood or ebb tidal currents modulated tsunami arrival by a few minutes and led to increased elevation in many situations, particularly during strong ebb flows, as compared to a static computation.

As part of the NTHMP USEC work, Tajalli-Bakshs *et al.*[24] modeled dynamic tsunami-tide interactions in Chesapeake Bay, with particular focus on assessing tsunami hazard in the James River, which is most affected by tidal currents and has the Norfolk Naval facility at its mouth and a nuclear power plant upstream. They considered the M2 tidal component in the Bay and combined it, for different phases, with the two worst case scenario PMTs identified for this area, i.e., tsunamis generated by an extreme CVV flank collapse and the historical Currituck underwater slide, whose site is located near the mouth of the Bay [25, 2] (Fig. 1.2). While results showed clear nonlinear tsunami-tide interactions, affecting both tsunami elevation and propagation speed in the river, maximum tsunami inundation did not exceed that computed over a static reference level equal to the maximum elevation of the dynamic tide at the river mouth (here, the 10%

exceedence maximum tide level).

Earlier studies summarized above all concluded that tsunami-tide interaction effects are largely site-specific. In the Chesapeake Bay, one of the two large estuaries located along the USEC considered in NHTMP work, Tajalli-Bakhsh *et al.* [24] concluded that this more advanced modeling approach was not necessary for a proper tsunami hazard assessment. Here, following a similar methodology, we simulate the combined effects of tides of various phases on the evolution of tsunami waves in the HRE (Fig. 1.1), to compute maximum inundation elevation and limits. (Note that for simplicity we use HRE to refer to the the New York Bay tidal system, including the Hudson River Estuary and East River.) Based on the earlier work on Atlantic tsunami source modeling summarized above, [8], the three PMTs selected to perform inundation mapping in the HRE area for NTHMP, representing the most likely extreme events that can potentially affect this region of the USEC, are: (1) A Currituck SMF proxy sited on the continental slope off of the Hudson River canyon [2] (see “Study Area 1” in Fig. 1.2); (2) A 80 km³ flank collapse of the Cumbre Vieja Volcano in the Canary Islands [14, 15]; and (3) A magnitude 9.0 earthquake in the Puerto Rico Trench [9, 10]. The HRE has particularly strong currents (1-2 kts, i.e., nearly twice the speed of currents in Chesapeake Bay) and also has been identified as one of the highest risk areas along the USEC for flooding caused by a tsunami resulting from a submarine mass failure (SMF) occurring in the Hudson River Canyon [4]; this led Grilli et al [2] to define CRT SMF proxy sources in the HRE canyon area (Fig. 1.2).

Besides being part of the NTHMP work scope of performing conservative tsunami hazard assessment for all the U.S. coastal areas, the HRE is another complex tidal system to assess the importance of nonlinear exchanges of energy between tide and tsunami, similar to the work done by [3] in the Columbia River.

There, Tolkova [3] found that tsunami signals propagating with the low tide were gradually damped out, while those traveling with the high tide were preserved or amplified. This was most apparent at the farthest upstream station for which data for the Tohoku 2011 tsunami were collected. Similar phenomena were observed by Tajalli-Bakhsh et al. [24] for tsunamis propagating up the James River, although as indicated this did not lead to higher inundation than for a static tide level. If the Hudson River results are consistent with Tolkova’s [3] findings, differences between static and dynamic tsunami-tide simulations should be larger at upstream locations when propagating over a high tide.

In the following, to assess dynamic tsunami-tide interactions in the HRE, we perform two sets of simulations. First, for each PMT, we simulate tsunami propagation into the HRE assuming a static tide level equal to the local MHW level. Then we perform joint tsunami-tide simulations for four phases of tidal forcing achieving a maximum level identical to MHW in the HRE. The methodology for performing combined tsunami-tide simulations, which is similar to that used by Tajalli-Bakhsh [24], is detailed in the next section. We then briefly detail the computational model and present grid set-up. We finally report in detail and compare results of the two sets of simulations. Note that the choice of the MHW for the static and maximum tide levels is consistent with the standard approach in tsunami inundation mapping done for NTHMP. Tajalli-Bakhsh [24] used slightly higher 10% exceedance tide level in Chesapeake Bay, for both static and dynamic simulations, because tsunami hazard was assessed at a nuclear power plant in the James River, which was required to be slightly more conservative.

Grid/ Type	SW Lat. (N deg.)	NE Lat. (N deg.)	SW Lon. (W deg.)	NE Lon. (W deg.)	Res.	N_x	N_y
G4/S	10	45	82	5	1 min	4,620	2,100
G3a/C	36.396	41.885 N	74.994	69.25	616 m	788	990
G3b/C	39.171	41.904	74.829	71.138	616 m	512	489
G2/C	40.003	41.355	74.437	72.266	154 m	1,188	980
G1/C	39.171	40.900	74.829	73.775	38.5 m	459	1,504

Table 1.1. Parameters of grids (Figs. 1.1 and 1.3) used in FUNWAVE-TVD model to compute the propagation of far-field (G4, G3b, G2, G1; CVV and PRT) and near-field (G3b, G2, G1; CRT SMF proxy 1; Fig. 1.2)) tsunami sources, and tides (G2, G1). “Res.” is resolution of Spherical (S) or Cartesian (C) type grids and N_x and N_y indicate the number of grid cells in each direction.

Figure 1.3. Regional and near-shore computational grids used in tide, tsunami-only, and tsunami-tide simulations with FUNWAVE-TVD (labeled *red boxes* correspond to grids defined in Table 2.1). Tide-only simulations are initiated in grid 3b, and then nested into grids 2 and 1. After being initiated in the 1 arc-min grid G4 (Fig. 1.1), simulations of far-field tsunami sources (CVV, PRT) are carried out in nested grids G3b, G2 and G1. Simulations of the near-field CRT proxy SMF tsunamis are performed in grids G3a, G2, and G1. All tsunami-tide simulations are initiated in grid G2 and then rerun in grid G1. Color scale is bathymetry (<0) and topography (>0) in meters.

1.2 Modeling methodology and model grids

1.2.1 Models and modeling methodology

All simulations, both tide and tsunami, alone or combined, are performed using the fully nonlinear and dispersive Boussinesq model FUNWAVE [26, 27, 9, 28], in its most recent Cartesian [16] and spherical [17] implementations referred to as FUNWAVE-TVD (the spherical implementation including Coriolis effects). FUNWAVE-TVD is fully parallelized for an efficient solution on shared memory clusters and uses an efficient total variation diminishing (TVD) algorithm to follow breaking wave fronts in shallow water. The model has a quadratic bottom friction term controlled by a Manning friction coefficient n and, unlike the original FUNWAVE, it models dissipation in breaking waves by turning off dispersive terms in areas where breaking is detected based on a breaking index criterion (see details in [16], [16]). While FUNWAVE-TVD's Cartesian implementation is fully nonlinear, its spherical implementation is only weakly nonlinear; hence, it is only used in areas where tsunami elevation over local depth is on the order of 10% or less. Therefore, in tsunami simulations, spherical grids are typically fairly coarse and used to model large ocean areas in relatively deeper waters, whereas Cartesian grids have a higher resolution and are used to model coastal tsunami impact. This approach was successfully used to model the Tohoku 2011 tsunami in both the near- and far-field [29, 17]. Both implementations of FUNWAVE-TVD have been fully validated against standard benchmarks as part of the NTHMP work [30, 31].

Simulations with FUNWAVE-TVD are performed in several levels of overlapping nested grids, using a one-way coupling methodology. This works by computing time series of free surface elevations and currents in a coarser grid level, for a large number of numerical gages (stations) defined along the boundary of the finer grid level. Computations in the finer nested grid level are then performed using these time series as boundary conditions. With this approach, reflected waves propagat-

ing from inside the area covered by each finer grid are included in the time series computed in the coarser grids along the finer grid boundaries, thus satisfying an open boundary condition. For far-field tsunami simulations, to reduce reflection in the first coarsest grid level (here the 1 arc-min Atlantic Ocean basin grid G4 used to compute the transoceanic propagation of the CVV and PRT sources; Fig. 1.1), 200-km-thick sponge (absorbing) layers are specified along all the open boundaries. For the near-field CRT SMF proxy tsunami, the first level grid is initialized with the surface elevation and horizontal velocity computed using the three-dimensional model NHWAVE [32] (see details in [2])

For the dynamic tsunami-tide simulations, we follow the methodology that was first applied by Tajalli-Bakhsh[24] in the Chesapeake Bay estuary, i.e., we:

1. Simulate the propagation of the selected PMTs from their source, in a series of nested grids, to a moderate resolution regional grid (here the 154 m resolution grid G2; Fig. 1.3; Table 2.1) encompassing the HRE (Fig. 1.3).
2. Simulate and calibrate the tide for its maximum elevation to reach the local MHW level in the HRE area, based on reference results available at a series of NOAA tide gages in the estuary; bottom friction values are adjusted in the model, if necessary, to achieve a better agreement (calibration phase). The uncalibrated tidal forcing (both surface elevation and current) is obtained from a separate global model (detailed later) and specified along the boundary of the medium size regional grid encompassing the HRE (here the 616 m resolution grid G3b; Fig. 1.3; Table 2.1).
3. Then jointly simulate tide and tsunami, by linearly superimposing incoming tsunami wave elevations and velocities with tidal forcing, along the offshore boundary of a computational grid selected with a depth large enough along its offshore boundary to justify such a linear superposition (here, grid G2).

Figure 1.4. High-resolution bathymetry/topography in HRE's area of interest, from FEMA's 8 m DEM [33], used to define the finest resolution grid G1's depth matrix. This data set was combined with the 90 m NOAA DEM data to define Grid G2's depth matrix (Table 2.1). Color scale is bathymetry (< 0) and topography (> 0) in meters, referenced to the NAVD88 vertical datum.

4. Finally, simulate effects of tide phase on the three incident tsunamis by considering four different phases when peak tsunami and time-shifted tide signals are superimposed along the boundary of grid G2.

1.2.2 Grid bathymetric data

Besides its footprint, resolution, and type (spherical or Cartesian), each model grid requires a depth matrix that is developed by interpolating bathymetric and topographic data of resolution commensurate with that of the grid. The key parameters for each model grid are listed in Table 2.1 and their footprints are shown in Figs. 1.1 and 1.3. In earlier NTHMP work, bathymetry and topography for such grids were interpolated from the most accurate sources available, i.e., the 1 arc-min ETOPO-1 data in deeper water, 1 or 3 arc-s (30 or 90 m) NOAA Coastal Relief model data [34] over the shelf, and 1/3 arc-s (10 m) NTHMP or Federal Emer-

gency Management Agency (FEMA) Region II Digital Elevation Models (DEMs) wherever available [33].

In analyzing NOAA’s detailed bathymetric data used for coastal hazard assessment in the HRE, we noted a paucity or even a lack of data in the vicinity of Manhattan Island. This region, however, is critical for considering tsunami effects in the Hudson River. Hence, in this region, nearshore bathymetry was obtained from FEMA, with a resolution of about 8 m (Fig. 1.4; [33]). Thus, grid G4’s depth matrix is based on ETOPO-1 data, while that of grids G3a,b is based on ETOPO-1 and 90m DEM data. All of grid G1 and part of grid G2’s depth matrix are based on FEMA’s 8 m DEM and parts of Grid G2 that are not included in this high-resolution data were completed using NOAA’s 90 m DEM. Figure 1.5 shows the resulting (interpolated) bathymetry and topography for grids G1 and G2. The vertical datum is referenced in all grids to NAVD88. Note that grid G1 is oriented at 18° clockwise from north (Fig. 1.3) to allow for a more efficient use of grid points, which significantly reduces the model computational time.

1.2.3 Fresh water discharge

Finally, the fresh water discharge from the Hudson River was estimated at $600 \text{ m}^3/\text{s}$ [35], compared to a maximum tidal volume flux through the mouth of the Hudson River during a MHW tide at Manhattan Island of over $6000 \text{ m}^3/\text{s}$. The latter is based on currents and surface elevations computed at the mouth of the Hudson River using FUNWAVE-TVD, for tide only simulations (see details in next section). The transect where the tidal flux calculation is made is marked in Figure 1.7. Because the river discharge is small as compared to the tidal flux and to isolate tidal effects, river discharge and related current are neglected in this study.

Figure 1.5. Interpolated bathymetry and topography used in grids G1 (*red box*) and G2 (footprint of the figure; Table 2.1), encompassing the HRE, where dynamic tsunami-tide simulations are performed and compared to simulations over a static tide level. Color scale and black contours are bathymetry (< 0) and topography (> 0) in meters, referenced to the NAVD88 vertical datum. Note the deep Hudson River canyon offshore of the HRE mouth. The *yellow bullet* marks the location of a numerical gage placed at the entrance to Lower Bay (-73.944 Lon. E., 40.501 Lat. N, local depth 16.9 m), where time series of surface elevations are computed in Fig. 1.13 for the 3 incident tsunamis.

1.3 Tide only simulations

Tide propagation is first simulated in the HRE to identify/calibrate conditions causing a maximum tidal elevation equal to the local MHW level at selected NOAA tide gages. A 24-h time interval was first identified, between 7:00 am on July 13, 2015 and 7:00 am on July 14, 2015, during which maximum tidal elevations nearly reached MHW at the tide gages (NOAA’s tide gage data listed in Table 1.2 indicates that MHW varies between 0.57 and 1.19 m NAVD88 in the HRE). Tide propagation was then simulated during this time interval with FUNWAVE-TVD, in the 616 m resolution grid G3b, based on boundary and initial conditions (surface elevation and horizontal velocity) obtained from a large-scale barotropic tide model: the “Oregon Tide Prediction Software” (OTPS). In the grid G3b simulations, tidal forcing was computed along the boundary of the 154 m resolution nested grid G2 (Fig. 1.3), in the form of time series of free surface elevations and currents at many control stations, following the one-way coupling method detailed before. This procedure was finally repeated for grid G1. Based on differences observed between modeled and reference surface elevations at 14 NOAA gages located within grid G1, simulations were repeated with modified bottom friction coefficients in grid G1, to achieve the best possible agreement.

OTPS’ latest version TPXO8 predicts tidal elevations and currents along the USEC, in a 2 arc-min grid. Considering this is a fairly coarse grid, OTPS’ results are more accurate offshore, in deeper water [36]. Accordingly, tide simulations with FUNWAVE were initiated in the larger, coarser resolution, domain G3b, whose boundary is mostly located in fairly deep water. Following Tajalli-Bakhsh et al.’s approach ([24]), boundary conditions were ramped-up using a “tanh” function, from zero to the OTPS’ model predictions, over nearly a half-semidiurnal tidal cycle (6 h). Model results were allowed to stabilize for another 12 hours before

being computed and specified along the boundary of grid G2, and so forth in grid G1. To validate and calibrate tide simulations, surface elevations were computed at the locations of twenty NOAA tide gages in the HRE area (Table 1.2; Fig. 1.6), which includes 2 actual tide gages, at Sandy Hook and Battery Point, and 18 virtual tide gages where corrections are made by NOAA with respect to the actual gages, based on a harmonic analysis. At some of these virtual gages, referred to as subordinate stations (numbered #2 to #7 and #13 to #16 in Fig. 1.6), only maximum and minimum tide levels and their time of occurrence are provided; at the other virtual stations, full time series are provided. Figure 1.6 shows that all 20 stations are located within grid G2, but only 14 stations are located within grid G1 (Fig. 1.7a). Numerical results obtained for the maximum surface elevation η_m during the second tidal cycle were compared to the reference maximum for each gage η_p , which we verified was close to the local MHW level for the selected time interval.

After this initial tide simulation, the Manning friction coefficient n was adjusted in grid G1 to improve the agreement between the modeled and known maximum elevations at NOAA's tide gages. By observing discrepancies at these stations (marked in Fig. 1.7a), n was adjusted to 0.015 in the Hudson and East Rivers, north of Battery Point, while a value $n = 0.025$ was used in the remainder of grid G1; this value was also used in grid G2. The corresponding friction coefficient, $C_d = gn^2/h^{1/3}$ is plotted in Fig. 1.7d for grid G1 (with g the gravitational acceleration and h , here, being the local depth with respect to NAVD88, i.e., not including surface elevation); C_d values are seen to vary between 0.001 and 0.005, i.e., they are both lower and higher than $C_d = 0.0025$, the standard value for coarse sand. The fairly straight Hudson River has a hard muddy bottom and a regular cross-section, supporting the use of a lower n value. Since both tide and tsunami

are long waves causing significant flow velocity near the bottom, it is reasonable to use the same bottom friction values in simulations of tsunami only propagation into the HRE, as well as in tsunami-tide simulations.

Following the calibration of bottom friction, the resulting maximum surface elevations modeled at the NOAA tide gages, and their absolute and relative differences with NOAA's reference values at the 20 stations in grids G1 and G2, are listed in Table 1.2 for each station. Additionally, Fig. 1.8 shows a comparison of time series of surface elevations modeled at the stations with NOAA's reference data (either full time series or only extrema, whichever are available). The agreement between these appears visually quite good, and more so for results in the higher resolution grid G1, particularly at gages #5 to #11 and #13 to #16, which are located in the most important areas considered here: New York harbor, and the Hudson and East Rivers around Manhattan. Table 1.2 shows that the modeled maximum tide elevations in grid G1 are within 0.02 m of NOAA's reference data at 8 of these 11 stations (with a 2.2% RMS for their relative difference); two stations (#7 and #11) have differences of 0.05 and 0.06 m and the largest difference (0.10 m) is observed at the Williamsburg Bridge station. For these 14 stations, the RMS of the relative difference between modeled and predicted results is 6.5%. Hence, the overall agreement of model results with NOAA's reference data in grid G1 appears to be good, particularly in the area where we will analyze dynamic effects of tides on tsunami inundation and runup. This is further detailed in Fig. 1.7b, which shows the envelope of maximum tidal elevations computed in grid G1 during the second tidal cycle (after model ramp-up); there is only a small variation in the maximum tide elevation (less than 0.08 m) from the mouth of the HRE to New York harbor and the East River. As should be expected, maximum elevation gradually decreases in the Hudson River, from Battery Point towards upstream, and

maximum elevations are larger in Long Island Sound due to funneling effects. The average maximum elevation computed in grid G1 is +0.72 m NAVD88; looking at Fig. 1.7b, we see that this level is achieved within ± 0.02 m in most of grid G1 (excluding Long Island Sound, the western part of the Lower Bay, and the upper East River). For comparison, Fig. 1.7c shows the local MHW level (referenced to NAVD88) computed over grid G1 using NOAA’s tool VDatum, which provides an empirical estimate based on values at reference stations and bathymetry. The pattern of VDatum’s MHW values appears to be very similar to that in simulations (Fig. 1.7b), with values however being slightly smaller at most locations; the average of VDatum data is +0.64 NAVD88 (with a standard deviation of 0.02 m), i.e., 0.08 m below that of simulations. This difference, however, is deemed small in view of the uncertainty in VDatum results and other uncertainties, and considering the good agreement of simulations with NOAA’s reference data at all the important stations in grid G1.

Results of the tide only simulations will be used to initialize FUNWAVE-TVD’s dynamic tsunami-tide simulations along the boundary of grid G2, in the form of time series of elevations and currents computed in grid G3b. Dynamic tsunami-tide simulations will be compared to tsunami only simulations performed over a static MHW level. For consistency with NOAA’s reference and VDatum data, we will set this level to +0.64 m NAVD88, although the average of maximum tide elevations computed with FUNWAVE-TVD in grid G1 is slightly larger, at +0.72 m; as indicated before, the small difference between these two levels is deemed negligible in view of other uncertainties. Hence, in the tsunami only simulations, +0.64 m will be added to the bathymetry matrix, creating a geodetic vertical static datum approximately referenced to MHW level. The technique of using a static water level corresponding to MHW in tsunami simulations is consis-

Figure 1.6. Footprint of grid G2 with marked locations of 20 NOAA tide gage stations (*numbered labels*); the red stars indicate actual tide gages at #1: Sandy Hook and #9 Battery Point, and the *black bullets* mark virtual tide gages where corrections are made with respect to the actual gages based on a harmonic analysis (see locations in Table 1.2). The *red box* marks the footprint of grid G1. Simulated and measured tide time series at the stations are plotted in Figure 1.8, and differences between these are quantified in Table 1.2. Color scale is bathymetry (< 0) and topography (> 0) in meters referenced to NAVD88 vertical datum.

tent with the methodology of [2].

1.4 Tsunami only simulations

Based on earlier work summarized in the introduction, three PMTs were selected and propagated into the HRE, as a result of: (1) a far-field M_w 9 seismic source in the Puerto Rico Trench (PRT) [9, 10]; (2) a far-field source from a 80 km³ partial collapse of the western flank of the Cumbre Vieja Volcano (CVV) in La Palma, Canary Island [14, 15] (deemed to be the likeliest extreme collapse scenario; see discussion in introduction); and (3) a near-field submarine mass failure (SMF) modeled as a Currituck (CRT) slide proxy on the continental slope,

Tide Gage Station				NOAA	Grid G2 (154 m)			Grid G1 (38.5 m)		
No.	Name	Lat. N. (Deg.)	Lon. E. (Deg.)	η_p (m)	η_m (m)	Absol. diff. (m)	Relat. diff. (%)	η_m (m)	Absol. diff. (m)	Relat. diff. (%)
#1	Sandy Hook	40.7003	-74.0135	0.71	0.78	0.07	9.42	0.73	0.02	3.24
#2	Atlantic Highlands	40.4183	-74.0200	0.72	0.78	0.06	8.58	0.74	0.02	2.56
#3	Waackaack	40.4483	-74.1433	0.72	0.82	0.11	15.25	0.78	0.07	9.37
#4	Princes Bay	40.5117	-74.2000	0.75	0.84	0.08	11.09	-	-	-
#5	Coney Island	40.5667	-73.9833	0.71	0.76	0.05	7.62	0.71	-0.00	-0.02
#6	Fort Hamilton	40.6083	-74.0350	0.73	0.80	0.07	9.14	0.71	-0.02	-3.39
#7	St. George	40.6433	-74.0733	0.66	0.78	0.11	17.19	0.72	0.05	7.79
#8	Gowanus Bay	40.6650	-74.0133	0.71	0.78	0.08	11.27	0.72	0.01	2.05
#9	Battery Point	40.4665	-74.0094	0.69	0.77	0.08	12.04	0.71	0.02	2.89
#10	Williamsburg Bridge	40.7117	-73.9683	0.59	0.72	0.13	22.73	0.69	0.10	17.44
#11	Horns Hook	40.7767	-73.9417	0.69	0.55	-0.14	-19.85	0.75	0.06	9.41
#12	Willeys Point	40.7933	-73.7817	1.16	0.98	-0.18	-15.77	-	-	-
#13	Union City	40.7667	-74.0183	0.65	0.65	0.00	0.03	0.67	0.02	2.66
#14	Edgewater	40.8133	-73.9783	0.63	0.51	-0.12	-18.73	0.63	-0.01	-0.82
#15	Spuyten Duyvil	40.8783	-73.9250	0.56	0.36	-0.20	-35.52	0.58	0.01	2.18
#16	Riverdale	40.9033	-73.9167	0.58	0.33	-0.25	-42.93	0.58	-0.01	-0.91
#17	Glen Cove	40.8633	-73.6550	1.19	0.96	-0.23	-19.62	-	-	-
#18	Long Neck Point	41.0383	-73.4800	1.08	0.90	-0.18	-16.50	-	-	-
#19	Cedar Beach	40.9650	-73.0433	0.97	0.77	-0.20	-20.95	-	-	-
#20	Northsville	40.9817	-72.6450	0.81	0.62	-0.19	-22.86	-	-	-

Table 1.2. Definition and location of NOAA Tide Gage Stations marked in Fig. 1.6. The maximum water level elevation with respect to the NAVD88 datum is given at each station for the NOAA high tide (MHW) prediction η_p , compared to surface elevation η_m modeled with FUNWAVE-TVD in the 154 m resolution grid G2 and 38.5 m resolution grid G1 (note, only 14 stations are located within this grid); the absolute ($\eta_m - \eta_p$) and relative differences between these ($(\eta_m - \eta_p)/\eta_p$) are listed for each grid.

Figure 1.7. Tide simulations in grid G1 (a - c; color scale in meters from NAVD88 datum): (a) bathymetry with marked locations of NOAA tide stations (*symbols/numbers*; see Fig. 1.8 and Table 1.2 for modeled and reference elevations) ; (b) envelope of maximum computed tidal elevations; (c) Local MHW calculated by VDatum; (d) friction coefficient C_d ; a Manning coefficient $n = 0.025$ was used throughout the domain except in the rivers, where a value $n = 0.015$ was used. The *red line* marked in (a) separates the Hudson River from Upper New York Bay and is the location of tidal flow rate calculations.

Figure 1.8. Time series of tide surface elevation (with respect to NAVD88) computed with FUNWAVE-TVD in grids G1 (*solid black*) and G2 (*dash black*) at the locations of 20 NOAA tide gage stations (Table 1.2; Figs. 1.6 and 1.7a), compared to NOAA's reference data (*solid red*; either full time series or extrema, whichever is available), for tides from 7:00 am on July 13, 2015 to 7:00 pm on July 14, 2015. The model was initialized in grid G3b with tide results computed with OTPS' model TPX08. The Manning bottom friction coefficient was calibrated in grids G1 and G2 to achieve a good agreement with reference data (Fig. 1.7d; Table 1.2 compares maximum water levels computed during the second tidal cycle with local MHW levels).

off of the Hudson River canyon [2]. The far-field tsunami sources (PRT, CVV) were specified and their propagation was first modeled in grid G4 (Fig. 1.1), in which time series of boundary conditions were computed to pursue simulations in grid G3b. The near-field tsunami source (CRT SMF proxy) was specified and its propagation first modeled in grid G3a (Fig. 1.2), in which time series of boundary conditions were computed to pursue simulations in grid G2. In view of the low resolution of bathymetric data used for grids G4 and G3a,b, as recommended by NOAA, no adjustment was made to the reference level in these grids. Simulations in grids G2 and G1 were performed based on a static reference level equal to the local average MHW level in grid G1, +0.64 m NAVD88. In the next section, simulations will be performed in these two grids, dynamically, in combination with the calibrated tide, by superposing tide and tsunami time series as a combined boundary condition along the boundary of grid G2.

Figure 1.9a shows the initial surface elevation of the M_w 9 PRT tsunami source computed with Okada’s 1985 method [37], based on 12 SIFT subfaults [10, 38], and Fig. 1.9b-d shows instantaneous surface elevations computed with FUNWAVE-TVD in the 1 arc-min resolution grid G4 (truncated here at Lon. E. -55 for more efficiency due to the tsunami directionality), after $t = 30$ min, 1h 42 min and 3h 20 min of propagation, respectively. We see that maximum tsunami elevations are quite directional south-to-north and focus on the upper USEC; this was already pointed out by [9]. After 200 min of propagation, the tsunami is entering the SE corner of grid G3b (Fig. 1.3).

Figure 1.10a,b show the surface elevation and horizontal velocity magnitude, respectively, computed by Abadie *et al.* [14] with the 3D Navier-Stokes model THETIS for the 80 km³ CVV flank collapse source, at $t = 20$ min into the event. At this time, surface elevations reach up to 30 m, with the largest values occur-

Figure 1.9. Simulations of the M_w 9 PRT seismic source with FUNWAVE-TVD, in grid G4 (truncated at Lon. E. -55; Fig. 1): (a) Initial surface elevation of tsunami source computed in lower red box with Okada’s method [37], based on 12 SIFT sub faults [10, 38]; the upper red box approximately represents the area of Fig. 1.3; (b-d) Instantaneous surface elevations computed after $t = 30$ min, $1\text{h } 42$ min and $3\text{h } 20$ min of propagation, respectively. All color scales are surface elevation in meters.

Figure 1.10. Simulations of the 80 km³ CVV flank collapse source with FUNWAVE-TVD, in grid G4 (Fig. 1.1: (a,b) Initial surface elevation (m) and horizontal velocity module (m/s), respectively, of tsunami source at $t = 20$ min into the event [14]; (c,d) Instantaneous surface elevations after $t = 4$ and 8 h of propagation into the event, respectively (*color scales* are surface elevation in meters).

Figure 1.11. Simulations of far-field tsunami sources with FUNWAVE-TVD. Instantaneous surface elevation in grid G3b (*color scale* is elevation and black contour bathymetry, both in meters), upon reaching the continental shelf, for the: (a) PRT tsunami at $t = 3, 4$ and 5h and (b) CVV tsunami at $t = 8, 9$ and 10h , since the event, from *top* to *bottom*.

Figure 1.12. Simulation with FUNWAVE-TVD of near-field Currituck (CRT) SMF proxy tsunami source located in Study Area 1 (Fig. 1.2), off of the Hudson River canyon. (a) SMF tsunami, generated with NHWAVE [32, 2] at $t = 13$ min. used to initialize FUNWAVE-TVD. Surface elevations simulated in grid G3a (*color scale* is elevation and black contour bathymetry, both in meters) at $t =$ (b) 30 min, (c) 1h 18 min and (d) 2h 8 min into the event.

ring in directions between 15 and 30 deg. south of West, as was already pointed out in other work [14, 15]. Fig. 1.10c,d shows instantaneous surface elevations computed with FUNWAVE-TVD in the 1 arc-min resolution grid G4, after $t = 4$ and 8h of propagation, respectively. These results confirm that maximum tsunami elevations are initially quite directional in a more or less WSW direction towards the Caribbean Islands and South America. Nevertheless, after 8h of propagation significant tsunami waves of 2-3 m elevation are approaching the upper USEC, entering the SE corner of grid G3b (Fig. 1.3).

Details of the subsequent propagation of the 2 far-field tsunamis, PRT and CVV, computed in the nested 616 m resolution grid G3b, are shown in Fig. 1.11a,b, as instantaneous surface elevations at $t = 3, 4$ and 5h, and 8, 9 and 10h (since the event), respectively. In each case, the first snapshot is at a time when the tsunami is starting to propagate over the continental shelf; due to refraction, the leading tsunami waves have gradually become more or less parallel to the local isobaths. Hence, the very different initial directionality of the propagation of these tsunamis (i.e., approximately from south to north for PRT and east to west for CVV) has been lost during refraction over the shelf slope. This property of long wave refraction was analyzed in more details for the CVV tsunami by Tehranirad *et al.* [15], who performed simple ray tracing analyses and showed the strong bathymetric control on nearshore tsunami propagation, causing waves to focus on or defocus away from specific areas of the coastline. In particular, here we see that in both cases incident waves refract away from the Hudson River canyon and focus on the shores of eastern New Jersey and western Long Island (NY). Partial reflection occurs along these shores and reflected waves interact later in time with other waves in the incident wave trains, creating more complex patterns offshore (see, e.g., Fig. 1.11b bottom panel).

Figure 1.12a shows the surface elevation for the near-field Currituck (CRT) SMF proxy source sited in Study Area 1 (Fig. 1.2) off of the Hudson River canyon, interpolated in the 616 m resolution grid G3a, at $t = 13$ min into the event. This source was computed using the 3D non-hydrostatic model NHWAVE [32, 2] and its elevation and corresponding horizontal velocity (not shown here) were used to initialize FUNWAVE-TVD's simulations in grid G3a. At the time of Fig. 1.12a, the tsunami caused by the SMF features a large leading depression wave (about -10 m) moving into the Hudson River Canyon, followed by a larger (15 m) elevation wave. Other waves in the wave train are propagating offshore (to the SE). Upon propagation (not shown here), these shorter dispersive waves develop an oscillatory tail of higher frequency waves. The onshore propagation of the CRT tsunami is detailed in Fig. 1.12b-d, which shows instantaneous surface elevations computed at $t = 30$ min, 1h 18 min and 2h 8 min into the event, respectively. Similar to the PRT and CVV cases, the shelf bathymetry induces a strong refraction of incoming tsunami waves, which gradually become parallel to local isobaths as they approach shore; as for the other two cases, waves are refracted away from the Hudson River Canyon and focus on shores located on either sides of it, where partial reflection occurs.

Simulations of the propagation of the 3 incident tsunamis (PRT, CVV, CRT) are pursued into the HRE with FUNWAVE-TVD, by one-way coupling, first in the 154 m resolution nested grid G2, using results of grids G3a,b as boundary conditions, and then similarly in the 39 m resolution nested grid G1 (Figs. 1.3, 1.6 and 1.7a). Figure 1.13 shows time series of surface elevations computed in grid G2 relative to the static water level for the three incident tsunamis, at a numerical gage located at the entrance of Lower Bay along the eastern boundary of grid G1 (-73.944 Lon. E., 40.501 Lat. N, local depth 16.9 m; see location in Fig. 1.5). For

comparison, similar time series are plotted at a gage located on the SE corner of grid G2, in deeper water (78 m depth; Table 2.1). At grid G2's SE corner: (1) the PRT tsunami has a leading 2 m elevation crest followed by 2 large waves (with height of about 2.3 and 1.3 m, respectively, and an 18 min period) and a tail of smaller oscillations; (2) by contrast, the CVV tsunami, which also has a leading crest with 1.6 m elevation, has a tail of more than 6 large long waves (of height 1.2 to 2.9 m and period 21 to 42 min), over which many shorter wavelength (higher frequency) waves are superimposed (with period 4 to 6 min), as a result of dispersion [15]; finally (3) due to the proximity to its source (Fig. 1.12a), the CRT SMF proxy tsunami is a large dipole wave of 6.5 min period with a leading depression of -9 m followed by a 12 m crest. After propagating over the shallow shelf from grid G2's SE corner to Lower Bay's entrance, for about 1h 45 min, Figure 1.13 shows that each tsunami wave train has significantly transformed, with wave elevations decreasing and some waves being damped out; maximum surface elevation for the PRT, CVV and CRT tsunamis have reduced to 0.8, 1.2 and 2 m, respectively. This results from the combination of energy spread out, due to wave refraction over the Hudson River Canyon bathymetry (see Figs. 1.11 and 1.12), and energy dissipation due to bottom friction. Tehranirad *et al.*[15] confirmed the significant effect of bottom friction on long wave propagation over a wide shallow shelf by comparing model results with an analytical solution.

Surface elevations for the entire propagation of the 3 tsunamis (in grids G3a,b, G2 and G1) are provided as supplementary online material, in the form of animations of model results: PRT.mp4, CVV.mp4, CRT.mp4. These show details of wave refraction, dissipation, and reflection off the coast during tsunami propagation over the shelf bathymetry (grids G3a,b), as well as tsunami propagation into the HRE and resulting flooding. (Note, for the CRT case, as can be seen in the

animation between $t = 31'19''$ and $36'20''$, to avoid using unnecessary wide sponge layers in grid G3a, the offshore moving part of the tsunami was truncated; this however does not affect the tsunami wave train propagating onshore towards the mouth of the HRE.) Animations in grid G1 are provided as side-by-side panels for both a static reference level, as discussed here, and dynamic tsunami-tide simulations, which are detailed in the next section. Besides color-coded elevations, these animations also show the instantaneous total current (i.e., tsunami plus tide if simulated dynamically) in the form of velocity vectors.

Envelopes of maximum surface elevation and inundation caused by each PMT in the HRE are plotted in Figure 1.14, based on results in grid G1. It should be noted that, to allow for a better comparison with the dynamic results, plots in the figure were computed from the results of tsunami propagation over the static MHW level (+0.64 m NAVD88; equal to the domain-averaged MHW), slightly corrected by the local difference between this averaged level and the actual space-varying MHW obtained with VDatum (shown in Figure 1.7c), which at most locations is within a few centimeters of the maximum elevation of the calibrated tide (shown in Figure 1.7b). In each case, maximum surface elevations in the HRE are found to be consistent with surface elevations computed at the entrance of Lower Bay (Fig. 1.13). For PRT, maximum coastal inundation/runup are 1 to 2 m, and for CVV these are 1.3 to 2.5 m; for CRT, surface elevations are divided into two regions: (1) outside of New York Harbor, coastal inundation/runup are 2 to 3.5 m in most areas; and (2) inside the harbor (including along the coast) maximum elevations are 1.5 to 2 m.

1.5 Dynamic tsunami-tide simulations

Simulations are repeated for the three incident PMTs in combination with the time-varying calibrated tide, which both modulates the reference water level

Figure 1.13. Time series of surface elevations computed with FUNWAVE-TVD in grid G2 (time is from the start of each respective event): (*red solid lines*) at the entrance of Lower Bay along the eastern boundary of grid G1 (-73.944 Lon. E., 40.501 Lat. N.; see location in Fig. 1.5), and (*black solid lines*) at the SE corner of grid G2 (Table 2.1), for the propagation of each incident PMT, over a +0.64 m NAVD88 static level (approximating the local MHW level): (a) PRT ; (b) CVV; and (c) CRT SMF proxy.

Figure 1.14. Envelope of maximum surface elevations (*color scale* in meter) computed with FUNWAVE-TVD for the propagation of each incident PMT in grid G1 over a +0.64 m NAVD88 static level (approximating the local MHW level), corrected by the local difference between this average level and that obtained with VDatum (shown in Figure 1.7c), to better represent the local MHW: (a) PRT (up to $t = 9$ h); (b) CVV (up to $t = 13.5$ h); and (c) CRT SMF proxy (up to $t = 6.5$ h). *Times in parenthesis* indicate the total duration of tsunami simulations from the start of each respective event.

Figure 1.15. Magnitude (*color scale* in m/s) and direction (vectors) of tidal currents computed with FUNWAVE-TVD for four phases of the calibrated tide reaching MHW level in grid G1 at its highest elevation; phases correspond to the tide arriving at Sandy Hook station #1 (Table 1.2; Fig. 1.6): (a) 1.5 h before; (b) concurrent with; (c) 1.5 h after; and (d) 3 h after high tide.

and causes a significant pre-existing as well as time-varying current. This tide was calibrated for its maximum elevation to closely match the local MHW level in the area of grid G1. To identify conditions leading to maximum inundation, we consider various phases of the tide at the time of tsunami arrival in Lower Bay.

In deeper water, offshore of the HRE, both tide and tsunamis are long waves of fairly small amplitude, as compared to depth and wavelength, which can thus be linearly combined (i.e., both elevation and current are additive) [1]. For such a superposition to be accurate closer to shore, water depth must be large enough compared to tsunami and tide elevation; here, considering that incident tsunami amplitudes are on the order of 2 m or less (Fig. 1.13) and tide amplitudes are up to 1 m (Fig. 1.8), water depth should be on the order of at least 20 m. As indicated before, incident tides computed in grid G3a and tsunamis computed in grid G3b are linearly combined at numerical gages (stations) located along the

Figure 1.16. Time series of surface elevations (in meter referenced to NAVD88) computed for the CRT SMF proxy at the Lower Bay entrance (*left*) and Battery Point (*right*) (Stations #1 and #3 in Fig. 1.17): (*blue lines*) dynamic tsunami-tide simulations; (*black lines*) tsunami simulations over a +0.64 m static MHW level; (*green lines*) linear superposition of tide and static tsunami simulations (tide elevations are shown as dashed lines for reference). Dynamic simulations were performed for the leading tsunami crest arriving at Sandy Hook station at four phases of the tide (see Fig. 1.15): (a) 1.5h before, (b) concurrent with, (c) 1.5h after, and (d) 3h after high tide.

Figure 1.17. Locations of numerical gages/stations #1 to 5 (*red stars*) in the HRE, where time series of surface elevations and currents are computed in grid G1, to perform a more detailed analysis of tsunami-tide interactions. See Figures 1.16, and 1.24 to 1.29 for results.

Figure 1.18. Differences (*color scale* in meters) between envelopes of maximum surface elevation computed in grid G1, for the PRT tsunami over dynamic and static (Fig 1.14) tide levels. (The latter are corrected to match MHW computed with VDatum) The initial tsunami crest arrives at the Sandy Hook station #1 (Table 1.2; Fig. 1.6): (a) 1.5 h before; (b) concurrent with; and (c) 1.5 h after high tide.

Figure 1.19. Same as Fig. 1.18 for the maximum crest of the CVV tsunami.

Figure 1.20. Same as Fig. 1.18 for the leading crest of the CVV tsunami.

Figure 1.21. Same as Fig. 1.18 for the CRT SMF proxy tsunami.

Figure 1.22. Envelope of maximum surface elevations (*color scale* in meters) computed for the propagation of each incident PMT into grid G1, over a dynamic tide (approximating the local MHW level at its highest elevation), over the four tested tide phases (arriving at Sandy Hook station #1 (Table 1.2; Fig. 1.6) 1.5 h before, concurrent with; 1.5 h after, and 3 h after high tide): (a) PRT (up to $t = 9$ h); (b) CVV (up to $t = 13.5$ h); and (c) CRT SMF proxy (up to $t = 6.5$ h). *Times in parenthesis* indicate the total time of tsunami simulations since the start of each respective event.

Figure 1.23. Difference of maximum surface elevation (*color scale* in meters) over grid G1 for the: (a) PRT; (b) CVV; and (c) CRT tsunamis, between the envelope of dynamic tide computations (envelope of envelopes for the four tested tide phases; Fig. 1.22) and of static tide computations (Fig. 1.14; the latter are corrected to match the local MHW computed with VDatum).

offshore boundary of grid G2. Fig. 1.5 shows that water depth is greater than 30 m at these gages, except for a few located on the northernmost part of the grid eastern boundary in Long Island, and the easternmost part of the grid southern boundary in New Jersey. Since these shallower areas are fairly small parts of the grid boundary, located far away from the entrance to the HRE (Fig. 1.5), potential nonlinear effects caused by a linear superposition of tide and tsunami signals are deemed negligible for simulations in the HRE. Along the offshore boundary of grid G2, tsunami-induced currents for the 3 PMTs are found to be 3-20 times larger than the maximum tidal currents, which supports their linear superposition. For each considered tsunami-tide combination, once computations are completed in grid G2, data is passed into grid G1 where the simulation is rerun, as was done for the tsunami simulations over a static water level. All simulations were performed using the same Manning friction coefficients as in the tide and tsunami only simulations.

Linear tsunami-tide combinations are specified along the boundary of grid G2 for 4 phases of the tidal signal, i.e., a minimum of 4 simulations are performed for each incident tsunami in order to identify the combination of tidal elevations and currents that best enhances the incident tsunamis and causes the maximum combined tsunami-tide elevations and coastal inundation in grid G1. These phases were selected such that the leading and/or maximum crest of each tsunami arrived at Sandy Hook, NJ (Station #1 in Figure 1.7): (1) 1.5 hours before; (2) concurrent with; (3) 1.5 hours after; and (4) 3 hours after high tide. The 1.5 h time interval between each combination roughly represents one-eighth of the dominant tidal period. For the PRT and CRT tsunamis, tsunami-tide synchronization was done for the leading crest (Fig. 1.13). For the CVV tsunami, a second taller crest arrived just over two hours after the initial crest (Fig. 1.13); accordingly, besides 4 simulations for the leading crest, 4 additional dynamic simulations were performed

for CVV, corresponding to the arrival of this second crest at the 4 phases of the tide.

Figure 1.15 shows the magnitude and direction of tidal currents computed in the HRE for the 4 selected phases of the calibrated tide. Panel (a), 1.5h before high tide, corresponds to the strongest flood currents. At high tide, in panel (b), weaker currents are still flowing into the Lower Bay, the Hudson and East Rivers, with currents being larger in the central channel. In panel (c), 1.5h after high tide, while strong currents are still flowing into the Hudson River, the East River is at slack, and strong currents are ebbing out of the Lower Bay. Finally, in panel (d), 3h after high tide, ebbing currents are flowing out of the Lower Bay and the rivers, and are strongest near the mouth of the Bay. In this simulation, the strongest currents nearly reach 1.5 m/s (3 knots), which is notably larger (more than twice) than the currents simulated by Tajalli-Bakhsh [24] (and observed) in the wider Chesapeake Bay and even in the James River.

In view of these current patterns, one might anticipate that the second and third phases of dynamic tsunami-tide simulations, in which the largest wave in each incident tsunami reaches the Sandy Hook gage, near the mouth of the HRE, concurrently or 1.5h after high tide, should lead to the maximum amplification of the incident tsunamis, at least, near the mouth of the Bay. Indeed, while for these phases tidal elevations are either maximum or have not yet decreased too much from their highest level, the tsunamis propagating into the Lower Bay will be facing opposite (ebbing) currents that will be increasing or be already quite strong (0.5 to 0.75 m/s in Fig. 1.15c); these opposite currents will continue to strengthen as the tsunamis propagate into New York Harbor and the Hudson and East Rivers (as seen in Fig. 1.15c and d) and should cause the tsunami surface elevation to rise, at least initially. In Fig. 1.15d, while currents are even stronger 3h after high

tide, tide surface elevations are starting to become negative and hence it will be harder to achieve higher elevations in the combined results.

Being both long waves, without nonlinear interactions, tide and tsunami should be propagating into the HRE at the same phase speed and their combined level should evolve in a way similar to the individual levels. Nonlinearity, however, will affect these features, first by causing amplitude dispersion effects that will move the maximum of the combined elevations ahead or behind the initially combined values, as well as amplification of the tsunami elevation by opposing (ebbing) currents and vice versa. Additionally, the larger/lower currents occurring near the seafloor in the dynamic tsunami-tide simulations will cause more/less dissipation of the tsunami by bottom friction. To identify and quantify nonlinear effects, results of each dynamic tsunami-tide simulation will be compared to those of the corresponding simulation done over a static water level equivalent to the space averaged MHW in grid G1 (+0.64 m NAVD88). Specifically, maximum and instantaneous computed surface elevations, and time series of those and of corresponding currents at selected reference gages, will be compared across grid G1.

Figure 1.16 compares surface elevations computed in the dynamic and static simulations for the CRT SMF proxy tsunami, which causes the largest incident tsunami in the HRE at the entrance of Lower Bay and at Battery Point (Stations #1 and #3 in Fig. 1.17), for the 4 selected phases of the tide. Additionally the linear superposition of the calibrated tide with the tsunami elevations is also plotted, which allows quantifying the importance of nonlinear interactions. (As before, static simulation results were slightly adjusted to the local MHW based on VDatum data, to illustrate the competing effects of a fluctuating mean water level (MWL) and opposing tidal currents.) At the entrance to Lower Bay (Station

#1), the surface elevation of the leading tsunami crest in the linear combination exceeds that of the dynamic simulation while the tide current is co-flowing (i.e., when the tsunami arrives 1.5 hours or less before high tide; cases (a) and (b)). When the current starts ebbing (cases (c) and (d)), the elevation of the dynamic tsunami-tide simulation gradually exceeds that of the linear combination. This pattern is also observed at Battery Point (station #3). Hence, the expected effect of an opposing current to enhance the leading tsunami crest is indeed predicted in the dynamic simulations. Figure 1.16 also shows that, at both stations, the simulation over a static MHW level yields a larger absolute surface elevation of the leading tsunami crest than the dynamic simulations, for all tide phases except case (b), when the tsunami arrives at high tide. Here, the elevation of the dynamic simulation slightly exceeds that of the static simulation for part of the time series. Results and differences between static and dynamic simulations are further detailed below.

Thus, Figures 1.18, 1.19, 1.20 and 1.21 show differences of maximum envelopes of surface elevations computed for the dynamic tsunami-tide and the static MHW level simulations, for the PRT, CVV (both leading and maximum crest cases), and CRT tsunamis, respectively. Results are shown for dynamic combinations corresponding to the selected tsunami crest arriving at three tidal phases: (1) 1.5h before; (2) concurrent with; and (3) 1.5h, after high tide. For PRT, Fig. 1.18 shows that the worst dynamic case scenario as far as coastal flooding is when the leading tsunami crest arrives at high tide at the Sandy Hook station, causing an increase in inundation in the HRE by 0.03 to 0.07 m up to Battery point. For CVV, when synchronizing the largest crest in the tsunami wave train with the tide, Fig. 1.19 shows that the worst dynamic case scenario is also for high tide, leading to a slightly increased flooding, by up to 0.05 m at the entrance to Lower

Bay. Upon entering Lower Bay, this crest interacts with tidal currents that have already been disturbed by more than 2h of tsunami propagation into the bay, and reflection coming back from the upper part of the HRE. The confused currents within the Bay are likely responsible for the mild decrease (by up to -0.05m) in surface elevation seen across the remainder of grid G1. This is confirmed in Fig 1.20, where instead the slightly smaller leading crest of the CVV tsunami was synchronized with the same 4 tide phases. Despite the lower crest, the dynamic simulation at high tide predicts inundations that exceed the static case by 0.05-0.1 m, up to the East River. Finally, Fig. 1.21 shows that the dynamic simulations of the CRT SMF proxy tsunami cause the largest increases in coastal flooding, by up to 0.15 m, again for the same high-tide phase. Unlike the other two PMTs, however, the largest increases in surface elevation are observed in the southern portion of Lower Bay and no or a negligible increase is observed in the Upper Bay.

As an overall summary of tsunami flooding hazard in the HRE, Figure 1.22 shows the maximum envelopes of surface elevations computed for each of the three incident tsunamis, over the 4 tested tide phases (i.e., envelopes of the dynamic simulations on which the difference plots of Figs. 1.18, 1.19, 1.20 and 1.21 are based, plus the 3h delay case), and their difference with the envelope of the same results over a static tide level (Fig. 1.14) is plotted in Fig. 1.23. (Note, as before, eight tide cases (4 for each of the two large crests) are included in the CVV dynamic envelope in Figure 1.22.) These results confirm that dynamic tsunami-tide interactions can cause a slightly increased flooding in the HRE (by up to 0.15 m), especially in the southern and southwest regions of Lower Bay, and in Battery Point and the East River around Manhattan. It also appears that the increase in flooding associated with the leading crest of the longer period far-field tsunamis (PRT and CVV) occurs relatively farther inland than that of the shorter period

near-field tsunami (CRT). While Fig. 1.16 shows that much of the higher frequency content of the CRT tsunami signal is filtered out by the time waves reach Battery Point, the lower frequencies of the PRT and CVV tsunamis are able to penetrate deeper.

As mentioned before, animations of model results for the three PMTs propagating into grids G3a,b, G2 and G1 are provided as supplementary online material: PRT.mp4, CVV.mp4, CRT.mp4. These more clearly show where the largest surface elevations occur in the HRE and their magnitude and phase. In grids G2 and G1, animations are based on results of dynamic tsunami-tide simulations for the case of each tsunami arriving at the Sandy Hook gage (station #1) concurrently with high tide. Results discussed above indicate that this represents the worst case (flooding) scenario at most locations. In grid G1, the animations show two side-by-side panels, one for the dynamic simulations and the second one, for comparison, for tsunami propagation over the static MHW level. Results in grid G1 also show instantaneous currents as velocity vectors (i.e., side-by-side for tsunami alone or combined tsunami-tide currents), which allows more easy understanding of how the tsunami velocity field is spatially modified. Currents are further detailed and analyzed in the next section.

1.6 Detailed Analysis of Results and Discussion

Results presented above indicate that tsunami-tide interactions may lead to increased flooding and stronger currents in some areas of the HRE, depending on tide phase. Here, we further analyze the physical mechanisms governing these interactions on the basis of time series of currents (Figs. 1.24, 1.25, 1.26) and surface elevations (Figs. 1.27, 1.28, 1.29) computed at 5 numerical gages (marked in Fig. 1.17), from the entrance of Lower Bay to the upper part of the Hudson River. Results are from simulations in the finer resolution grid G1, for the 3

PMTs combined with the three main phases of the tide considered so far, i.e., 1.5 h before, concurrent with, and 1.5 h after high tide, plus the 3h after high tide phase, to have a case with stronger ebbing currents. Note that here currents have been projected in the local main direction of tsunami propagation at each station. In Figs. 1.24, 1.25 and 1.26, we compare currents computed in the dynamic (nonlinear) tsunami-tide simulations, the static tsunami simulations over a MHW level, and the linear superposition of the corresponding tide and tsunami currents. For surface elevations in Figs. 1.27, 1.28 and 1.29, we compare the detided dynamic and static results; detiding is done by subtracting the corresponding tide surface elevations (MHW or dynamic level, depending on the considered case).

Figures 1.27, 1.28 and 1.29 show similar current patterns for the three tsunami cases, which have the expected behavior. At all stations, when the tide is co-flowing (flooding current), the nonlinearly combined tide and tsunami currents are either equal to or have a slightly smaller magnitude than the linearly combined currents, likely as a result of the increased bottom friction dissipation. When the tide is opposite (ebbing current), this trend reverses itself and the nonlinearly combined currents become larger than the linearly combined currents. For the selected tide phases, however, which are aimed at maximizing the combined tsunami-tide elevations, the latter mostly occurs in the tail of the tsunami wave trains and, as we shall see, when surface elevations are lower; hence, the impact on maximum flooding is minimal. In all cases, higher frequency oscillations seen in the incident tsunami currents are gradually damped as the tsunamis propagate up the estuary, also likely as a result of dissipation by increased bottom friction. Most of this damping happens by the time the tsunamis reach station #2. A slightly early arrival of tsunami currents is observed in the dynamic case when there is a favorable (co-flowing) tidal current, which indicates an increase in wave phase

velocity; consistent with this, a later arrival is observed for ebbing currents and this difference in arrival time progressively increases as the tsunamis propagate up the HRE (from station #1 to 5).

In Figs. 1.24, 1.25 and 1.26, the patterns of surface elevations are similarly divided between the same stations, with at stations #2 to #5, most of the higher frequency oscillations having been filtered out. As could be expected from elementary long wave theory, when facing an opposite (ebbing) current tsunami elevations increase as compared to the simulations performed over a static MHW level, and they decrease when traveling with the (flooding) current. This becomes more prominent as the tsunamis propagate upstream the HRE. However, as this dynamic increase in surface elevation mostly occurs in the tail of the tsunami trains, while the tidal elevations are decreasing, this does not affect maximum flooding. When facing a co-flowing (flooding) tidal current, tsunami elevations, as for the currents, are equal or slightly smaller than those found over a static MHW level. Consistent with observations made for the currents, changes in tsunami phase speed are observed, with the dynamic cases slightly lagging behind the static cases when facing an opposite (ebbing) tide current, and vice versa. Overall, observed differences between dynamic and static tsunami-tide simulations are consistent with predictions of elementary wave theory on wave-current interactions. In grid G1, tidal currents exceed 1 m/s (2 kts), which is comparable to currents caused by the incoming PMTs. This similarity in current magnitude results in meaningful nonlinear interactions between the two long wave trains. At the entrance to Lower Bay (station #1 in Fig. 1.17), slack tide occurs about 1 h after high tide, but in the Hudson River, flooding currents persist for more than two hours after high tide. Hence, leading tsunami crests arriving 1.5 h after high tide (during the initiation of the ebbing current) will still experience a favorable current in the Hudson River and

thus will tend to decrease in elevation. Another phenomenon affecting dynamic tsunami-tide simulations is that during lower tide elevation periods, the tsunamis propagate over shallower water areas and hence could end up shoaling somewhat more in some areas than in the static simulations; however, as results of the propagation over a lower tide have shown (i.e., 3 h delay), this in general does not lead to increased maximum flooding in the HRE over the entire simulation. Differences between surface elevation time series for the dynamic tsunami-tide simulations and those for the tsunami propagating over a static MHW level thus mostly result from nonlinear interactions between tide and tsunami currents.

1.7 Conclusions

We performed simulations of dynamic tsunami-tide interactions in the Hudson River estuary (HRE) and compared results to the standard tsunami simulations, which are performed over a static MHW tide level. In both cases, the maximum tide level (static or maximum dynamic) was selected as the average maximum tidal elevation reached in the HRE for an MHW tide during the period spanning 7:00 am on 13 July 2015 to 7:00 pm on 14 July 2015. Overall, dynamic tsunami-tide simulations only predict a modest increase in maximum inundation in the HRE, 0.05-0.15 m for the three selected PMTs and four tide phases, as compared to static simulations. More specifically, Figs. 1.18, 1.19 and 1.21, which show maximum envelopes of differences between the dynamic and static results computed for each PMT and the three main tide phases considered here, and Fig. 1.23, which shows the envelopes of these, indicate that areas with the largest increases in surface elevation resulting from nonlinear tsunami-tide interactions are located at both the entrance and the southern region of Lower Bay (Sandy Hook Bay and Raritan Bay). Tsunamis arriving at the entrance to Lower Bay are trains of long waves that have nearly depth-uniform currents with maximum magnitude similar to that of

Figure 1.24. Time series of currents computed in grid G1, at stations #1-5 (labels; Fig. 1.17), for the PRT tsunami (*black*), tide (*red*), and their linear (*green*) and nonlinear (*blue*) combinations: 1.5 h before, concurrent with, 1.5 h after and 3 h after high tide, from leftward to rightward columns. Time is measured from the beginning of the event.

Figure 1.25. Same results as in Fig. 1.24 for the CVV tsunami.

Figure 1.26. Same results as in Fig. 1.24 for the CRT SMF proxy tsunami.

Figure 1.27. Time series at stations #1-5 (labels) in grid G1 (Fig. 1.17) of tidal currents projected in the tsunami direction of propagation (*red*), and surface elevations for the tide (*green*) and PRT tsunami: over static MHW level (*black*), and in dynamic combination with the tide (*blue*): 1.5h before, concurrent with, 1.5h after and 3 h after high tide from leftward to rightward columns. Tsunami surface elevations (static and dynamic) have been detided by subtracting the corresponding tide surface elevations (i.e., static MHW level or dynamic level). Time is measured from the beginning of the event.

Figure 1.28. Same results as in Fig. 1.27 for the CVV tsunami.

Figure 1.29. Same results as in Fig. 1.27 for the CRT SMF proxy tsunami.

currents caused by the selected tide. Results show that, for opposite currents, the three considered PMTs experience dynamic increases in surface elevations near high tide as far inland as Upper Bay (New York Harbor). Further upstream, the increase in bottom friction resulting from tsunami-tide interactions leads to reduced surface elevations as compared to a simulation over a static MHW level.

Although the maximum increases in surface elevation resulting from dynamic tsunami-tide interactions are both localized and not very significant in view of the maximum absolute tsunami flooding in the HRE (up to 3.5 m; Fig. 1.22), they nevertheless indicate that nonlinear interactions between tide and tsunami currents are meaningful. Hence, such interactions could become a significant factor in tsunami hazard assessment (such as performed in the NTHMP work), in bays or estuaries with larger tidal currents than in the HRE. Since dynamic tsunami-tide effects are highly site-specific; however, when one suspects that significant tidal currents can occur, high-resolution simulations should be performed to accurately estimate their effects on local tsunami hazard, particularly if the coastline geometry and bottom topography are complex.

As a final illustration of this work, Figure 1.30 shows the extent of maximum tsunami inundation on Staten Island predicted for the three PMTs (in each case, the envelope of the maximum elevation for the four considered tide phases), in the static and dynamic simulation cases. Staten Island is on the west side of Lower Bay, which is an area especially vulnerable to tsunami inundation. We see that the inundation extent of the dynamic case encompasses that of the static case, except in a very small area for the CVV case.

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Figure 1.30. Extent of maximum tsunami inundation in eastern Staten Island from : (*yellow*) maximum envelopes computed over four tidal phases in dynamic tsunami-tide simulations (Fig. 1.22); and (*red*) the linear superposition of tsunamis over a static tide level (local MHW), for the three considered PMTs: PRT (*left*), CVV (*center*), and CRT SMF proxy (*right*).

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List of References

- [1] R. Dean and R. Dalrymple, *Water wave mechanics for engineers and scientists*, ser. Advanced Series on Ocean Engineering. Prentice-Hall, 1991, vol. 4.
- [2] S. Grilli, C. O'Reilly, J. Harris, T. Tajalli-Bakhsh, B. Tehranirad, S. Banihashemi, J. Kirby, C. Baxter, T. Eggeling, G. Ma, and F. Shi, "Modeling of SMF tsunami hazard along the upper US East Coast: Detailed impact around Ocean City, MD," *Natural Hazards*, vol. 76, no. 2, pp. 705–746, 2015.
- [3] E. Tolkova, "Tide-tsunami interaction in the Columbia River, as implied by historical data and numerical simulations," *Pure and Applied Geophysics*, vol. 170, no. 6–8, pp. 1115–1126, 2013.
- [4] S. T. Grilli, O.-D. S. Taylor, C. D. Baxter, and S. Marezki, "Probabilistic approach for determining submarine landslide tsunami hazard along the upper East Coast of the United States," *Marine Geology*, vol. 264, no. 1–2, pp. 74–97, 2009.
- [5] B. Tehranirad, S. Banihashemi, J. T. Kirby, J. A. Callahan, and F. Shi, "Tsunami inundation mapping for Ocean City, MD NGDC DEM," Center for Applied Coastal Research, Department of Civil and Environmental Engineering, University of Delaware, Tech. Rep., 2014, research Report No. CACR-14-04.

- [6] U. ten Brink, D. Twichell, E. Geist, J. Chaytor, J. Locat, H. Lee, B. Buczkowski, R. Barkan, A. Solow, B. Andrews, T. Parsons, P. Lynett, J. Lin, and M. Sansoucy, "Evaluation of tsunami sources with the potential to impact the U.S. Atlantic and Gulf Coasts," USGS Administrative report to the U.S. Nuclear Regulatory Commission, Tech. Rep., 2008, 300 pps.
- [7] U. S. ten Brink, J. D. Chaytor, E. L. Geist, D. S. Brothers, and B. D. Andrews, "Assessment of tsunami hazard to the US Atlantic margin," *Marine Geology*, vol. 353, pp. 31–54, 2014.
- [8] S. Grilli, A. Grilli, B. Tehranirad, and J. Kirby, "Modeling tsunami sources and their propagation in the Atlantic Ocean for coastal tsunami hazard assessment and inundation mapping along the US East Coast," in *Proc. 2015 COPRI Solutions to Coastal Disasters Conf.* Boston, USA: American Soc. Civil Eng., September 2015, p. 12.
- [9] S. T. Grilli, S. Dubosq, N. Pophet, Y. Pérignon, J. Kirby, and F. Shi, "Numerical simulation and first-order hazard analysis of large co-seismic tsunamis generated in the Puerto Rico trench: near-field impact on the north shore of Puerto Rico and far-field impact on the US East Coast," *Natural Hazards and Earth System Sciences*, vol. 10, pp. 2109–2125, 2010.
- [10] A. Grilli and S. Grilli, "Modeling of tsunami generation, propagation and regional impact along the upper U.S East Coast from the Puerto Rico trench," 2013b. [Online]. Available: <http://personal.egr.uri.edu/grilli//grilli-grilli-cacr-13-02>
- [11] R. Barkan, U. ten Brick, and J. Lin, "Far field tsunami simulations of the 1755 Lisbon earthquake: Implication for tsunami hazard to the U.S. East Coast and the Caribbean." *Marine Geology*, vol. 264, no. 1–2, pp. 109–122, 2009.
- [12] A. Grilli and S. Grilli, "Modeling of tsunami generation, propagation and regional impact along the U.S. East Coast from the Azores Convergence Zone," 2013a, research Report no. CACR-13-04, 20 pps. [Online]. Available: <http://personal.egr.uri.edu/grilli//grilli-grilli-cacr-13-04>
- [13] S. Ward and S. Day, "Cumbre Vieja Volcano potential collapse at La Palma, Canary Islands," *Geophysical Research Letter*, vol. 28, no. 17, pp. 3397–3400, 2001.
- [14] S. Abadie, J. Harris, S. Grilli, and R. Fabre, "Numerical modeling of tsunami waves generated by the flank collapse of the Cumbre Vieja Volcano (La Palma, Canary Islands): Tsunami source and near-field effects," *J. Geophys. Res.*, vol. 117, no. CH03050, 2012.

- [15] B. Tehranirad, J. Harris, A. Grilli, S. Grilli, S. Abadie, J. Kirby, and F. Shi, "Far-field tsunami hazard in the North Atlantic basin from large scale flank collapses of the Cumbre Vieja Volcano, La Palma," *Pure Appl. Geophys.*, p. 28 pps, 2015.
- [16] F. Shi, J. T. Kirby, J. C. Harris, J. D. Geiman, and S. T. Grilli, "A high-order adaptive time-stepping TVD solver for Boussinesq modeling of breaking waves and coastal inundation," *Ocean Modelling*, vol. 43–44, pp. 36–51, 2012.
- [17] J. Kirby, F. Shi, T. B., J. Harris, and S. Grilli, "Dispersive tsunami waves in the ocean: Model equations and sensitivity to dispersion and Coriolis effects," *Ocean Modelling*, vol. 62, pp. 39–55, 2013.
- [18] Z. Kowalik, T. Proshutinsky, and A. Proshutinsky, "Tide-tsunami interactions," *Science of Tsunami Hazards*, vol. 24, no. 4, pp. 242–256, 2006.
- [19] Z. Kowalik and A. Proshutinsky, "Tsunami-tide interactions: A Cook Inlet case study," *Continental Shelf Research*, vol. 30, no. 6, pp. 633–642, 2010.
- [20] Y. Zhang, R. Witter, and G. Priest, "Tsunami-tide interaction in 1964 Prince Williams Sound tsunami," *Ocean Modelling*, vol. 40, no. 3–4, pp. 246–259, 2011.
- [21] H. Yeh, E. Tolkova, D. Jay, S. Talke, and H. Fritz, "Tsunami hydrodynamics in the Columbia River," *J. Disaster Res.*, vol. 7, no. 5, pp. 604–608, 2012.
- [22] E. Tolkova, H. Tanaka, and M. Roh, "Tsunami observations in rivers from a perspective of tsunami interaction with tide and riverine flow," *Pure and Applied Geophysics*, vol. 172, no. 3–4, pp. 953–968, 2015.
- [23] S. Nakada, M. Hayashi, S. Koshimura, S. Yoneda, and E. Kobayashi, "Tsunami simulation generated by the greatest earthquake scenario along the Nankai Trough under consideration of tidal currents in a large bay," in *Proceedings of the Twenty-fifth (2015) International Ocean and Polar Engineering Conference*. International Society of Offshore and Polar Engineers, 2015, pp. 811–816.
- [24] T. S. Tajali-Bakhsh, S. T. Grilli, and A. R. Grilli, "Dynamic tide effects on tsunami coastal hazard in large estuaries: Case of the Chesapeake Bay/James River, USA," Center for Applied Coastal Research, Dept. of Civil and Environmental Engineering, University of Delaware, Tech. Rep., 2014, research Report no. CACR-15-09, 42 pps.
- [25] E. Geist, P. Lynett, and J. Chaytor, "Hydrodynamic modeling of tsunamis from the Currituck landslide," *Marine Geology*, vol. 264, pp. 41–52, 2009.

- [26] G. Wei, J. Kirby, S. Grilli, and R. Subramanya, “A fully nonlinear Boussinesq model for free surface waves. part i: Highly nonlinear unsteady waves,” *Journal of Fluid Mechanics*, vol. 294, pp. 71–92, 1995.
- [27] S. Grilli, M. Ioualalen, J. Asavanant, F. Shi, J. Kirby, and P. Watts, “Source constraints and model simulation of the December 26, 2004 Indian Ocean tsunami,” *Journal of Waterway, Port, Coastal, and Ocean Engineering*, vol. 133, no. 6, pp. 414–428, 2007.
- [28] M. Ioualalen, J. Asavanant, N. Kaewbanjak, S. T. Grilli, J. T. Kirby, and P. Watts, “Modeling the 26th December 2004 Indian Ocean tsunami: Case study of impact in Thailand,” *J. Geophys. Res.*, vol. 112, p. C07024, 2007.
- [29] S. Grilli, J. Harris, T. Tajalli-Bakhsh, T. Masterlark, C. Kyriakopoulos, J. Kirby, and F. Shi, “Numerical simulation of the 2011 Tohoku tsunami based on a new transient FEM co-seismic source: Comparison to far- and near-field observations,” *Pure and Applied Geophysics*, vol. 170, pp. 1333–1359, 2013.
- [30] B. Tehranirad, F. Shi, J. Kirby, J. Harris, and S. Grilli, “Tsunami benchmark results for fully nonlinear Boussinesq wave model FUNWAVE-TVD, version 1.0,” Center for Applied Coastal Research, University of Delaware, Tech. Rep., 2011, research Report No. CACR-11-02.
- [31] F. Shi, J. Kirby, and B. Tehranirad, “Tsunami benchmark results for spherical coordinate version of FUNWAVE-TVD (version 1.1),” Center for Applied Coastal Research, Univ. of Delaware, Tech. Rep., 2012b, research Report no. CACR-12-02.
- [32] G. Ma, F. Shi, and J. Kirby, “Shock-capturing non-hydrostatic model for fully dispersive surface wave processes,” *Ocean Modelling*, vol. 43–44, pp. 22–35, 2012.
- [33] *Region II Storm Surge Project - Coastal Terrain Processing Methodology*, Federal Emergency Management Agency (FEMA), Department of Homeland Security, 500 C Street, SW Washington DC, 20472, September 2014.
- [34] NOAA-NGDC, “National geophysical data center, U.S. coastal relief mode,” 2013, retrieved December 2013. [Online]. Available: <http://www.ngdc.noaa.gov/mgg/coastal/crm.html>
- [35] “Estimates of monthly and annual net discharge, in cubic feet per second, of Hudson River at New York, N.Y. (mouth),” 2010, U.S. Department of the Interior, US Geological Survey. [Online]. Available: http://ny.water.usgs.gov/projects/dialer_plots/Hudson_R_at_NYC_Freshwater_Discharge.htm
- [36] D. Stammer, R. D. Ray, O. B. Andersen, B. K. Arbic, W. Bosch, L. Carrere, Y. Cheng, D. S. Chinn, B. D. Dushaw, G. D. Egbert, S. Y. Erofeeva, H. S.

- Fok, J. A. M. Green, S. Griffiths, M. A. King, V. Lapin, F. G. Lemoine, S. B. Luthcke, F. Lyard, J. Morison, M. Muller, L. Padman, J. G. Richman, J. F. Shriver, C. K. Shum, E. Taguchi, and T. Yi, “Accuracy assessment of global barotropic ocean tide models,” *Reviews of Geophysics*, vol. 52, no. 3, pp. 243–282, 2014.
- [37] Y. Okada, “Surface deformation due to shear and tensile faults in a half space,” *Bull. Seismol. Soc. America*, vol. 75, no. 4, pp. 1135–1154, 1985.
- [38] E. Gica, M. Spillane, V. Titov, C. Chamberlin, and J. Newman, “Development of the forecast propagation database for NOAA’s short-term inundation forecast for tsunamis,” National Oceanographic and Atmospheric Administration,” NOAA Tech. Memo., 2008, OAR PMEL-139.

MANUSCRIPT 2

**Modeling coastal tsunami hazard from submarine mass failures: effect
of slide rheology, experimental validation, and case studies off the US
East coast**

by

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Abstract. We first validate two models simulating tsunami generation by deforming submarine mass failures (SMFs) against laboratory experiments for SMF made of glass beads moving down a steep slope. These are two-layer models, in which the upper layer is water, simulated with the 3D (σ -layer) non-hydrostatic model NHWAVE, and the SMF bottom layer is simulated with depth-integrated equations and represented either as a dense Newtonian fluid or a granular medium. Using the dense fluid model, we assess model convergence with grid resolution, and sensitivity of slide motion and generated surface elevations to slide parameters (i.e., viscosity, bottom friction, and initial submergence). A more limited validation is conducted for the granular slide model. Both models can accurately simulate time series of surface elevations measured at 4 gages, while providing a good simulation of both the geometry and kinematics of the moving slide material. In its current state of development, which has some limitations, the granular model provides a lesser agreement with measured surface elevations than the viscous flow model, but describes slide geometry and motion slightly better.

The viscous slide model, which at present is the only one that can be applied to an arbitrary bottom bathymetry, is then used to simulate the historic Currituck SMF motion, in order to determine relevant viscous slide parameters to simulate SMF tsunamis on the east coast. The same parameters are then applied to simulate tsunami generation from a possible SMF sited near the Hudson River Canyon. As for Currituck, this SMF was simulated in earlier work as a rigid slump supposed to provide a conservative estimate of tsunami generation. Simulations are performed for 3 deforming slides with different dissipation parameters and the rigid slump, and results compared; all SMFs have the same initial volume, location, and geometry. As expected from its larger velocity and acceleration throughout motion as well as constant geometry (while the deforming slides “flow”), a larger tsunami is

generated by the rigid slump than when assuming a deforming slide. Simulations of tsunami propagation are then pursued for the 4 generated tsunamis in two levels of nested grids, using the Boussinesq model FUNWAVE-TVD, and maximum surface elevations computed along a 5 m depth contour off of the coast of New Jersey and New York. From very large initial surface elevations (up to 25 m for the rigid slump), nearshore tsunami elevations significantly reduce in all cases (up to 6.5 m), due to both directional energy spreading and bottom friction over the wide shelf. At most nearshore locations, however, surface elevations caused by the rigid slump are significantly larger (up to a factor of 2) than those caused by the 3 deforming slides. Hence, the rigid slump provides a conservative estimate of SMF tsunami impact in terms of maximum inundation/runup at the coast, while using a more realistic rheology with some level of SMF deformation, in general, leads to a reduced tsunami impact at the coast. This validates as conservative the tsunami hazard assessment and inundation mapping performed to date as part of NTHMP, on the basis of Currituck SMF proxies simulated as rigid slump.

Ongoing and future work will focus on extending the granular slide model, which features more complete and realistic to more accurately simulate tsunami generation from deforming SMFs, in a variety of context and rheology.

2.1 Introduction

2.1.1 General context

As evidenced by the two recent catastrophic events, of 2004 in the Indian Ocean (IO; e.g., [1, 2]) and 2011 in Tohoku (TO), Japan ([3, 4]), extreme tsunamis can devastate the world’s increasingly populated coastal areas, causing high fatalities (over 200,000 for the IO tsunami and over 18,000 for the TO tsunami), destroying fragile coastal infrastructures, and imposing enormous loss to the economy of the most impacted countries (e.g., an over \$300B loss due to the TO tsunami im-

pact in Japan;[5]). Commensurate with their catastrophic impact, the IO and TO events were associated with the 3rd and 5th largest earthquakes ever witnessed in human history (M_w 9.3 and 9.1, respectively), and were essentially triggered by the coseismic seafloor deformation induced in the subduction zones (SZ) where their epicenter was located (Andaman SZ and Japan Trench, respectively). As these SZs were in close proximity to the Indonesian and Japanese coastline, respectively, both the tsunami propagation time from the start of each event was short (15-25 min), thus reducing warning times, and the coastal impact was maximized on these countries' coastline as there was no time and space for wave energy spreading to occur; in both cases extreme tsunami runups of over 40 m were measured.

However, for the TO tsunami, despite the large earthquake magnitude, many studies have shown that the largest coastal inundation and runup that impacted the 80 km long Sanriku coast (to the North of the main rupture area), could not be explained by the seismically triggered tsunami alone (e.g., [3] and references herein). Studies based on tsunami waveform inversion (e.g.,[6, 7, 8]) showed that an additional source of tsunami generation near the trench, to the north of the main rupture, was required to explain these observations. While several seismically related mechanisms were proposed to explain this additional source of wave generation (e.g., [9]), Tappin *et al.* [4] showed that there is strong evidence that seismic waves may have triggered a large submarine mass failure (SMF) near the trench, in 3,000-4,000 m depth, north of the main rupture (volume estimated to about 500 km³). This SMF would have generated the large amplitude, higher frequency (3-4 min period), waves observed at nearshore GPS Iwate buoys off of Sanriku, that superimposed with the longer (20-40 min period) coseismically triggered tsunami waves, and caused a large focused impact on the Sanriku coast. By comparing pre- and post-tsunami seafloor surveys as well as performing inverse

ray tracing of higher-frequency waves, Tappin *et al.* [4] found evidence for a SMF covering an area approximately 30 km along-trench and 20 km across-trench; a slope stability analysis confirmed the high likelihood for a seismically triggered slope failure at the proposed location. Based on the presence of many other large historical slumps in the area, they modeled the SMF as a rigid slump, with small angular motion and runout, consistent with the observed seafloor deformation. Following the methodology detailed in [10], they modeled wave generation from this SMF using the non-hydrostatic three-dimensional (3D) (sigma layer) model NHWAVE [11], in which they specified the SMF law of motion and geometry as bottom boundary conditions, based on earlier numerical and experimental work by Grilli and Watts [12, 13], Watts *et al.* [14], and Enet and Grilli [15, 16, 17]. Once the waves were generated and slump motion terminated, they continued modeling tsunami propagation towards the coast of Japan with the two-dimensional (2D) fully nonlinear and dispersive Boussinesq long wave model FUNWAVE-TVD, in a series of nested grids of increasingly fine resolution [18, 19]. Waves predicted at the locations of nearshore GPS buoys and offshore DART buoys, as well as the modeled coastal inundation and runup, agreed very well with observations, when the SMF was triggered with a 2'30" time delay, consistent with the propagation time of seismic waves from the earthquake epicenter to the assumed SMF location. Tappin *et al.* concluded that the potential for large SMFs to be triggered by megathrust earthquakes and contribute large waves to tsunami generation had important implications for assessing coastal hazard from similar future events.

Although less catastrophic to society, at least to modern society, than that associated with the TO event, large tsunamis have been triggered by SMFs or subaerial slides in the distant to recent past. While it is not our purpose to provide an exhaustive list, we will point out important events, such as the 24-50ky BP 165

km^3 Currituck SMF [20, 21, 10], 8150 BP 3,500 km^3 Storegga subaerial slide [22], the 200 km^3 1929 Grand Bank SMF [23, 24], the 40 km^3 1946 Unimak SMF [25, 26, 27], 0.03 km^3 1958 Lituya Bay subaerial slide [28, 29], the 0.027 km^3 1975 Kitimat subaerial slide [30, 31], and the 0.0003 km^3 1994 Skagway subaerial slide [32] events, that have been the object of an increasing body of work (see, e.g., the review of historical SMFs by Harbitz *et al.* [33] for additional cases and a discussion). [It should be noted that many of these historical cases occurred as highly deforming debris flows.] However, studies of SMF tsunamis really intensified after the seminal 1998 Papua New Guinea (PNG) event (e.g. [34, 35]). Here, a moderate M_w 7.1 earthquake, that should not have been significantly tsunamigenic, had apparently triggered a very large tsunami, causing flow depths of up to 15 m on the Sissano spit, killing over 2,200 people in the process. In view of this conflicting evidence, a large number of field and numerical studies were conducted, which showed that the earthquake had triggered a large (about 6 km^3) underwater slump, with a 15 min delay, 1,600 m deep off the Sissano Spit, whose waves were responsible for the large inundation and runup focused on the spit [35]. This event was the first well documented case supporting the large tsunamigenic and destructive potential of SMF tsunamis, that led to a large number of studies and developments in theoretical/numerical models (e.g., [12, 13], [25, 14], [36], [37, 38], [39, 40, 41], [42, 43], [44], [29], [45]; for details, see the very exhaustive review by Yavari-Ramshe and Ataie-Ashtiani, [46]) and laboratory experiments (e.g., [15, 16, 17], [13], [28, 47], [48], [41], [49], [50]), as well as a reanalysis of past events (e.g., 1946 Unimak) in view of new evidence ([27]), that had been potential SMF candidates, but whose analysis or modeling had not been fully conclusive [26].

Past volcanic eruptions have also been associated with the generation of large and destructive tsunamis, from pyroclastic flows and/or caldera collapse, such as

the 1883 Krakatau [51] or more recently the 2002 Stromboli [52], eruptions. Additionally, fast growing young volcanoes suffer periodic large mass failures, causing partial to total flank collapses of large tsunamigenic potential. This has been documented for the Hawaii volcanoes by Moore *et al.* [53], the Canary Islands volcanoes (in particular the Cumbre Vieja Volcano (CVV) on La Palma) by McMurtry *et al.* [54], which estimate that the latest large-scale collapse of the CVV may have occurred 250ka ago, and in a recent study of the Fogo volcano on the Cape Verde island, by Ramalho *et al.* [55] who showed that a flank collapse may have catastrophically happened 73ka ago, as at least one fast voluminous event that triggered tsunamis of enormous height and energy, causing over 270 m runup on the nearby Santiago island. In more recent history, in 1888, a 5 km³ flank collapse of Ritter Island (Papua New Guinea) produced damaging tsunami waves at distances of up to 500 km; the resulting landslide removed most of the subaerial Ritter Island, reducing the 800 m high edifice to a crescent-shaped remnant [56, 57]. In 1975, a M7.2 earthquake struck the Big Island of Hawaii in Kalapana, causing a tsunami that created a 7-14 m runup on the nearby shores. In a detailed study, Day *et al.* [58] showed that a slump motion of a large part of the Kilauea volcano Southeast flank (with a 7,200 km³ volume) was likely responsible for a large part of the observed near-field tsunami. Finally, in 1980, the eruption of Mount St Helens caused a 2.5 km³ rockslide-debris avalanche [59].

2.1.2 Specific context (USEC)

Although this study has more general goals, its main focus is to improve the modeling of tsunami generation, and coastal inundation and runup, from SMFs located off of the United States (US) East Coast (USEC), along the Atlantic Ocean margin and shelf slope, in relation to tsunami hazard assessment and inundation mapping carried out under the umbrella of the US National Tsunami Hazard Mit-

igation Program (NTHMP; <http://nthmp.tsunami.gov/index.html>). Since 1995, in the wake of the devastating IO tsunami, NTHMP has supported the systematic development of tsunami inundation maps for selected areas of the US coastline (see maps at: <http://www1.udel.edu/kirby/nthmp.html>), based on high resolution numerical modeling, to allow for better tsunami hazard assessment and mitigation. As part of this activity, since 2009, most of the authors have been involved in the development of tsunami hazard maps for the USEC, in the form of envelopes of coastal inundation caused by the most probable maximum tsunamis (PMTs) in the considered oceanic basin. For the USEC, this is the Atlantic Ocean basin where PMTs due to a variety of geological processes (or sources) were identified, including [60, 61, 62]: (i) far-field coseismic sources, such as caused by a M9 earthquake affecting the entire Puerto Rico Trench (PRT) [63], or a repeat of the M8.7-8.9 1755 Lisbon earthquake in the Açores Convergence Zone (ACZ) [64]; (ii) a far-field subarerial landslide source due to a large volcanic collapse of the CVV in the Canary Islands [65, 66, 67]; and (iii) near-field SMFs, on or near the continental shelf break [68, 10, 69, 60, 70, 71, 72]. The improved modeling of the latter SMF sources, particularly those occurring on the mainly silicate shelf of the US North East and their potential impact on the USEC are the object of the present paper; this will include estimating the sensitivity of tsunami generation and coastal hazard to slide rheology.

While few historical tsunamis caused by SMFs have been clearly identified to have impacted the USEC (e.g., the 1929 Grand Bank tsunami; [24, 23]), ten Brink *et al.* [60, 70, 71, 61], Chaytor *et al.* [69] and Twichell *et al.* [73] have mapped numerous paleo-SMFs, whose total surface area covers a significant portion of the USEC continental slope and rise, with many of these being of large volume (tens to over 100 km³). In recent history, the USEC has only experienced a moderate

seismicity, up to M7.2, but this was sufficient to trigger a 200 km³ debris flow and a damaging tsunami in the Grand Banks in 1929. Other historical SMFs, such as the 1998 PNG, have confirmed that this level of seismicity can trigger large SMFs and cause catastrophic tsunamis. Specific studies of the tsunamigenic potential of SMFs off of the USEC have shown that these may pose the largest tsunami hazard among all possible PMTs in the Atlantic basin [60, 70, 71, 68, 10]. Factors that increase tsunami hazard from SMFs are (e.g., [25, 13, 14, 35, 68, 10]): (i) a possible occurrence in fairly shallow water, enhancing tsunami generation; (ii) a location at a short distance from shore, reducing energy spreading and warning times; and (iii) a more directional and focused wave generation, causing significant inundation along a narrow section of the coast.

In part due to the lack of specific data on past SMF geometry and parameters at the time and the uncertainty in identifying these and the locations of future tsunamigenic SMFs, Grilli *et al.* [68] performed Monte Carlo Simulations (MCS) of SMFs triggered by seismicity along the USEC. They defined a series of transects across the coast, initially from New Jersey to Cape Cod, but these were later extended to southern Florida for a total of 91 transects. Thousands of SMFs were simulated along each transect, with random values of seismicity, sediment properties, SMF type (slides and slumps), location/depth, geometry, and excess pore pressure being picked from known probability distributions and/or site-specific field data. Slope stability analyses were performed for each of these and, for each SMF that failed, tsunami generation and runup were estimated based on semi-empirical methods [13, 14]. Statistics of results were performed, which allowed estimating the 100- and 500-year return period runups along the entire USEC. North of Virginia, the MCS predicted a 500-year runup of up to 5-6 m, which is commensurate with the largest impact from other PMTs [62] and similar to or

greater than the 100-year hurricane storm surge in the region. South of Virginia, the MCS predicted a significantly reduced runup. In view of these results, Grilli *et al.* [10] performed geophysical and geotechnical analyses in regions of the upper USEC deemed at higher risk of large SMF tsunami runup. This led to the selection of 4 areas where large tsunamigenic SMFs could be expected to occur, due to high seismicity combined with large bottom slopes and a sufficient accumulation of sediment (Fig. 2.1.2). For the first generation of NTHMP inundation map, which are based on PMTs, one large SMF was then parameterized and tsunami generation modeled in each area, using the characteristics of the largest historical failure in the region, the 165 km³ Currituck slide complex (Fig. 2.1.2); these were referred to as SMF Currituck proxy sources.

As indicated before, Currituck is the largest paleo-SMF identified along the western Atlantic Ocean margin and occurred between 24 and 50 ka ago, in a period when sea level was much lower. This event has been extensively studied from the geological and slide triggering points of view (e.g., [20], and references herein). Tsunami generation from a reconstituted Currituck SMF was first studied by Geist *et al.* [21], using a simplified SMF tsunami generation model. Before simulating the SMF Currituck proxies in areas 1 to 4, Grilli *et al.* [10] modeled tsunami generation and coastal impact for the historical Currituck slide. Similar to the methodology detailed above for the TO tsunami, they used the non-hydrostatic 3D model NHWAVE [11] to simulate tsunami generation and then propagated the tsunami in a series of nested grids, using the 2D Boussinesq model FUNWAVE-TVD. In the bathymetry provided to NHWAVE, they recreated the unfailed slope, assuming the SMF had a quasi-Gaussian profile with an elliptical footprint. To maximize tsunami generation, as suggested by Grilli *et al.* (2005) [13], they modeled the SMF kinematics as a rigid slump, using information on the maximum

Figure 2.1. Location of areas identified by Grilli *et al.* [10], off of the upper USEC, as having high potential for large tsunamigenic SMFs (from Shelby *et al.* [74]). The red box marks the boundary of the 500 m resolution NHWAVE Cartesian grid CT used to simulate the Currituck SMF motion (see Table 1). Depth is in meters, in the color scale and bathymetric contours.

velocity of the SMF center of mass from Locat *et al.*'s (2009) work [20]. The SMF geometry and kinematics were both specified as a bottom boundary condition in NHWAVE. Results showed that, with the present sea level, Currituck would have generated a large tsunami, causing significant inundation (over 5 m) in the area of Norfolk, VA, and south of it, as well as along the shores of the Delmarva peninsula, including Ocean City, MD. Following the same methodology, Grilli *et al.* [10] then modeled tsunami generation from the four SMF Currituck proxies in areas 1-4 (Fig. 2.1.2), assuming they behaved as rigid slumps, and computed coastal inundation and runup, to be included in NTHMP inundation maps.

In this work, in view of recent progress made in SMF tsunami models, the modeling of tsunami generation and coastal impact on the USEC using SMF Currituck proxies is revisited, to estimate the impact of slide rheology on tsunami hazard. Earlier work (e.g., [75, 13]) indicates that moderate SMF deformation should generally result in a slightly reduced coastal hazard, but overall does not significantly affect tsunami generation. This may not always be the case, however, as local bathymetry, seafloor properties and SMF geometry could play an important role on slide kinematics and tsunami generation. For the purpose of developing the most realistic and accurate inundation maps for NTHMP, it is important to use the most appropriate slide model. Accordingly, in this paper, we first summarize the equations and methods for two recent two-layer SMF models developed and applied by some of the authors. In both models, the upper layer is water modeled with the 3D model NHWAVE and the slide is a depth-integrated layer modeled as: (i) a dense Newtonian fluid [31]; and (ii) a saturated granular medium [76]. Both models are then validated and benchmarked against recent laboratory experiments performed for SMFs made of small glass beads moving down a plane slope. We then apply the dense fluid model (note the granular model is not yet applicable to an arbitrary slope), to the historical Currituck slide and the SMF Currituck proxy in area 1, and compare both slide motion and tsunami generation with those of the same SMFs modeled as rigid slumps. We then draw some conclusions regarding tsunami hazard assessment.

2.2 Numerical Models and modeling methodology

A variety of numerical models have been proposed to simulate tsunami generation by SMFs or subaerial slides, particularly since the 1998 PNG event, which was caused by a deep SMF moving as a rigid slump [34, 35]. While we do not intend to provide an exhaustive review of SMF tsunami models, we will mention

models simulating tsunami generation: (1) by rigid SMFs or subaerial slides, by specifying their geometry and motion using bottom boundary conditions, with the fluid motion being modeled based on potential flow theory (e.g., [12, 13, 14, 36], dispersive long wave equations (e.g., [37, 38, 52, 40, 44], or Navier-Stokes equations (e.g., [77, 48, 41, 42, 43, 50, 10]; and (2) by deforming SMFs or subaerial slides, based on long-wave equations for both slide and coupled fluid motion (e.g., [78, 79, 32, 24]), on Navier-Stokes equations for both slide and fluid (e.g., [22, 33, 77, 41, 80, 65, 42, 43, 66, 29, 45]), and on a depth-integrated layer for the SMF and Navier-Stokes equations for the fluid [76, 31]. For details, see the exhaustive review by Yavari-Ramshe and Ataie-Ashtiani (2016) [46].

In this work, tsunami generation by SMFs is modeled with the 3D non-hydrostatic model NHWAVE [11], which uses a horizontal Cartesian grid and a boundary fitting σ -coordinate grid in the vertical direction. Once the tsunami is fully generated, the modeling of wave propagation is pursued with the 2D fully non-linear and dispersive long wave Boussinesq model FUNWAVE-TVD [18, 19], in a series of nested grids of increasingly fine resolution (see, e.g., [3, 10, 62, 4], for more details on this approach). It should be pointed out that both NHWAVE and FUNWAVE are non-hydrostatic, i.e., dispersive, wave models; the inclusion of frequency dispersion in tsunami models has been shown by many authors to be necessary for accurately modeling SMF tsunami generation and propagation, essentially, because of the typically smaller wavelength to depth ratio of SMF tsunami waves (see, e.g., [12, 13, 25, 35, 11, 81]). Failing to include dispersive/non-hydrostatic effects in the model used to simulate SMF tsunami generation has been shown to cause large errors in the shape and kinematics of initial waves (e.g., [35, 11]). Additionally, failing to include dispersion in the model used to propagate the highly dispersive wave trains that are generated will also cause large errors in wave height and

steepness (and hence coastal impact), due to a lack of constructive-destructive wave-wave interferences during propagation; the generated wave trains will also typically lack the long oscillatory (dispersive) tail observed experimentally (see the many experimental works listed before), in the field (e.g., [4]), or numerically (e.g., [25, 35, 65, 66, 67, 10]) for landslide tsunamis, and will be limited to one or two leading waves.

We model tsunami generation by rigid SMFs following the methodology detailed in Grilli *et al.* [10], in which the SMF law of motion (slump or slide) and geometry are specified as bottom boundary conditions in NHWAVE (see also [13, 14, 17]); the reader is referred to the references for the details of this method. We model tsunami generation from deforming SMFs by representing them as a layer of dense fluid [31] or a saturated granular medium layer [76], coupled along the deforming SMF-water interface with NHWAVE, which is used to simulate the resulting wave motion. In both cases, SMF equations of mass and momentum conservation are depth-integrated, similar to those obtained in a long wave generation model, and these include volumetric and bottom friction dissipation terms. In experiments, the actual initial shape of the SMF is modeled, whereas in field case studies, both rigid and deforming slide geometry is modeled as an initial sediment mound with quasi-Gaussian cross-sections and an elliptical footprint over the slope (see, [17, 10], for details).

The two deforming SMF models will be validated and benchmarked against laboratory experiments for tsunami generation by slides made of glass beads, moving down a plane slope. However, only the dense fluid slide model will be applied to field case studies, because the other model cannot yet simulate a bathymetry with arbitrary depth $h(x, y)$ (the extension to an arbitrary bathymetry is in progress). Below, we provide the governing equation for the dense fluid slide model layer,

Figure 2.2. Sketch of a vertical cross-section in Kirby *et al.*'s [31] deforming slide model, with definition of key geometric parameters.

and the reader is invited to consult the references for additional details and information, particularly on numerical implementation and methods. Following Jiang and Leblond (1992) [78] and Fine *et al.* (1998) [32], Kirby *et al.* [31] modeled the motion of a dense Newtonian fluid layer, coupled to NHWAVE, using the depth-integrated momentum equation,

$$\begin{aligned} \frac{\partial \mathbf{F}}{\partial t} + \frac{6}{5} \nabla_h \cdot \left(\frac{\mathbf{F} \mathbf{F}}{D} \right) = & -gD \left\{ -\frac{\rho_s - \rho_w}{\rho_s} \nabla_h \chi + \frac{\rho_w}{\rho_s} \nabla_h \eta \right\} \\ & - \frac{\mu}{\rho_s} \left\{ 3 \frac{\mathbf{F}}{D^2} - \nabla_h^2 \mathbf{F} \right\} - \frac{g n^2}{D^{1/3}} \frac{\mathbf{F} |\mathbf{F}|}{D^2}. \end{aligned} \quad (2.1)$$

where $\mathbf{F} = (DU, DV)$ with $D = h - \chi$ (where h is the fixed bathymetry without slide, and χ denotes the distance from the still water level (swl) to the slide interface; Fig. 2.2), $\mathbf{U} = (U, V)$ is the depth-averaged horizontal velocity vector (directions x and y), $\nabla_h = (\partial/\partial x, \partial/\partial y)$ is the horizontal gradient operator, ρ_w and ρ_s are the water and slide bulk density, respectively, μ is the slide (dense fluid) dynamic viscosity, g is gravitational acceleration, and n is the Manning coefficient for the slide-substrate bottom friction. Terms in the right-hand side of Eq. 2.1 represent the pressure gradient due to the slide-water interface and free surface

slopes, (viscous) dissipation within the slide, and bottom friction dissipation at the slide-substrate interface, respectively. Note that Eqs. 2.1 are expressed in the Cartesian horizontal axes (x, y) and assume that the slide layer is thin as compared to the scale of along slope variation, similar to a typical long wave equation model for which horizontal scales are assumed to be larger than vertical scales; hence, this also leads to assuming a mild slope and hence neglecting vertical accelerations as compared to gravity. For steeper slopes, however, such as in the laboratory experiments presented hereafter, the vertical acceleration term may no longer be negligible.

In field case studies, after the SMF has stopped moving or is too deep to be tsunamigenic (for time $t > t_f$) (whether rigid or deforming), surface elevation $\eta(x, y)$ and horizontal velocity $\mathbf{U}(x, y)$ are interpolated onto FUNWAVE-TVD's grid for further simulations. Both Cartesian coordinates, fully nonlinear [18], and spherical coordinates, weakly nonlinear [19], implementations of FUNWAVE-TVD are available, and simulations are performed in a series of one-way coupled nested grids, with increasingly fine resolution and commensurately accurate bathymetric and topographic data towards the coast. The rationale for this hybrid modeling approach is that: (i) FUNWAVE cannot currently simulate waves generated by a moving bottom; (ii) NHWAVE can simulate a moving bottom for a rigid SMF and has been extended to include a deforming slide layer [76, 31]; (iii) Being 3D, NHWAVE can more accurately simulate SMF tsunami generation in deeper water (i.e., for more dispersive waves), during which velocities can be much more non-uniform over depth than during subsequent tsunami propagation (see, e.g., [36]); (iv) Being fully nonlinear, FUNWAVE-TVD is more accurate than NHWAVE for simulating coastal wave transformations and impact, in particular, regarding wave breaking dissipation and the moving shoreline algorithm; (v) Although both models

are heavily and efficiently parallelized, FUNWAVE-TVD is more computationally efficient as it is only 2D, and hence uses grids at least 3 times smaller than the minimum NHWAVE grid with 3 σ -layers that provides a similar approximation of horizontal velocities in the vertical direction; and finally (vi) FUNWAVE-TVD also has a spherical implementation, which allows a more accurate simulation of far-field tsunami propagation. Note, in the present applications, because all simulations will be performed in regional or nearshore grids with small latitudinal and longitudinal ranges, spherical coordinates will not be necessary, although combinations of spherical and Cartesian nested grids have been used in earlier work with FUNWAVE-TVD (see, e.g., [3, 10, 62, 19]).

The one-way coupling method used in simulations of tsunami propagation in nested grids with FUNWAVE-TVD works as follows: time series of surface elevation and depth-averaged current are computed at a large number of stations/numerical wave gages, defined in a coarser grid, along the boundary of the finer grid used in the next level of nesting. Computations are fully performed in each coarser grid and then these are restarted in the next level of finer grid, using the station time series as boundary conditions. As these include both incident and reflected waves computed in the coarser grid, the coupling method closely approximates open boundary conditions. It was found in NTHMP work that a nesting ratio with a factor 3-4 reduction in mesh size allowed achieving good accuracy in tsunami simulations. Note that to prevent reflection in the first grid level, sponge layers are used along all the offshore boundaries (see details in [18]). For each grid level, whenever possible, bathymetry and topography are interpolated from data of accuracy commensurate with the grid resolution. In deeper water, we use NOAA's 1 arc-min resolution ETOPO-1 data. In shallower water and on continental shelves, we use NOAA's NGDC 3" and 1" Coastal Relief Model data.

Figure 2.3. Set-up for laboratory experiments of tsunami generation by underwater slides made of glass beads performed in the Ecole Centrale de Marseille (IRPHE) precision tank of (useful) length $l = 6.27\text{ m}$, width $w = 0.25\text{ m}$, and water depth $h = 0.330\text{ m}$. Upon release, beads are moving down a $\theta = 35$ degree slope. (a) Longitudinal cross section with marked location of sluice gate and 4 wave gages (WG1, WG2, WG3, WG4). (b,c) Zoom-in on side- and cross-section views of slope and sluice gate (dimensions marked in mm). (d) Picture of experimental set-up around slope and sluice gate.

2.3 Experimental validation of deforming slide models

2.3.1 Description of laboratory experiments and results

Laboratory experiments of tsunami generation by underwater slides made of glass beads were performed at the Ecole Centrale de Marseille (IRPHE), France, in a precision tank of (useful) length $l = 6.27$ m and width $w = 0.25$ m (Fig. 2.3). In each experiment, a Mass W_b of beads of density $\rho_b = 2,500$ kg/m³ was submerged in fresh water of density $\rho_w = 1,000$ kg/m³, in a reservoir of triangular shape located over a $\theta = 35$ degree slope, fronted by a movable sluice gate. Experiments were started by withdrawing the gate into a bottom cavity within the slope (using a highly repeatable motion controlled by springs, visible in Fig. 2.3d). Fifty eight experiments were performed with each experiment being repeated twice (i.e., there were 29 individual sets of parameters). Independent parameters in experiments were the tank water depth, set to $h = 0.320$ to 0.370 m, the glass bead diameter, set to either $d_b = 4$ or 10 mm and their total dry weight, set to $W_b = 1.5$ to 2.5 kg. In 20 experiments, a thin layer of glass beads was glued on the slope while glass beads were free to slide in other experiments. Each experiment was recorded using a high speed video camera (at 1,000 frames per sec) set on the side of the tank, showing both the underwater motion of the beads and the resulting free surface deformation (Fig. 2.4). Additionally, time series of free surface elevations were measured at 4 wave gages WG1-WG4 (Fig. 2.3a), $\eta_i(t)$ ($i = 1, 2, 3, 4$).

Fig. 2.4 shows snapshots extracted from the video taken during one of the landslide tsunami experiments, with parameters: $h = 0.330$ m; $d_b = 4$ mm, $W_b = 2$ kg, and no glued beads on the slope, up to $t = 0.60$ sec (note the starting time of each experiment, $t = 0$, is defined when the gate has just withdrawn into its cavity). In Fig. 2.4a, we see that glass beads are initially stored in the triangular reservoir with the sluice gate up. The experiment is started by withdrawing the gate into a bottom cavity. Although this is quick, it takes about 0.125 s to do so

in experiments; hence, here, we define the start of each experiment, $t = 0$, when the gate has just withdrawn into its cavity. Once free to move, glass beads slide down the 35 degree slope and their motion creates an initial depression of the water surface (Fig. 2.4b), which then rebounds, creating two sets of waves moving down and up the tank (Fig. 2.4c). The “onshore” moving waves cause runup on the slope whereas the “offshore” waves reflect on the far end of the tank and propagate back towards the generation area (Fig. 2.4d-f). This behavior is also clearly observed in time series measured at wave gages WG1-WG4 (Fig. 2.5). A detailed analysis of experimental results shows that this and each experiment is highly repeatable, with almost unnoticeable differences between wave gage measurements for two replicates of the same experiment. More details of the experimental set-up and methods, and of results of the 58 experiments will be reported elsewhere.

Here, we benchmark the two models of tsunami generation by deforming slides discussed above (i.e., Ma *et al.*'s , 2015 [76] and Kirby *et al.*'s , 2016 [31] models) by simulating a single representative experiment corresponding to the case and parameters of Fig. 2.4. In this simulation, the entire tank geometry is modeled in NHWAVE, starting from the situation shown just before the stage of Fig. 2.4b when the gate has just withdrawn, defined as $t = 0$. Numerical wave gages are located in the model at the same locations as shown in Fig. 2.3a and simulated time series are compared to experimental data. This is detailed in the next section.

2.3.2 Numerical modeling of laboratory experiments

We model one laboratory experiment of tsunami generation by a deforming slide made of glass beads of density $\rho_b = 2,500 \text{ kg/m}^3$ in fresh water of density $\rho_w = 1,000 \text{ kg/m}^3$, with parameters: $h = 0.330 \text{ m}$; $d_b = 4 \text{ mm}$, $W_b = 2 \text{ kg}$, and no glued beads on the slope (Fig. 2.4), using the two-layer models detailed above, i.e., with either one dense Newtonian fluid sublayer [31] or a saturated granular

Figure 2.4. Snapshots of laboratory experiments of tsunami generation by underwater slides made of glass beads (Fig. 3), for $h = 0.330$ m; $d_b = 4$ mm, $W_b = 2$ kg, and no glued beads on the slope, at times $t =$ (a) -0.105 ; (b) 0.02 ; (c) 0.17 ; (d) 0.32 ; (e) 0.47 ; and (f) 0.62 sec. Note, glass beads are initially stored within the glass bead reservoir with the sluice gate up; at later times, after the gate is withdrawn, the deforming slide moves down the 35° slope while the free surface is deformed. The starting time of experiments $t = 0$ is defined when the gate has just withdrawn into its cavity.

medium sublayer [76]. To apply the latter model, we simply define a granular material made of spherical grains of density and diameter identical to the glass beads used in experiments. To apply the former model, however, we first need to estimate the density of a dense fluid equivalent to the initial mixture of glass beads and water. Also, because only the diameter and total dry weight of glass beads was measured in experiments, for both models, we need to compute the location of the resting interface between slide and water (Fig. 2.4a), i.e., the slide initial submergence h_s , whose value significantly affects tsunami generation (Grilli and Watts, [13]). Finally, for both models, we need to specify realistic values of the Manning bottom friction coefficient n and the parameter controlling the volumetric energy dissipation. For the dense fluid, the latter is the dynamic viscosity μ and for the granular medium it is the internal Coulomb friction angle φ_{int} and dynamic bed friction angle φ_{bed} , plus a calibration parameter λ controlling the pore pressure and friction near the bed. As detailed below, results from theoretical and experimental work on granular media will be used to estimate the dense fluid dynamic viscosity and the granular, bed and internal, friction angles will be based on recommended values from earlier experimental work, which will then be slightly adjusted. Finally, for both models, the Manning coefficient will be calibrated for the first generated wave to match observations.

While the maximum regular packing of spheres is $\phi_m = \pi/(3\sqrt{2}) = 74.0\%$ of the volume occupied by the spheres, studies of randomly packed spheres show that the packing value is typically near (and bounded by) $\phi = 63.4\%$ [82]. Here, we will assume this value, meaning that for a saturated medium, $1 - \phi = 36.6\%$ of voids are filled with water. Hence, the equivalent fluid density for modeling the slide as a dense fluid is found as: $\rho_s = (1 - \phi)\rho_w + \phi\rho_b = 1,951 \text{ kg/m}^3$. Slide volume is then found as, $V_s = W_s/(\phi\rho_b) = 1.262 \times 10^{-3} \text{ m}^3$, and the initial front

thickness of the slide is derived from the triangular geometry of the glass bead reservoir (Figs. 3 and 4a) as, $T = \sqrt{2 \tan \theta V_s / w} = 0.084$ m. Based on dimensions shown in Fig.3b, we finally find the initial slide submergence for the considered experiment as, $h_s = h - T - 0.291 \tan \theta = 0.0422$ m.

Modeling the deforming slide as a dense fluid of density ρ_s requires estimating relevant values of the fluid's dynamic viscosity μ_s and Manning coefficient n for the slide-substrate bottom friction. Assuming the suspension of water with glass beads still behaves as a Newtonian fluid, expressions have been proposed in earlier work for the ratio of μ_s to the suspending fluid viscosity μ , $\eta_r = \mu_s / \mu$, assuming a high concentration of particles (here the glass beads).

$$\eta_r = \left(1 + \frac{1.25\phi}{(1 - \frac{\phi}{\phi_m})}\right)^2, \quad \frac{9}{8} \left(\frac{(\frac{\phi}{\phi_m})^{\frac{1}{3}}}{1 - (\frac{\phi}{\phi_m})^{\frac{1}{3}}}\right), \text{ and } \left(1 - \frac{\phi}{\phi_m}\right)^{-2} \quad (2.2)$$

based on Eilers (1941) [83], Frankel and Acrivos (1967) [84], and Quemada (1977) [85], respectively. [For details regarding theoretical developments and experimental validations of these expressions, see Mendoza and Santamaria-Holek (2009) [86], and Mueller *et al.* (2010) [87]]. Hence, for the above values of ϕ and ϕ_m , we find based on Eqs. 2.2, $\eta_r = 42.7, 21.3$ and 48.9 , respectively. Assuming the dynamic viscosity of fresh water is $\mu = 0.001$ kg/(m.s), these equations predict μ_s values in the range $[0.0213 - 0.0489]$ kg/(m.s). Just after the initiation of motion (e.g., Fig. 2.4b), the slide still nearly has the maximum random packing density ϕ and, hence, using a viscosity $\mu_s \in [0.0213 - 0.0489]$ kg/(m.s) is a reasonable assumption. However, as the slide deforms and moves down the slope (Fig. 2.4c-d), the glass bead volume fraction decreases, which based on Eqs. 2.2 causes the equivalent slide density and viscosity to decrease. To account for this, in the dense fluid model, we will be using a lower value of viscosity, $\mu_s = 0.01$ kg/(m.s). Additionally, using a lower value of viscosity can help compensate for the nearly free slip of glass beads

along the slope and the low bottom friction that results, when glass beads are not glued to the slope. A lower viscosity can also enhance the slide initial acceleration, which as indicated above, may be underestimated in the model Eq.2.1. For this value of viscosity, a Manning coefficient value $n = 0.04$ was calibrated so that the modeled slide reach the bottom of the slope at the time measured in experiments (Fig. 2.4).

The selected glass bead experiment is modeled with NHWAVE using the initial slide geometry computed above (T, h_s) and the selected parameters, ρ_s , μ_s and n , for the dense fluid representing the slide. A sensitivity to grid resolution detailed later shows that results are well converged when using a horizontal resolution with grid size $\Delta x = \Delta y = 0.01$ m and 9 σ -layers in the vertical direction. Fig. 2.5 shows a comparison of time series of surface elevations computed in this grid to those observed at the four wave gages WG1-WG4 (Fig. 2.3a). Overall, there is a good agreement of model results with the measurements, and more so for the incident wave train than for reflected waves, later in the time series, where discrepancies slightly increase, perhaps due to effects of far-end wall roughness in experiments, not represented in the model. More specifically, at gage WG1, the model reproduces well the 2 leading waves (i.e., first two crests and first trough), both incident and reflected, but shows increasing discrepancies for the smaller trailing waves; for the two leading incident waves, the Root Mean Square (RMS) difference of the modeled $\eta_m(t)$ and experimental $\eta_e(t)$ surface elevations, scaled by the maximum wave height H_{max} measured at the gage (trough to crest), $\epsilon_{RMS} = \sqrt{\sum(\eta_m - \eta_e)^2}/H_{max} = 7.2$ %. At gages WG2-WG4, due to dispersion, the two leading waves evolve into an increasing number of larger waves, which are all well simulated in the model (incident or reflected). We note that both the elevation and phase of waves are well simulated, indicating that dissipations in the

Figure 2.5. Comparison of observed (blue) time series of surface elevation at wave gages WG1 to WG4 (Fig. 3a), from top to bottom, to those computed (red) using NHWAVE, with a dense Newtonian fluid layer underneath with: $\rho_s = 1,951 \text{ kg/m}^3$, $\mu_s = 0.01 \text{ kg/(m.s)}$ and $n = 0.04$. The model grid uses $\Delta x = \Delta y = 0.01 \text{ m}$ horizontally and 9 σ -layers in the vertical direction. Note, the origin of the time axis corresponds to the arrival of the first elevation wave at gage WG1.

slide and frequency dispersion effects during propagation are adequately modeled. The RMS error for the first three and four waves observed at gages WG2 and WG3, and for the entire time series at gage WG4, is $\epsilon_{RMS} = 7.0, 6.6$, and 8.4% , respectively.

Simulations of this experiment were repeated to study the sensitivity of model results to grid resolution. Thus, Fig. 2.6 shows results using $\Delta x = \Delta y = 0.01 \text{ m}$ in the horizontal grid, with the number of σ -layers being 3, 6, or 9. The simulation with 3 σ -layers yields slightly lower values of crest and trough elevations, whereas results are nearly identical using 6 or 9 σ -layers; 9 σ -layers will be used from now on. Fig. 2.7 shows results using 9 σ -layers, with the horizontal grid resolution set to $\Delta x = \Delta y = 0.005, 0.01$, or 0.02 m . A significant difference is observed when reducing the horizontal grid resolution from 0.02 to 0.01 m , and a much smaller difference is observed when further reducing the grid step to 0.005 m . Hence, results are deemed to have nearly converged for $\Delta x = \Delta y = 0.01 \text{ m}$ and, for more

Figure 2.6. Case of Fig. 2.5. Sensitivity of surface elevations computed at wave gages for $\Delta x = \Delta y = 0.01$ m horizontally, to the number of σ -layers : 3 (green), 6 (black), and 9 (red), as compared to experimental data (blue).

efficiency, this resolution will be used from now on.

Fig. 2.8 compares the slide cross-sections computed at four times $t =$ (a) 0.02; (b) 0.17; (c) 0.32; and (s) 0.47 sec, to those observed in experiments with the high speed camera. The computed flow velocity module is also plotted in the figures. A reasonable agreement is observed between the computed and measured profiles, but clearly, the depth-integrated slide model cannot reproduce the bulbous shape, associated with a recirculating flow within the slide, that gradually develops at the front of the slide, as time increases. Also, Figs. 2.8a,b show large changes in flow velocity over a small horizontal distance, between the near-initial stage at $t = 0.02$ sec, and $t = 0.17$ sec, indicating that the flow acceleration down the slope is large, with likely large associated vertical accelerations due to the steep slope. As indicated before, the latter are neglected in the current model Eq. 2.1 and, although this does not appear to significantly affect the modeling of tsunami generation (as inferred from the good agreement in Fig. 2.5 of surface elevations at wave gages), their inclusion in the equations might further improve the agreement of the modeled and observed slide cross-sections.

Figure 2.7. Case of Fig. 2.5. Sensitivity of surface elevations computed at wave gages for 9 σ -layers, to the horizontal grid resolution, $\Delta x = \Delta y = 0.02$ (green), 0.01 (red), and 0.005 (black) m, as compared to experimental data (blue).

The same experiment was finally simulated using Ma *et al.*'s (2015) [76] two-layer granular-slide model, using the same grid discretization in NHWAVE as in the converged case with the first model (i.e., $\Delta x = \Delta y = 0.01$ m and 9 σ -layers). As this model is still being developed and not yet applicable to arbitrary topography (and thus practical tsunami hazard assessment), the goal at this stage was only to verify that this model with more complete physics can also simulate the glass bead experiments. Values of the granular friction parameters in the model were first selected similar to those used by Watts and Grilli (2003) [75], who modeled 2D experiments with 3 mm diameter glass beads using the Bingham model BING, i.e., $\varphi_{int} = 29$ deg. and $\varphi_{bed} = 6$ deg. Then, to improve the agreement of the first generated wave with laboratory measurements, the bed friction was slightly adjusted to: $\varphi_{bed} = 9$ deg., together with using $\lambda = 0.04$ (note that a larger value of $\lambda \in [0, 1]$ will result in a reduced bed Coulomb friction). Finally, the (water on slide or substrate) friction coefficient was set to $C_f = 0.6$. This is a fairly large value, but it allows accounting for sidewall friction in experiments; results will show that with this model, the oscillatory tail in the wave train is still too large as

Figure 2.8. Case of Fig. 2.5. Comparison of computed slide cross-sections (top) with those observed in experiments with the high speed camera (bottom), at four times $t =$ (a) 0.02; (b) 0.17; (c) 0.32; and (d) 0.47 s. Color scales are computed flow velocity modules in m/s.

compared to experiments, indicating that bottom friction might need to be further increase.

Surface elevations calculated at the four gage locations with the granular slide model are shown in Fig. 2.9 and compared to experiments, as well as to results of the viscous slide model (case of Fig. 2.5). We see that the first two waves generated by the granular slide model are in reasonable agreement with experimental results (particularly at gages WG2 to WG4), but waves appear to be slightly longer, shed a larger oscillatory tail, and decay slower than waves observed in experiments and modeled with the viscous fluid model; they also have a slightly higher phase velocity. These differences with the viscous fluid model are likely due to insufficient energy dissipation in the granular slide, but in view of the more complex governing equations, more work is needed to clearly identify the causes and how model parameters should be adjusted. Additionally, during the early stages of slide motion, unlike the viscous slide model which has no slip boundary conditions, the granular slide model lateral boundary conditions cause a small amount of friction at the edges of the domain. This creates a small variation in slide motion and tsunami forcing across the width of the tank, which leads to some lateral spreading in the resulting waves. Nonetheless, as indicated the granular model produces surface elevations, which agree reasonably well with experimental results.

2.3.3 Sensitivity of tsunami generation to slide rheology

When modeling SMFs as deforming slides represented by a dense Newtonian fluid, besides density, two main parameters control the slide center of mass motion $S(t)$ down the slope (measured parallel to the slope): the equivalent slide viscosity μ_s and Manning coefficient n . When setting these parameters to larger values, this increases the frictional dissipation and resulting forces that are slowing down slide motion in the model. Hence, when μ_s or n are increased, we expect that slide

Figure 2.9. Case of Fig. 2.5. Comparison of observed (blue) time series of surface elevation at wave gages WG1 to WG4 (Fig. 3a), from top to bottom, to those computed using NHWAVE's viscous model (red), and granular flow [76] model (black).

Figure 2.10. Case of Fig. 2.5. Sensitivity of surface elevations computed at gages WG1-WG4 (Fig. 2.3a; right panels from top to bottom), as compared to experimental measurements, to the equivalent slide viscosity $\mu_s = 0.01$ (red), 0.1 (green), and 1 (black) $\text{kg}/(\text{m.s})$, with $n = 0.04$. The leftward panels show the corresponding slide center of mass motion $S(t)$, velocity $U(t)$, and acceleration $A(t)$. Note, zero time for left panels is when the sluice gate has fully retracted, whereas for right panels it is when the first elevation wave reaches gage WG1.

acceleration and velocity will be reduced, once the slide picks up speed and these forces become effective. For the initial part of slide motion down the slope, however, slide velocity is small and the motion should be close to be a purely accelerating motion, $S(t) \sim (1/2)A_o^2(t)$, with A_o the initial slope acceleration down the slope [13, 14]; hence one should not observe significant differences in slide motion. Since slide velocity and acceleration are reduced when μ_s or n are increased, one would also expect that tsunami generation be reduced.

These predictions are confirmed in Figs. 2.10 and 2.11, which show results of simulations corresponding to the case of Fig. 2.5, in which we first set slide

viscosity to $\mu_s = 1, 0.1, \text{ and } 0.01 \text{ kg/(m.s)}$, with $n = 0.04$, and then set the slide-substrate Manning coefficient to $n = 0.01, 0.04 \text{ and } 0.07$, with $\mu_s = 0.01 \text{ kg/(m.s)}$. In both figures, the computed slide center of mass motion $S(t)$, velocity $U(t) = dS/dt$ and acceleration $A(t) = dU/dt$ are plotted, together with the free surface elevations computed at gages WG1-WG4 (Fig. 2.3a), as compared to the experimental measurements. As expected, in both figures, for short times, $t < 0.1$ sec or so, slide kinematics is not affected by the value of μ_s or n , consistent with this being a purely accelerating motion. For later times, however, while the slide bulbous front is still traveling down the slope (i.e., for $t < 0.5$ sec or so; Fig. 2.4), both slide center of mass velocity and acceleration, and thus distance traveled by the slide on the slope in a given time, are reduced when μ_s or n are increased. Once the slide is on the flat bottom of the tank, this behavior may be slightly more complex. Results are also as expected regarding the free surface elevations modeled at wave gages, with a reduction in tsunami generation, as μ_s or n are increased.

2.3.4 Sensitivity of tsunami generation to slide submergence

Earlier work indicates that besides slide volume/geometry, density, and rheology, one important parameter that affects tsunami generation is the initial slide submergence depth h_s [13, 14, 17], with tsunami wave elevation increasing, the smaller the submergence depth. This is also observed in the full set of laboratory experiments with glass beads, which is not detailed here due to lack of space. In the model, this is verified in Fig. 2.12, where computations for the experimental case of Fig. 2.5 have been repeated for a lower and higher submergence, i.e., $h_s = 0.022 \text{ and } 0.062 \text{ m}$, for the initial same location of the slide on the slope, by setting the water depth to, $h = 0.31 \text{ and } 0.35 \text{ m}$, respectively, with $\mu_s = 0.01 \text{ kg/(m.s)}$ and $n = 0.04$. The figure shows, as expected, that for the shallower

Figure 2.11. Case of Fig. 2.5. Sensitivity of surface elevations computed at gages WG1-WG4 (Fig. 2.3a; right panels from top to bottom), as compared to experimental measurements, to the slide-substrate Manning coefficient $n = 0.01$ (red), 0.04 (green), and 0.07 (black) $\text{kg}/(\text{m.s})$, with $\mu_s = 0.01 \text{ kg}/(\text{m.s})$. The leftward panels show the corresponding slide center of mass motion $S(t)$, velocity $U(t)$, and acceleration $A(t)$. Note, zero time for left panels is when the sluice gate has fully retracted, whereas for right panels it is when the first elevation wave reaches gage WG1.

Figure 2.12. Case of Fig. 2.5. Sensitivity of surface elevations computed at gages WG1-WG4 (Fig. 3a; right panels from top to bottom), as compared to experimental measurements (blue), to the slide initial submergence depth, $h_s = 0.022$ (green), 0.042 (red), and 0.062 (black) m, for a water depth $h = 0.31$ (green), 0.33 (red), and 0.35 (black) m, respectively, with $\mu_s = 0.01$ kg/(m.s) and $n = 0.04$. Time is given in seconds after the initial crest reaches the first gage.

submergence depth the largest leading trough and crest are significantly increased, as compared to the based case, particularly upon generation at gage WG1. For a larger submergence depth, the trend is opposite but by a lesser amount, indicating that nonlinear effects are at play. At gages WG2-WG4, the same trends are preserved but as dispersion distributes the change in the leading wave elevation to the training waves in the oscillatory tail, the effect of submergence depth is less marked than at the first gage.

2.4 Case studies off of the US East Coast

In the previous section, on the basis of laboratory experiments, we assessed the accuracy of two models simulating tsunami generation by deforming underwater slides, as well as the sensitivity of the dense fluid model results to parameters such as the equivalent viscosity μ_s and the slide-substrate Manning friction coefficient n . In the following we apply the viscous fluid model to the simulation of tsunami generation by SMFs located off of the USEC, on the continental shelf break and slope, and evaluate the effect of slide rheology on coastal tsunami hazard, as compared to earlier simulations in which the SMFs were modeled as rigid slumps [10]. As before, we will limit ourselves to the upper USEC where SMF tsunami hazard was deemed to be higher [68] and consider SMFs with the characteristics of the historical Currituck slide, i.e., Currituck SMF proxies, located in area 1 in Fig. 2.1.2. Before this simulation, however, we revisit the simulation of the historical Currituck slide (which was modeled by Grill *et al.* [10] as a rigid slump) by applying the deforming fluid-like slide model, with the goal of estimating relevant values of μ_s and n , that provide a motion of the slide center of mass $S(t)$, similar to that of the earlier slump simulations, and thus a similar runout S_f after a main time of motion t_f . Simulations of the Currituck SMF proxy located in area 1, in the Hudson River canyon, are then performed using these parameters

Figure 2.13. Currituck slide complex (Fig. 2.1.2). (a) High-resolution bathymetry around the site in post-failed (current) conditions (the two headwalls identified by Locat *et al.* [20] are clearly visible). (b) Zoom-in around initial slide location and geometry ($V_s = 134 \text{ km}^3$; maximum thickness $T = 750 \text{ m}$, downslope length $b = 30 \text{ km}$, and width $w = 20 \text{ km}$; density $\rho_s = 1,900 \text{ kg/m}^3$), with the center of the pre-failed volume located at (-74.7E , 36.4N), in the unfailed recreated bathymetry. (c) Final location of rigid slump[10], for $t_f = 11.9 \text{ min}$. Depth (< 0) is marked in meters and time of 5 sec marked in panel (b) correspond to the first time step in NHWAVE.

Figure 2.14. Simulated kinematics of Currituck SMF (Fig. 2.13). Time variation of center of mass motion S , velocity U , and acceleration A , for a deforming fluid-like slide, with $\mu_s = 500 \text{ kg}/(\text{m}\cdot\text{s})$ and a Manning friction coefficient $n = 0.05$ (black), 0.10 (red), and 0.15 (green), as compared to Grilli *et al.*'s [10] rigid slump kinematics (blue).

Grid	SW Lat. (N deg.)	NE Lat. (N deg.)	SW Lon. (W deg.)	NE Lon. (W deg.)	Res.	N_x	N_y
CT	35.430	36.559	74.900	72.7126	500 m	393	256
G1	37.401	40.993	74.464	69.838	1,000 m	400	400
G2	40.003	40.967	74.437	72.245	154 m	1,188	700

Table 2.1. Parameters of Cartesian horizontal grids (Figs. 2.1.2 and 2.16) used in NHWAVE (CT, G1) and FUNWAVE-TVD (G1, G2) models to compute tsunami generation by SMFs and propagation to the coast. “Res.” is grid resolution and N_x and N_y indicate the number of grid cells in each direction.

and the generated tsunami and coastal impact are compared to those previously obtained by modeling the SMF as a rigid slump.

2.4.1 Currituck slide complex

As mentioned above, we simulate the historical Currituck slide using the deforming fluid-like slide model, in order to derive realistic values for the slide equivalent viscosity μ_s and Manning coefficient n , that yield a law of motion of the slide center of mass similar to that of the rigid slumps used earlier. Hence, we focus here on slide kinematics, rather than tsunami generation, which is also computed in the model but will not be detailed here.

Fig. 2.1.2 shows the area of the Currituck slide complex, on the continental

slope off of Chesapeake Bay. Grilli *et al.* [10] model tsunami generation by the Currituck slide complex using NHWAVE, representing it as a $V_s = 134 \text{ km}^3$ rigid slump (i.e., within the range of the total volume estimated by Locat *et al.* [20] for the failed area, 128-165 km^3), moving over a recreated unfailed bathymetry. Although the failure clearly occurred in two stages, as inferred from the two existing headwalls (Fig. 2.13a), the initial slump geometry was idealized as a single sediment mount with quasi-Gaussian cross-sections and an elliptical footprint on the seafloor [17]; maximum thickness was $T = 750 \text{ m}$, downslope length $b = 30 \text{ km}$, and width $w = 20 \text{ km}$, with the ellipse centered at (-74.7E, 36.4N), in water depth $h \simeq 1,800 \text{ m}$ (Fig. 2.13b). Assuming a bulk density $\rho_s = 1,900 \text{ kg/m}^3$, the slump kinematics was specified as a bottom boundary condition in the model, based on the analytical law of motion developed by Grilli and Watts [13], using the maximum velocity $U_{max} = 35 \text{ m/s}$ proposed by Locat *et al.* [20] based on independent simulations. With these parameters the slump time of motion were found as, $t_f = 11.9 \text{ min}$ and runout $S_f = 15.8 \text{ km}$ (Fig. 2.13c).

Here, we repeat these simulations for the same unfailed bathymetry and geometric representation of the SMF as in Grilli *et al.* [10] (Fig. 2.13b), but modeling its kinematics instead as a deforming fluid-like slide of density $\rho_s = 1,900 \text{ kg/m}^3$, using the NHWAVE model with a dense fluid layer underneath [31]. As shown in Figs. 2.10 and 2.11, changes in equivalent slide viscosity have a relatively smaller effect on slide kinematics and tsunami generation than changes in the bottom friction coefficient. Additionally, the analysis of field data at the Currituck slide complex [88, 20] indicated that it had more likely failed as a high viscosity debris flow. Accordingly, and following Jiang and Leblond [78], we selected a fairly high equivalent slide viscosity $\mu_s = 500 \text{ kg/(m.s)}$ corresponding to this type of rheology. We then varied the bottom friction coefficient n and observed its effect on

the deforming slide kinematics; in the simulations we used a horizontal grid with $\Delta x = \Delta y = 500$ m, approximately covering the area shown in Fig. 2.1.2 (see details in Table 2.1), and 5 σ -layers in the vertical direction. Note that this grid covers a fairly small area, extending more to the SE direction where the deforming slide is moving and tsunami generation occurs, which may not be sufficient to perform actual tsunami hazard assessment, but is sufficient at this stage for the purpose of comparing slide kinematics between rigid and deforming SMFs.

Thus, Fig. 2.14 shows the slide center of mass motion S , velocity U and acceleration A computed during the NHWAVE simulations of the Currituck SMF, when specifying $n = 0.05, 0.10$, or 0.15 , as compared to the kinematics of the equivalent rigid slump modeled by Grilli *et al.* [10]. To properly compare tsunami generation for rigid and deforming SMFs, we selected $n = 0.1$, for which the slide motion and velocity during the majority of the failure (up to $t = 600$ s or 10 min) are closest to those of the rigid slump. This way, both the deforming and rigid SMF center of mass achieve approximately the same runout at the same time, while reaching a similar maximum velocity $U = 32 - 34$ m/s during motion. Based on these parameters, Fig. 2.15 shows snapshots of the modeled slide motion, from $t = 0 - 12$ min. We see that as the slide moves down the continental slope, it both vertically deforms and laterally spreads out, covering an increasing surface area on the seafloor. Comparing the snapshot in Fig. 2.15 at 12 min to the final slump motion in Fig. 2.13c, we see that, by contrast, the slump has just covered a pendulum-like rigid motion, while keeping its shape and footprint on the seafloor constant.

2.4.2 Hudson River canyon SMF

We simulate tsunami generation and propagation to the coast by a Currituck SMF proxy located in Area 1 of Fig. 2.1.2, in the Hudson River Canyon, and

Figure 2.15. Currituck slide complex (Fig. 2.13). Snapshots of slide motion modeled using NHWAVE, with a dense fluid layer underneath with $\mu_s = 500 \text{ kg}/(\text{m.s})$ and a Manning friction coefficient $n = 0.10$. Time of snapshots is marked in min:sec.

Figure 2.16. Simulations of tsunami generation by a Currituck SMF proxy located in the Hudson River Canyon (area 1 in Fig. 2.1.2). (a) Boundary (red lines) of nested Cartesian computational grids used in simulations: G1 (1000 m resolution) and G2 (154 m resolution) (see details in Table 1). Tsunami generation is performed in grid G1 using NHWAVE with 5 σ -layers in the vertical direction, and results are used as initial condition in FUNWAVE-TVD, also for grid G1. The thick yellow line marks the 5 m depth isobath along which tsunami elevation is computed in Fig. 2.19. (b) Zoom-in of SMF location, with the Hudson River canyon to the east of it. Color scale is bathymetry (< 0) and topography (> 0) in meters. Red ellipse marks SMF footprint.

evaluate the effect of slide rheology on coastal tsunami hazard. The SMF has the same volume as the Currituck slide, $V_s = 134 \text{ km}^3$ and a quasi-Gaussian initial geometry of thickness $T = 750 \text{ m}$ and elliptical footprint of downslope length $b = 30 \text{ km}$ and width $w = 20 \text{ km}$ (see footprint in Fig. 2.16). Because this is a hypothetical SMF occurring for an existing slope, upon failure, the SMF is initially located below the seafloor (by contrast with the actual Currituck paleoslide whose initial geometry was recreated over the current seafloor). We simulate the SMF motion and related tsunami generation with NHWAVE, assuming either that it behaves as a rigid slump (as in [10]) or as a deforming slide. Once tsunami generation is completed in NHWAVE, simulations are initialized and continued with FUNWAVE-TVD.

In the rigid slump simulations, the SMF moves in the assumed azimuthal direction of motion $\theta = 136 \text{ deg.}$ (from North), and its geometry and kinematics are computed as detailed in [10] and specified as a bottom boundary condition in NHWAVE. During this simulation, as in [10], bottom friction between water and substrate is neglected, which is acceptable considering the large water depth. In the deforming slide simulations, we use the new NHWAVE two-layer model, in which the bottom layer is a dense fluid with material parameters calibrated for the historical Currituck slide case (see above): $\rho_s = 1,900 \text{ kg/m}^3$, $\mu_s = 500 \text{ kg/(m.s)}$ and $n = 0.05, 0.1$ or 0.15 for the slide-substrate bottom friction. With these parameters and $n = 0.1$, earlier simulations of the historical Currituck SMF (see above) indicate that the deforming slide reaches nearly the same runout S_f as the slump at time t_f (see details above and Fig. 2.14). Although the bottom topography is different, we find that the law of motion of the Currituck SMF proxy in Area 1 is very closely that derived for the historical Currituck slide, whether modeled as a rigid slump or a deforming slide, shown in Fig. 2.14 for

these parameters (red curves); hence the Hudson Canyon SMF kinematics is not replotted here.

Fig. 2.16a shows the boundary of two nested grids used in simulations (see details in Table 2.1). Tsunami generation is computed with NHWAVE in the 1,000 m resolution Cartesian grid G1, with 5 σ -layers in the vertical direction and simulations are pursued up to $t = 20$ min when the deforming slide motion becomes negligible (the rigid SMF stops moving at $t_f = 11.9$ min; see discussion below). Beyond this time, it is assumed that the deforming slide, although still moving is no longer tsunamigenic, due to both the spreading and thinning of the sediment layer and the large depth (see Figs. 2.17b and 2.18b, discussed later). For $t > 20$ min, computations are carried out with NHWAVE, first in grid G1 and then by one-way coupling in the finer, 154 m resolution, Cartesian grid G2; in all simulations with FUNWAVE-TVD, a Manning coefficient $n = 0.025$ (i.e., the value for coarse sand) is used to define bottom friction.

Fig. 2.17 shows results of NHWAVE simulations for slide motion and wave generation at time $t = 13.3$ min (800 s), for both the rigid slump and the deforming slide, i.e., shortly after the slump has stopped moving. At this time, whereas the rigid slump has moved in the specified downslope direction (transect in Fig. 2.18a) while preserving its shape, the deforming slide has flowed asymmetrically, following the steepest slope on the seafloor, while spreading in all directions (Fig. 2.17a,b). Fig. 2.18b shows a downslope transect through the time dependent bathymetry of both the rigid slump and the deforming slide; here, we see as expected that both SMFs achieve approximately the same runout when the slump stops moving (at $t = 714$ s). However, as a result of lateral spreading, the deforming slide thickness over the seafloor gradually reduces along the transect. Also note that at $t = 300$ s, both SMFs approximately have the same cross-section along their axis, indicating

Figure 2.17. NHWAVE simulations of tsunami generation by Currituck SMF proxy in Area 1 (Fig. 2.1.2) in the Hudson River Canyon (Fig. 2.16), modeled as a: (a,c) rigid slump [10]; (b,d) deforming slide represented as a dense fluid layer (with $\mu_s = 500 \text{ kg/(m.s)}$ and $n = 0.1$), with same initial geometry, location, volume, density $\rho_s = 1,900 \text{ kg/m}^3$ and runout at the time the slump stops moving ($t_f = 11.9 \text{ min}$). Panels (a,b) show in gray the SMF locations after 13.3 min (to the left of the Hudson River canyon), and panels (c,d) show the surface waves generated after 13.3 min (the black ellipses mark the initial footprint of the SMFs).

Figure 2.18. Case of Fig. 2.16. Downslope ($ξ$; direction $θ = 136$ deg. from North) and cross-slope ($ζ$) transects (marked in panel (a); centered on the SMFs' initial location), through: (b) bottom topography and SMFs; and (c,d) tsunami waves generated by a rigid slump (blue) and a deforming slide (red), simulated with NHWAVE. Times in panels (b-d) are $t = 300$ (solid), 600 (chained) and 800 (dashed) sec.

that lateral spreading of the deforming slide is not very significant at small time, likely as a result of the high slide viscosity. Despite the difference in kinematics and geometry, the patterns of generated tsunami waves shown in Figs. 2.17c,d at $t = 13.3$ min are similar for both SMFs. As a result of variations in thickness over the seafloor, there is a marked asymmetry in wave generation for the deforming slide, with larger waves generated on its western side; this is further discussed below. Overall, results show that onshore moving waves have lower elevations in the deforming slide case.

Differences in generated waves for both SMFs are more apparent in the transects of Fig. 2.18 through free surface elevations computed with NHWAVE, downslope (Fig. 2.18c; ξ direction, $\theta = 136$ deg. from North) and cross-slope (Fig. 2.18d; ζ direction), at times $t = 300, 600$ and 800 s. At $t = 300$ s, in both cases, the free surface is essentially a large two-pronged depression with a large elevation wave that has propagated offshore. The depression in the free surface, which is due to the draw down caused by the large volume of sediment moving downslope, has a larger offshore maximum trough for the rigid slump (60 m deep) than for the deforming slide (about 28 m deep), and a second onshore depression (nearly 50 m) of closely identical shape in both cases. For both SMFs, at $t = 300$ s, a rebound wave has already formed in the middle of the depression. Fig. 2.18c shows that, at $t = 600$ and 800 s, this rebound wave propagates onshore as the leading elevation wave of both SMF tsunami wave trains, preceded by a depression wave originated from the initial draw down. At these times, for both SMFs, both the offshore elevation wave and trough propagate faster offshore (due to the increase in phase velocity over the deep ocean), while decreasing in elevation; Figs. 2.17c,d show that this decrease is a result of the 2D horizontal energy spreading, also visible in the cross-slope transects of Fig. 2.18d. Additionally, the latter show that, while

waves generated by the rigid slump are fairly symmetric with respect to the SMF axis of motion ($\zeta = 0$), those generated by the deforming slide are larger on its western side ($\zeta < 0$) than on its eastern side, and the maximum initial free surface depression shown at $t = 300$ s is also shifted westward, whereas it is on the SMF axis for the rigid slump. As discussed above, this is a result of the asymmetry of the deforming slide thickness on the seafloor.

As indicated, once the tsunamis are generated with NHWAVE, simulations are initialized in grid G1 in FUNWAVE-TVD, by specifying surface elevation and horizontal velocity interpolated at the relevant depth ($0.531h$; [18]). Simulations are performed in grid G1 for 3 h 8' of tsunami propagation and coastal impact (note that sponge layers are specified along the grid boundary) and time series of surface elevations are computed along the boundary of the 154 m resolution nested grid G2 to be used in the one-way coupling methodology. Simulations are then restarted in grid G2 using the time series as boundary conditions, and surface elevations are computed along the 5 m isobath marked in Fig. 2.16a. Following Grilli et al. [10], computations were first initialized in FUNWAVE-TVD at $t = 13.3$ min (800 s), based on surface elevation and horizontal velocity (interpolated at 0.531 times the local depth) computed with NHWAVE. However, while this was acceptable for the slump, which by this time has stopped moving, the deforming slide was still significantly moving at $t = 800$ s and, while it was no longer tsunamigenic, initializing FUNWAVE-TVD at this time caused the appearance of a large spurious rebound wave, likely due to inconsistencies in bottom boundary conditions. Hence, simulations with NHWAVE were pursued until $t = 1,200$ s (20 min) before initializing FUNWAVE-TVD, which prevented the appearance of the spurious wave in the deforming slide case; for consistency, initialization was done at the same time for the rigid slump case as well. It should be pointed out that, similar

Figure 2.19. Case of Figs. 2.16 to 2.18. Envelope of maximum surface elevation (color scale in meters) after 3 h 8 min, computed with FUNWAVE-TVD (initialized at $t = 1,200$ s with NHWAVE results; $n = 0.025$) during propagation of SMF tsunamis generated by: (a) a rigid slump, and (b) a deforming slide with $n = 0.10$ (bathymetric contours are shown in meters). (c,d) Maximum η_{max} and minimum η_{min} surface elevations computed in grid G2 along the 5 m isobath (yellow line marked in (a) and (b)) for: (thick blue) the tsunamis generated by a rigid slump, and three different deforming slides, with $n = 0.05$ (black), $n = 0.10$ (red), and $n = 0.15$ (green), for the slide-substrate bottom friction; the curvilinear distance s along the 5 m contour is computed from its southern end; the region of lower maximum surface elevations (210 to 260 km) corresponds to the Hudson River Estuary complex, with the New Jersey shore to the south (left), and Long Island shore to the north (right).

to Grilli *et al.* (2015a) [10], for sake of computational efficiency, before simulations were initialized in FUNWAVE-TVD the offshore propagating waves were filtered out, allowing using a smaller domain in both models and narrower sponge layers in FUNWAVE-TVD, that do not significantly interfere with the coastline of interest. This was done to the right of a diagonal line that is visible on both Figs. 2.19a,b. It was verified that this did not affect results in the area of interest in grid G2.

Figs. 2.19a,b show envelopes of maximum surface elevations η_{max} computed in grids G1 and G2 during simulations with FUNWAVE-TVD of the propagation of tsunamis generated by the rigid slump and a deforming slide with $n = 0.01$, respectively. While patterns and directionality of the maximum envelopes are similar, the westward increase in wave generation is apparent for the deforming slide case. Overall, the maximum height of waves generated by the rigid slump is larger (over 25 m near the source; also see Figs. 2.18c,d) than for the deforming slide (up to 12 m near the source). As waves propagate onshore over the wide shallow shelf, in both cases, surface elevations significantly decrease and are similarly modulated alongshore, as a result of bottom friction dissipation over the wide shallow shelf and bathymetric control on wave focusing/defocusing, respectively [67].

This is detailed in Fig. 2.19c where, as in [10] who compared tsunami elevations from various sources along the USEC, the alongshore variation of maximum wave elevation η_{max} computed in grid G2 for both SMFs has been plotted as a function of the curvilinear distance s along the 5 m isobath marked by a yellow line in Figs. 2.16 and 2.19a,b. The middle section of this isobath (centered around 225 km) is within the Hudson River Estuary complex where, due to bathymetric defocusing, tsunami elevations are significantly lower than those impacting the coasts of Long Island (> 260 km) and New Jersey (< 210 km) (see Shelby *et al.*, 2016 [74] for detail). Consistent with the maximum envelopes of Figs. 2.19a,b, waves caused by the rigid

slump are larger at most locations, by up to a factor of 2, as compared to waves generated by the reference deforming slide case (with $n = 0.01$). As pointed out by Tehranirad *et al.* [67], waves generated in both cases, although of different height, closely have the same pattern of alongshore variation, which is due to the bathymetric control on long wave focusing/defocusing. For comparison, the figures also shows wave elevations caused by the deforming slides with a reduced or a larger slide-substrate friction (i.e., $n = 0.05$ or 0.15), and expectedly these increase as the inverse of n . Reducing friction in the deforming slide, however, is not sufficient to cause wave elevations that are larger than those of the rigid slump, except slightly in a very small area to the right of the figure.

Finally, some critical coastal infrastructures, such as power plants, which have cold water intakes, are also severely impacted by low water conditions occurring during the initial SMF tsunami impact in the form of a large depression wave. Hence, in Fig. 2.19d, we have similarly plotted the envelope of minimum surface elevation η_{min} computed along the 5 m isobath for all SMF cases considered in Fig. 2.19c. Here we see that consistent with the larger initial depression wave it generates, the rigid slump also causes a larger drawdown at the coast, for most locations, than the deforming slide cases, except to the north (for $x > 330$ km), likely due to site-specific bathymetric effects.

2.5 Conclusions

In this work, we first validated two models simulating tsunami generation by deforming submarine mass failures (SMFs), against laboratory experiments conducted at IRPHE in a small precision wave tank. One case was used, from a large set of highly repeatable laboratory experiments performed for SMF made of glass beads moving down a steep slope. Both validated models are two-layer models in which the fluid is modeled with the non-hydrostatic 3D (σ -layer) non-hydrostatic

model NHWAVE and the SMF layer is depth-integrated and represented either as a dense Newtonian fluid [31] or a granular medium [76]. The latter model is currently limited to a plane slope and does not include dilatancy effects in the granular medium. Both models currently neglect vertical accelerations and hence should be more accurate for moderate to small slopes.

Most of the validation tests were performed for the dense fluid model, including assessing model convergence regarding both horizontal and vertical grid resolution, and sensitivity of slide motion and generated surface elevations to slide parameters; specifically, result sensitivity to slide viscosity (i.e., internal energy dissipation), bottom friction, and initial submergence. A more limited validation was conducted for the granular slide model. Overall, once calibrated, results of the viscous model were found in better agreement with time series of surface elevations measured at 4 gages, than the granular flow model, while providing a good simulation of both the geometry and kinematics of the moving slide material. In its current state of development, which has some limitations, the granular model provided a reasonable to good agreement with surface elevations while describing slide geometry and motion slightly better than with the viscous fluid approximation.

The viscous slide model, which at present is the only one that can be applied to an arbitrary bottom bathymetry, was then used to simulate the historic Currituck SMF motion, in order to determine relevant viscous slide parameters to simulate SMF tsunamis on the east coast. This historical SMF and the generated tsunami were previously modeled by Grilli *et al.* [10] as a rigid slump and here slides parameters were first adjusted based on sediment properties, as well as to match the earlier values of the Currituck SMF runout and maximum velocity, when simulated as a rigid slump. Having determined that slide behavior is more sensitive to bottom friction than viscosity at this scale, particularly on the mild continental slope,

assuming for this moderately deforming slide a large viscosity $\mu_s = 500 \text{ kg/(m.s)}$, 3 values of Manning’s n coefficient ($n = 0.05, 0.10$, and 0.15) were selected to cover a realistic range of possible deforming slide rheologies.

The same parameters were finally applied to simulate tsunami generation from a possible SMF sited near the Hudson River Canyon. As for Currituck, this SMF had also been simulated by Grilli *et al.* [10] as a rigid slump, as this was supposed to provide a conservative (i.e., worst case scenario) estimate of tsunami generation. Simulations were performed as before for 3 deforming slides and the rigid slump, and results compared; all SMFs had the same initial volume, location, and geometry. The SMFs’ center of mass motion, velocity and acceleration were compared and it was verified that the deforming slide with $n = 0.1$ achieved nearly the same runout as the rigid slump, at the time the latter stopped moving. While initial acceleration was larger in the deforming slides, both velocity and accelerations stayed larger for a longer time for the rigid slump; this, combined with the reducing thickness of the deforming slide during motion led to a smaller tsunami generation in all cases of deforming slides than for the rigid slump. Tsunami propagation was computed for the 4 generated tsunamis in two levels of nested grids, using the Boussinesq model FUNWAVE-TVD, and the maximum surface elevations was computed along a 5 m depth contour off of the coast of New Jersey and New York. From very large initial surface elevations (up to 25 m for the rigid slump), nearshore tsunami elevations were found to be significantly smaller in all cases (up to 6.5 m), due to both directional energy spreading and bottom friction over the wide shelf. At most locations, nearshore tsunami surface elevations caused by the rigid slump were found to be significantly larger (up to a factor of 2) than those caused by the 3 deforming slide cases; for those, surface elevation slightly increased when n decreased. Nearshore minimum surface elevations (tsunami drawdown) were sim-

ilarly computed and it was found, again, that the rigid slump caused the largest drawdown in most cases, except along a stretch of the coast of Long Island, likely to site specific focusing bathymetric effects.

In view of these results (see, Fig. 2.19), we conclude that, as expected from earlier work (e.g., [75, 13]), the rigid slump provides a conservative estimate of SMF tsunami impact in terms of maximum inundation/runup at the coast and, in most locations, of maximum drawdown; by contrast, using a more realistic rheology with some level of SMF deformation, in general, will reduce tsunami impact at the coast, whether maximum inundation or drawdown. This validates as conservative the tsunami hazard assessment and inundation mapping performed to date as part of NTHMP, on the basis of Currituck SMF proxies simulated as rigid slump.

Clearly the two-layer granular flow NHWAVE model features more complete and realistic physics and thus has greater potential for more accurately modeling SMF tsunami generation, in a variety of context and rheology, once it is extended to arbitrary bottom topography; as indicated, this extension is in progress. Additionally, it is planned to extend the granular layer depth-integrated governing equations to both properly include effects of vertical accelerations, which may be important over steeper slopes, and dilatancy effect. The latter was for instance included in the two-phase model recently proposed by Bouchut *et al.* (2016) [89], who indicates that when dilation occurs the fluid is sucked into the granular material, the pore pressure decreases and the friction force on the granular phase increases. In the case of contraction (the opposite of dilation), the fluid is expelled from the mixture, the pore pressure increases and the friction force decreases. Including dilatancy will thus allow simulating both volumetric and bed dissipation effects that vary with slide shape and kinematics.

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List of References

- [1] S. Grilli, M. Ioualalen, J. Asavanant, F. Shi, J. Kirby, and P. Watts, "Source constraints and model simulation of the December 26, 2004 Indian Ocean tsunami," *Journal of Waterway, Port, Coastal, and Ocean Engineering*, vol. 133, no. 6, pp. 414–428, 2007.
- [2] M. Ioualalen, J. Asavanant, N. Kaewbanjak, S. T. Grilli, J. T. Kirby, and P. Watts, "Modeling the 26th December 2004 Indian Ocean tsunami: Case study of impact in Thailand," *J. Geophys. Res.*, vol. 112, p. C07024, 2007.
- [3] S. Grilli, J. Harris, T. Tajalli-Bakhsh, T. Masterlark, C. Kyriakopoulos, J. Kirby, and F. Shi, "Numerical simulation of the 2011 Tohoku tsunami based on a new transient FEM co-seismic source: Comparison to far- and near-field observations," *Pure and Applied Geophysics*, vol. 170, pp. 1333–1359, 2013.
- [4] D. R. Tappin, S. T. Grilli, H. J. C., G. R. J., M. Timothy, K. J. T., F. Shi, G. Ma, K. Thingbaijam, and P. Maig, "Did a submarine landslide contribute to the 2011 Tohoku tsunami ?" *Marine Geology*, vol. 357, pp. 344–361, 2014.
- [5] "Japan quake's economic impact worse than first feared," reuters, published/accessed 14 April 2011. [Online]. Available: <http://www.reuters.com/article/japan-economy-idUSL3E7FC09220110412>
- [6] Y. Fujii, K. Satake, S. Sakai, M. Shinohara, and T. Kanazawa, "Tsunami source of the 2011 off the Pacific coast of Tohoku, Japan earthquake," *Earth, Planets, and Space*, vol. 63, pp. 815–820, 2011.
- [7] K. Satake, Y. Fujii, T. Harada, and Y. Namegaya, "Time and Space Distribution of Coseismic Slip of the 2011 Tohoku Earthquake as Inferred from Tsunami Waveform Data," *The Bulletin of the Seismological Society of America*, vol. 103, pp. 1473–1492, May 2013.

- [8] H. G. T. V. Wei Y., Newman A. and L. Tang, "Tsunami forecast by joint inversion of real-time tsunami waveforms and seismic or gps data: application to the tohoku 2011 tsunami," *Pure Appl Geophys*, vol. 171, no. 12, pp. 3281–3305, 2014.
- [9] F. Romano, A. Piatanesi, N. Lorito, S. and D'Agostino, K. Hirata, S. Atzori, Y. Yamazaki, and M. Cocco, "Clues from joint inversion of tsunami and geodetic data of the 2011 Tohoku-oki earthquake," *Scientific Reports*, vol. 2, no. 385, 2012. [Online]. Available: <http://dx.doi.org/10.1038/srep00385>
- [10] S. Grilli, C. O'Reilly, J. Harris, T. Tajalli-Bakhsh, B. Tehranirad, S. Banihashemi, J. Kirby, C. Baxter, T. Eggeling, G. Ma, and F. Shi, "Modeling of SMF tsunami hazard along the upper US East Coast: Detailed impact around Ocean City, MD," *Natural Hazards*, vol. 76, no. 2, pp. 705–746, 2015.
- [11] G. Ma, F. Shi, and J. Kirby, "Shock-capturing non-hydrostatic model for fully dispersive surface wave processes," *Ocean Modelling*, vol. 43–44, pp. 22–35, 2012.
- [12] S. Grilli and P. Watts, "Modeling of waves generated by a moving submerged body. applications to underwater landslides," *Engng Analys Boundary Elements*, vol. 23, pp. 645–656, 1999.
- [13] S. T. Grilli and P. Watts, "Tsunami generation by submarine mass failure Part I : Modeling, experimental validation, and sensitivity analysis," *Journal of Waterway Port Coastal and Ocean Engineering*, vol. 131, no. 6, pp. 283–297, 2005.
- [14] P. Watts, S. T. Grilli, D. R. Tappin, and G. J. Fryer, "Tsunami generation by submarine mass failure. ii: predictive equations and case studies," *Journal of Waterway, Port, Coastal, and Ocean Engineering*, vol. 131, no. 6, pp. 298–310, 2005. [Online]. Available: <http://nora.nerc.ac.uk/504819/>
- [15] F. Enet, S. Grilli, and P. Watts, "Laboratory experiments for tsunamis generated by underwater landslides: Comparison with numerical modeling," in *Proc 13th Offshore and Polar Engng Conf (ISOPE03)*, Honolulu, USA, 2003, pp. 372–379.
- [16] F. Enet and S. Grilli, "Tsunami landslide generation: Modeling and experiments," in *Proc 5th Intl on Ocean Wave Measurement and Analysis (WAVES 2005)*. Madrid, Spain: IAHR Pub, 2005, pp. 88–98.
- [17] F. Enet and S. Grilli, "Experimental study of tsunami generation by three-dimensional rigid underwater landslides," *J Waterw Port Coast Ocean Eng*, vol. 133, no. 6, pp. 442 – 454, 2007.

- [18] F. Shi, J. T. Kirby, J. C. Harris, J. D. Geiman, and S. T. Grilli, “A high-order adaptive time-stepping TVD solver for Boussinesq modeling of breaking waves and coastal inundation,” *Ocean Modelling*, vol. 43–44, pp. 36–51, 2012.
- [19] J. Kirby, F. Shi, T. B., J. Harris, and S. Grilli, “Dispersive tsunami waves in the ocean: Model equations and sensitivity to dispersion and Coriolis effects,” *Ocean Modelling*, vol. 62, pp. 39–55, 2013.
- [20] J. Locat, H. Lee, U. ten Brink, D. Twichell, E. Geist, and M. Sansoucy, “Geomorphology, stability and mobility of the Currituck slide,” *Mar Geo*, 2009.
- [21] E. Geist, P. Lynett, and J. Chaytor, “Hydrodynamic modeling of tsunamis from the Currituck landslide,” *Marine Geology*, vol. 264, pp. 41–52, 2009.
- [22] C. Harbitz, “Model simulations of tsunamis generated by the Storegga slides,” *Mar Geol*, vol. 105, pp. 1–21, 1992.
- [23] D. J. W. Piper, P. Cochonat, and M. L. Morrison, “The sequence of events around the epicentre of the 1929 Grand Banks earthquake: initiation of debris flows and turbidity current inferred from sidescan sonar,” *Sedimentology*, vol. 46, no. 1, pp. 79–97, 1999. [Online]. Available: <http://dx.doi.org/10.1046/j.1365-3091.1999.00204.x>
- [24] I. Fine, A. Rabinovich, B. Bornhold, R. Thomson, and E. Kulikov, “The Grand Banks landslide-generated tsunami of November 18, 1929: preliminary analysis and numerical modeling,” *Mar Geol*, vol. 215, pp. 45–57, 2005.
- [25] P. Watts, S. T. Grilli, J. T. Kirby, G. J. Fryer, and D. R. Tappin, “Landslide tsunami case studies using a boussinesq model and a fully nonlinear tsunami generation model,” *Natural Hazards and Earth System Sciences*, vol. 3, no. 5, pp. 391–402, 2003. [Online]. Available: <http://www.nat-hazards-earth-syst-sci.net/3/391/2003/>
- [26] G. Fryer, P. Watts, and L. Pratson, “Source of the great tsunami of 1 April 1946: a landslide in the upper Aleutian forearc,” *Mar Geol*, vol. 203, pp. 201–218, 2004.
- [27] R. von Huene, S. Kirby, J. Miller, and P. Dartnell, “The destructive 1946 unimak near-field tsunami: New evidence for a submarine slide source from reprocessed marine geophysical data,” *Geophysical Research Letters*, vol. 41, no. 19, pp. 6811–6818, 2014, 2014GL061759. [Online]. Available: <http://dx.doi.org/10.1002/2014GL061759>
- [28] H. Fritz, W. Hager, and M. H.-E., “Lituya Bay case: rock slide impact and wave runup,” *Sc Tsunami Haz*, vol. 19, pp. 3–22, 2001.

- [29] R. Weiss, H. M. Fritz, and K. Wünnemann, “Hybrid modeling of the mega-tsunami runup in lituya bay after half a century,” *Geophysical Research Letters*, vol. 36, no. 9, pp. n/a–n/a, 2009, 109602. [Online]. Available: <http://dx.doi.org/10.1029/2009GL037814>
- [30] T. S. Murty, “Submarine slide-generated water waves in Kitimat Inlet, British Columbia,” *Journal of Geophysical Research: Oceans*, vol. 84, no. C12, pp. 7777–7779, 1979. [Online]. Available: <http://dx.doi.org/10.1029/JC084iC12p07777>
- [31] J. T. Kirby, F. Shi, D. Nicolsky, and S. Misra, “The 27 April 1975 Kitimat, British Columbia submarine landslide tsunami: A comparison of modeling approaches,” *Landslides*, 2016, published online Feb. 26, 2016.
- [32] I. Fine, A. Rabinovich, E. Kulikov, and et al., “Numerical modelling of landslide generated tsunamis with application to the skagway harbor tsunami of november 3, 1994,” in *Proc Int Conf on Tsunamis*, Paris, 1998, pp. 211–223.
- [33] C. Harbitz, G. Pedersen, and B. Gjevik, “Numerical simulations of large water waves due to landslides,” *J Hydraul Eng*, vol. 119, no. 1, pp. 1325–1342, 1993.
- [34] D. Tappin, P. Watts, G. McMurtry, Y. Lafoy, and T. Matsumoto, “The Sissano, Papua New Guinea tsunami of July 1998 - offshore evidence on the source mechanism,” *Marine Geology*, vol. 175, pp. 1 – 23, 2001. [Online]. Available: <http://www.sciencedirect.com/science/article/pii/S0025322701001311>
- [35] D. R. Tappin, P. Watts, and S. T. Grilli, “The Papua New Guinea tsunami of 1998: anatomy of a catastrophic event,” *Natural Hazards and Earth System Sciences*, vol. 8, pp. 243–266, 2008.
- [36] S. T. Grilli, S. Vogelmann, and P. Watts, “Development of a 3d Numerical Wave Tank for modeling tsunami generation by underwater landslides,” *Engineering Analysis with Boundary Elements*, vol. 26, no. 4, pp. 301–313, 2002.
- [37] P. Lynett and P.-F. Liu, “A numerical study of submarine landslide generated waves and runup,” *Proc R Soc London*, vol. A458, pp. 2885–2910, 2002.
- [38] P. Lynett and P.-F. Liu, “A numerical study of the run-up generated by three-dimensional landslides,” *J Geophys Res*, vol. 110, no. C3, pp. 2156–2202, 2005.
- [39] B. Ataie-Ashtiani and A. Najafi-Jilani, “Prediction of submerged landslide generated waves in dam reservoirs: an applied approach,” *Dam Eng*, vol. 17, no. 3, pp. 135–155, 2006.

- [40] B. Ataie-Ashtiani and A. Najafi-Jilani, "A higher-order Boussinesq-type model with moving bottom boundary: applications to submarine landslide tsunami waves," *Int J Numer Meth Fluid*, vol. 53, no. 6, pp. 1019 – 1048, 2007.
- [41] B. Ataie-Ashtiani and A. Najafi-Jilani, "Laboratory investigations on impulsive waves caused by underwater landslide," *Coastal Engineering*, vol. 55, no. 12, pp. 989–1004, 2008.
- [42] S. Abadie, D. Morichon, S. Grilli, and S. Glockner, "VOF/Navier-Stokes numerical modeling of surface waves generated by subaerial landslides," *La Houille Blanche*, vol. 1, pp. 21–26, Feb 2008.
- [43] S. Abadie, D. Morichon, S. Grilli, and S. Glockner, "Numerical simulation of waves generated by landslides using a multiple-fluid Navier-Stokes mode," *Coastal Engineering*, vol. 57, pp. 779–794, 2010.
- [44] D. Fuhrman and P. Madsen, "Tsunami generation, propagation, and run-up with a high-order Boussinesq model," *Coastal Engng*, vol. 56, pp. 747–758, 2009.
- [45] J. Horrillo, A. Wood, G. Kim, and A. Parambath, "A simplified 3-d navier-stokes numerical model for landslide-tsunami: Application to the gulf of mexico," *J Geophys Res*, vol. 118, pp. 6934–6950, 2013.
- [46] S. Yavari-Ramshe and B. Ataie-Ashtiani, "Numerical modeling of subaerial and submarine landslide generated tsunami waves - recent advances and future challenges," *Landslides*, 2016, (in press).
- [47] H. Fritz, W. Hager, and H.-E. Minor, "Nearfield characteristics of landslide generated impulse waves," *J Waterw Port Coast Ocean Eng*, vol. 130, pp. 287–302, 2004.
- [48] P.-F. Liu, T.-R. Wu, F. Raichlen, C. Synolakis, and J. Borrero, "Runup and rundown generated by three-dimensional masses," *J Fluid Mech*, vol. 536, pp. 107–144, 1995.
- [49] F. Mohammed and H. M. Fritz, "Physical modeling of tsunamis generated by three-dimensional deformable granular landslides," *Journal of Geophysical Research: Oceans*, vol. 117, no. C11, pp. n/a–n/a, 2012, c11015. [Online]. Available: <http://dx.doi.org/10.1029/2011JC007850>
- [50] S. Viroulet, D. Cébron, O. Kimmoun, and C. Kharif, "Shallow water waves generated by subaerial solid landslides," *Geophysical Journal International*, vol. 193, no. 2, pp. 747–762, 2013. [Online]. Available: <http://gji.oxfordjournals.org/content/193/2/747.abstract>

- [51] F. Maeno and F. Imamura, “Tsunami generation by a rapid entrance of pyroclastic flow into the sea during the 1883 Krakatau eruption, Indonesia,” *J. Geophys. Res.*, vol. 116, no. B9, 2011.
- [52] S. Tinti, A. Manucci, G. Pagnoni, A. Armigliato, and F. Zaniboni, “The 30 December 2002 landslide-induced tsunamis in Stromboli: sequence of the events reconstructed from the eyewitness accounts,” *Natural Hazards and Earth System Sciences*, vol. 5, no. 6, pp. 763–775, 2005. [Online]. Available: <http://www.nat-hazards-earth-syst-sci.net/5/763/2005/>
- [53] J. G. Moore, D. A. Clague, R. T. Holcomb, P. W. Lipman, W. R. Normark, and M. E. Torresan, “Prodigious submarine landslides on the Hawaiian Ridge,” *Journal of Geophysical Research: Solid Earth*, vol. 94, no. B12, pp. 17 465–17 484, 1989. [Online]. Available: <http://dx.doi.org/10.1029/JB094iB12p17465>
- [54] G. M. McMurtry, D. R. Tappin, P. N. Sedwick, I. Wilkinson, J. Fietzke, and B. Sellwood, “Elevated marine deposits in bermuda record a late quaternary megatsunami,” *Sedimentary Geology*, vol. 200, no. 3–4, pp. 155 – 165, 2007, sedimentary Features of Tsunami Deposits - Their Origin, Recognition and Discrimination: An Introduction. [Online]. Available: <http://www.sciencedirect.com/science/article/pii/S0037073806002831>
- [55] R. S. Ramalho, G. Winckler, J. Madeira, G. R. Helffrich, A. Hipólito, R. Quartau, K. Adena, and J. M. Schaefer, “Hazard potential of volcanic flank collapses raised by new megatsunami evidence,” *Science Advances*, vol. 1, no. 9, 2015. [Online]. Available: <http://advances.sciencemag.org/content/1/9/e1500456>
- [56] S. Ward and S. Day, “Ritter Island Volcano - lateral collapse and the tsunami of 1888,” *GEOPHYSICAL JOURNAL INTERNATIONAL*, vol. 154, no. 3, pp. 891–902, SEP 2003.
- [57] S. Day, P. Llanes, E. Silver, G. Hoffmann, S. Ward, and N. Driscoll, “Submarine landslide deposits of the historical lateral collapse of Ritter Island, Papua New Guinea,” *Marine and Petroleum Geology*, vol. 67, pp. 419–438, 2015.
- [58] S. Day, P. Watts, S. Grilli, and J. Kirby, “Mechanical models of the 1975 Kalapana, Hawaii earthquake and tsunami,” *Marine Geology*, vol. 215, pp. 59–92, 2005.
- [59] H. Glicken, “Rockslide-debris avalanche of May 18, 1980, Mount St. Helens Volcano, Washington,” US Geological Survey, Reston, VA, USGS Open-file Report 96-677, 1996, 90 pp.
- [60] U. ten Brink, D. Twichell, E. Geist, J. Chaytor, J. Locat, H. Lee, B. Buczkowski, R. Barkan, A. Solow, B. Andrews, T. Parsons, P. Lynett,

- J. Lin, and M. Sansoucy, "Evaluation of tsunami sources with the potential to impact the U.S. Atlantic and Gulf Coasts," USGS Administrative report to the U.S. Nuclear Regulatory Commission, Tech. Rep., 2008, 300 pps.
- [61] U. S. ten Brink, J. D. Chaytor, E. L. Geist, D. S. Brothers, and B. D. Andrews, "Assessment of tsunami hazard to the US Atlantic margin," *Marine Geology*, vol. 353, pp. 31–54, 2014.
 - [62] S. Grilli, A. Grilli, B. Tehranirad, and J. Kirby, "Modeling tsunami sources and their propagation in the Atlantic Ocean for coastal tsunami hazard assessment and inundation mapping along the US East Coast," in *Proc. 2015 COPRI Solutions to Coastal Disasters Conf.* Boston, USA: American Soc. Civil Eng., September 2015, p. 12.
 - [63] S. T. Grilli, S. Dubosq, N. Pophet, Y. Pérignon, J. Kirby, and F. Shi, "Numerical simulation and first-order hazard analysis of large co-seismic tsunamis generated in the Puerto Rico trench: near-field impact on the north shore of Puerto Rico and far-field impact on the US East Coast," *Natural Hazards and Earth System Sciences*, vol. 10, pp. 2109–2125, 2010.
 - [64] R. Barkan, U. ten Brick, and J. Lin, "Far field tsunami simulations of the 1755 Lisbon earthquake: Implication for tsunami hazard to the U.S. East Coast and the Caribbean." *Marine Geology*, vol. 264, no. 1–2, pp. 109–122, 2009.
 - [65] F. Løvholt, F.vholt, G. Pedersen, and G. Gisler, "Oceanic propagation of a potential tsunami from the la palma island," *Journal of Geophysical Research: Oceans*, vol. 113, no. C9, pp. n/a–n/a, 2008, c09026. [Online]. Available: <http://dx.doi.org/10.1029/2007JC004603>
 - [66] S. Abadie, J. Harris, S. Grilli, and R. Fabre, "Numerical modeling of tsunami waves generated by the flank collapse of the Cumbre Vieja Volcano (La Palma, Canary Islands): Tsunami source and near-field effects," *J. Geophys. Res.*, vol. 117, no. CH03050, 2012.
 - [67] B. Tehranirad, J. Harris, A. Grilli, S. Grilli, S. Abadie, J. Kirby, and F. Shi, "Far-field tsunami hazard in the North Atlantic basin from large scale flank collapses of the Cumbre Vieja Volcano, La Palma," *Pure Appl. Geophys.*, p. 28 pps, 2015.
 - [68] S. T. Grilli, O.-D. S. Taylor, C. D. Baxter, and S. Marezki, "Probabilistic approach for determining submarine landslide tsunami hazard along the upper East Coast of the United States," *Marine Geology*, vol. 264, no. 1–2, pp. 74–97, 2009.

- [69] J. Chaytor, U. ten Brink, J. Solow, and B. Andrews, "Size distribution of submarine landslides along the U.S. Atlantic Margin," *Mar Geol*, vol. 264, pp. 16–27, 2009.
- [70] U. S. ten Brink, H. J. Lee, E. L. Geist, and D. Twichell, "Assessment of tsunami hazard to the U.S. East Coast using relationships between submarine landslides and earthquakes," *Marine Geology*, vol. 264, pp. 65 – 73, 2009, tsunami hazard along the U.S. Atlantic coast. [Online]. Available: <http://www.sciencedirect.com/science/article/pii/S0025322708001710>
- [71] U. ten Brink, R. Barkan, B. Andrews, and J. Chaytor, "Size distributions and failure initiation of submarine and subaerial landslides," *Earth and Planetary Science Letters*, vol. 287, pp. 31 – 42, 2009. [Online]. Available: <http://www.sciencedirect.com/science/article/pii/S0012821X09004464>
- [72] J. S. Schnyder, J. T. Kirby, F. Shi, B. Tehranirad, G. P. Eberli, T. Mulder, and E. Ducassou, "Potential for tsunami generation by submarine slope failures along the western Great Bahama Bank," *AGU Fall Meeting Abstracts*, Dec. 2013.
- [73] D. C. Twichell, J. D. Chaytor, U. S. ten Brink, and B. Buczkowski, "Morphology of late quaternary submarine landslides along the U.S. Atlantic continental margin," *Marine Geology*, vol. 264, pp. 4 – 15, 2009, tsunami hazard along the U.S. Atlantic coast. [Online]. Available: <http://www.sciencedirect.com/science/article/pii/S0025322709000255>
- [74] M. Shelby, S. T. Grilli, and A. R. Grilli, "Tsunami hazard assessment in the hudson river estuary based on dynamic tsunami-tide simulations," *Pure and Applied Geophysics*, pp. 1–39, 2016. [Online]. Available: <http://dx.doi.org/10.1007/s00024-016-1315-y>
- [75] P. Watts and S. Grilli, "Underwater landslide shape, motion, deformation, and tsunami generation," in *Proc 13th ISOPE*, Honolulu, Hawaii, 2003, pp. 364–371.
- [76] H. T. Ma G, Kirby JT and F. Shi, "A two-layer granular landslide model for tsunami wave generation: theory and computation," *Ocean Model*, vol. 93, pp. 40–55, 2015.
- [77] S. Assier-Rzadkiewicz, C. Mariotti, and P. Heinrich, "Numerical simulation of submarine landslides and their hydraulic effects," *J Waterw Port Coast Ocean Eng*, vol. 123, pp. 149–157, 1997.
- [78] L. Jiang and P. LeBlond, "The coupling of a submarine slide and the surface waves which it generates," *J Geophys Res*, vol. 97, no. C8, pp. 12 731–12 744, 1992.

- [79] P. Heinrich, “Nonlinear water waves generated by submarine and aerial landslides,” *J Waterw Port Coast Ocean Eng*, vol. 118, pp. 249–266, 1992.
- [80] B. Ataie-Ashtiani and G. Shobeyri, “Numerical simulation of landslide impulsive waves by incompressible smoothed particle hydrodynamics,” *Intl J Num Meth Fluids*, vol. 56, no. 2, pp. 209 – 232, 2008.
- [81] S. Glimsdal, G. Pedersen, C. Harbitz, and L. F., “Dispersion of tsunamis: does it really matter ?” *Nat. Hazards Earth Syst. Sci.*, vol. 13, pp. 1507–1526, 2013.
- [82] C. Song, W. P, and H. Makse, “A phase diagram for jammed matter,” *Nature*, pp. 629–632, 2008.
- [83] H. Eilers, “Die viskosität von emulsionen hochviskoser stoffe als funktion der konzentration,” *Kolloid-Zeitschrift*, vol. 97, no. 3, pp. 313–321, 1941. [Online]. Available: <http://dx.doi.org/10.1007/BF01503023>
- [84] N. Frankel and A. Acrivos, “On the viscosity of a concentrated suspension of solid spheres,” *Chem Eng Sci*, vol. 22, pp. 847–853, 1967.
- [85] D. Quemada, “Rheology of concentrated disperse systems and minimum energy dissipation principle,” *Rheologica Acta*, vol. 16, no. 1, pp. 82–94, 1977. [Online]. Available: <http://dx.doi.org/10.1007/BF01516932>
- [86] C. I. Mendoza and I. Santamaría-Holek, “The rheology of hard sphere suspensions at arbitrary volume fractions: An improved differential viscosity model,” *The Journal of Chemical Physics*, vol. 130, no. 4, 2009. [Online]. Available: <http://scitation.aip.org/content/aip/journal/jcp/130/4/10.1063/1.3063120>
- [87] S. Mueller, E. W. Llewellyn, and H. M. Mader, “The rheology of suspensions of solid particles,” *Proceedings of the Royal Society of London A: Mathematical, Physical and Engineering Sciences*, vol. 466, no. 2116, pp. 1201–1228, 2010. [Online]. Available: <http://rspa.royalsocietypublishing.org/content/466/2116/1201>
- [88] D. B. Prior, E. H. Doyle, and T. Neurauter, “The currituck slide, mid-atlantic continental slope - revisited,” *Marine Geology*, vol. 73, no. 1, pp. 25 – 45, 1986. [Online]. Available: <http://www.sciencedirect.com/science/article/pii/002532278690109X>
- [89] F. Bouchut and S. Boyaval, “Unified derivation of thin-layer reduced models for shallow free-surface gravity flows of viscous fluids,” *European Journal of Mechanics - B/Fluids*, vol. 55, Part 1, pp. 116 – 131, 2016. [Online]. Available: <http://www.sciencedirect.com/science/article/pii/S0997754615001120>

MANUSCRIPT 3

**Algorithms for tsunami detection by High Frequency Radar :
development and case studies for tsunami impact in British Columbia,
Canada**

by

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Abstract. A shore-based High-Frequency (HF) WERA radar was recently installed by Ocean Networks Canada (ONC) near Tofino, British Columbia (Canada), to mitigate the elevated tsunami hazard along the shores of Vancouver Island, from both far- and near-field seismic sources and, in particular, from the Cascadia Subduction Zone (CSZ). With this HF radar, ocean currents can be measured up to a 70-85 km range, depending on atmospheric conditions, based on the Doppler shift they cause in ocean waves at the radar Bragg frequency. In earlier work, the authors (and others) have shown that tsunami currents need to be at least 0.15-0.20 m/s to be reliably detectable by HF radar, when considering environmental noise and background currents (from tide and mesoscale circulation). This would limit the direct detection of tsunami-induced currents to shallow water areas where they are sufficiently strong due to wave shoaling and, hence, to the continental shelf. It follows that, in locations with a narrow shelf, warning times based on such a tsunami detection method may be small.

To detect an approaching tsunami in deeper water, beyond the continental shelf, the authors have proposed a new detection algorithm that does not require “inverting” currents, but instead is based on spatial correlations of the raw radar signal at two distant locations along the same wave ray, shifted in time by the tsunami propagation time along the ray. A change in pattern of these correlations would indicate the presence of a tsunami. They validated this algorithm for idealized tsunami wave trains propagating over a simple seafloor geometry in a direction normally incident to shore. Here, this algorithm is extended and validated for realistic tsunami case studies conducted for seismic sources and using the bathymetry off of Vancouver Island, BC. Tsunami currents computed with a state-of-the-art long wave model are spatially averaged over the HF radar cells aligned along individual wave rays, obtained by solving geometric optic equations.

A model simulating ONC radars backscattered signal in space and time, as a function of the simulated tsunami currents, is applied on the Pacific Ocean side of Vancouver Island. Numerical experiments are performed, showing that the proposed algorithm works for detecting a realistic tsunami. Correlation thresholds relevant for tsunami detection can be inferred from the results.

3.1 Introduction

Major tsunamis can be enormously destructive and cause large numbers of fatalities along the world's increasingly populated and developed coastlines [1, 2]. While the brunt of tsunami impact cannot be easily attenuated, loss of life, however, can be mitigated or even eliminated by providing early warning to coastal populations. Such warnings can be issued based on early detection and assessment of the mechanisms of tsunami generation (e.g., seismicity) as well as detection of the tsunami itself as soon as possible after its generation. The latter is particularly important when the tsunami source is located close to the nearest coastal areas, and thus both energy spreading is low and propagation time is short. This is the case, for instance for co-seismic tsunamis generated in nearshore subduction zones (SZ) (e.g., Japan Trench, Puerto Rico Trench, Cascadia SZ,...), or for submarine mass failures (SMFs), that can be triggered on or near the continental shelf slope by moderate seismic activity [3, 4, 5]; meteotsunamis, also, may be generated on the continental shelves by fast moving pressure systems (e.g., derechos)[6].

The detection of offshore propagating tsunamis from a nearshore generation area is usually made in deep water, at bottom-mounted pressure sensors (so-called DART buoys), based on which a warning is issued for far-field locations. The detection of onshore propagating tsunamis in shallow water, over the continental shelf, is typically made by bottom pressure sensors and tide gauges that may not survive the impact of large tsunamis; additionally such detection is local (i.e., point-based)

and often takes place too late to be used in early warning systems. Hence, with the current detection technology used in tsunami warning systems, there may not be enough time to issue a warning for near-shore seismic or SMF tsunami sources, based on actual tsunami data. When the earthquake is the tsunami triggering mechanism, a warning can be issued based on detecting seismic waves, and from these estimating the earthquake parameters and the likelihood for tsunami generation. For non-seismically induced nearshore SMF tsunamis or for meteotsunamis, a warning can only be issued based on detecting the tsunami at nearshore sensors and, hence, there may not even be enough time to issue it before the tsunami impacts the coast; this is particularly true in the case of a narrow shelf.

The use of shore-based High Frequency (HF) radars to detect incoming tsunami waves was proposed almost 40 years ago by Barrick [7] and, more recently, was supported by numerical simulations (see, e.g., [8], [9], [10], [11]), and by HF radar measurements made during the Tohoku 2011 tsunami in Japan [12, 13, 14], in Chile [15], and in Hawaii [16]. No realtime tsunami detection algorithms were in place, but an a posteriori analysis of the radar data identified the tsunami current in the measurements. As for other nearshore currents, this works by measuring the Doppler shift tsunami currents induce on the radar signal and from this estimating time series of radial surface currents (i.e., projected on the radar line-of-sight) over a grid of radar cells covering the radar sweep area (typical cell size is one to a few km in each direction, with a range of 10s to 100s of km in the radial direction, depending on radar frequency and power). This dense spatial coverage is another advantage of HF radar detection over standard instrument methods. Tsunami detection and warning algorithms were proposed in some of these earlier studies, based on both a sufficient magnitude of the tsunami current inferred from the radar Doppler spectrum, combined with identifying its oscillatory nature in

space and time. In earlier work based on a 4.5 MHz HF radar (Stradivarius) with a 200 km range, Grilli *et al.* [17] showed that such algorithms reliably work when tsunami currents are at least $U_t \sim 0.15 - 0.20$ m/s and thus rise above background noise and currents. Hence, this limits a direct detection of tsunami currents to fairly shallow water and thus nearshore locations, and also means short warning times, unless there is a very wide shelf.

To detect a tsunami in deeper water, beyond the continental shelf, the authors have proposed a new detection algorithm that does not require “inverting” currents, but instead is based on spatial correlations of the raw radar signal at two distant locations along the same wave ray, shifted in time by the tsunami propagation time along the ray. A change in pattern of these correlations indicates the presence of a tsunami, since no other geophysical phenomenon can be responsible. They validated this algorithm only for idealized tsunami wave trains, propagating over a simple seafloor geometry in a direction normally incident to shore [17]. Here, this algorithm is extended and validated for realistic tsunami case studies conducted for seismic sources and using the bathymetry off of the Pacific Ocean side of Vancouver Island, in British Columbia (Canada). To mitigate tsunami hazard in this area from both far- and near-field seismic sources, in particular, from the Cascade Subduction Zone (CSZ), Ocean Networks Canada (ONC) has been developing a Tsunami Early Warning System (TEWS), combining various instruments deployed on the seafloor as part of their Neptune Observatory (seismometers, pressure sensors), and a shore-based WERA HF radar deployed near Tofino (TF), BC. This HF radar can remotely sense ocean currents up to a 70-85 km range (Fig. 3.1b). In this paper, we perform numerical experiments showing that the proposed algorithm also works for a site with complex bathymetry and using realistic tsunami data; based on these we discuss relevant correlation thresholds for tsunami

Grid	SW corner (Lat/Lon)	$N_x \times N_y$ meshes	Resolution (actual)	Resolution ($\sim$ m)
G0	(10.00,-180.00)	1950 x 1560	0.6 min (S)	3,600
G1	(44.00,-129.01)	700 x 600	2 min (S)	1,089
G2	(46.99, -127.52)	766 x 900	270 m (C)	270
G3	(48.25, -126.90)	1800 x 1200	90 m (C)	90

Table 3.1. Parameters of nested grids in FUNWAVE-TVD simulations, in which G0 is a spherical (S) grid with 100 km thick sponge layers on the outside boundary, and G1, G2 and G3 are spherical (S) or Cartesian (C) grids centered on the WERA radar sweep area in Tofino, BC (Figs. 3.1a and 3.2a). G1-G3 simulations are performed by one-way coupling.

detection. In these numerical experiments, tsunami currents computed with the state-of-the-art long wave model FUNWAVE-TVD [18, 19] are spatially averaged over HF radar cells aligned along individual wave rays (obtained by solving geometric optic equations). Although simulations were performed for large seismic sources in the CSZ [20], in this paper, we only detail results for a single M_w 9.1 far-field seismic source in the Semidi Subduction zone (SSZ; Fig. 3.1a). Based on time series of tsunami radial currents simulated in each radar cell, we compute the radar signal in the cells using a backscattering model [17], which is applied here for the characteristics of the radar installed in TF (carrier electromagnetic wave (EMW) frequency $f_{EM} = 13.5$ MHz).

3.2 Tsunami simulations

3.2.1 Numerical models, tsunami source and numerical grids

We extend a HF radar simulator and tsunami detection algorithm proposed earlier by Grilli *et al.* [17], and apply both to the sweep area of the WERA radar, off of Tofino, BC, based on simulated tsunami currents corresponding to the arrival of a tsunami generated by a M_w 9.1 seismic source in the SSZ (Fig. 3.1). This source was designed by the SAFFR (Science Application for Risk Reduction) group to have the same magnitude as the Tohoku 2011 event and cause maxi-

Figure 3.1. (a) Zoom-in on Pacific Ocean area covered by 2 arc-min grid G0, with initial surface elevation (color scale in meters) of the M_w 9.1 SAFRR seismic source in the Semidi Subduction Zone (SSZ), and boundary of nested model grids G1 (0.6 arc-min), G2 (270 m), and G3 (90 m) around Vancouver Island, BC) (Table 3.1). (b) Zoom-in on area of grid G3 around WERA HF radar deployment site in Tofino (TF); the thick black line marks the measurement (sweep) area (85 km radius) covered by the radar (color scale and contours are bathymetry (< 0) and topography (> 0) in meters).

mum impact in northern California [21]. Simulations of tsunami propagation are performed with FUNWAVE-TVD, a Boussinesq long wave model with extended dispersive properties, which is fully nonlinear in Cartesian grids [18] and weakly nonlinear in spherical grids [19]. The model was efficiently parallelized for use on a shared memory cluster (over 90% scalability is typically achieved), which easily allows using large grids (such as here the G0 grid, which has over 3 million meshes, as detailed below). FUNWAVE (its earlier version) and FUNWAVE-TVD have been widely used to simulate tsunami case studies [22, 23, 2, 5, 1, 4, 24, 25, 26]. Since 2010, the authors have used this model and related methodology to compute tsunami inundation maps for the US East Coast, under the auspice of the US National Tsunami Hazard Mitigation Program (NTHMP) (see work reported at: <http://chinacat.coastal.udel.edu/nthmp.html>) Both spherical and Cartesian versions of FUNWAVE-TVD were validated through benchmarking and approved for NTHMP work [27].

The initial surface elevation of the SAFFR seismic source was obtained from Kirby *et al.* [21] (source labeled “KirbyAlaskaPeninsulaTotal”; see <http://atom.giseis.alaska.edu>) and used as initial condition in FUNWAVE-TVD (Fig. 3.1a). Four levels of nested grids were used in simulations, with the coarser one G0, a 2 arc-min resolution spherical coordinate grid, covering a large area of the Pacific Ocean, and G1, G2, and G3 being spherical and Cartesian nested grids centered on Tofino, BC, with higher resolutions of 0.6 arc-min (1,089 m), 270 m, and 90 m (Figs. 3.1a and 3.2a). Table 3.1 provides parameters for each grid, including numbers of meshes. To eliminate reflection, 100 km thick sponge layers are used along the outside boundary of grid G0; simulations in finer nested grids are performed by one-way coupling. In this method, time series of surface elevation and depth-averaged current are computed for a large number of stations/numerical

Figure 3.2. Tsunami simulations with FUNWAVE-TVD off of Vancouver Island, BC: (a) Details of nested grids G1, G2 and G3 (Fig. 3.1a; Table 3.1) centered on Tofino, BC; (b) Area of grid G3 with numbered locations of numerical wave gauge stations; and (c) Zoom-in onto radar cell mesh in radar sweep area (Figs. 3.1b, 3.3), with marked locations (red dots) of G3 grid cells contained in one radar cell. [Color scale and contours show bathymetry (< 0) and topography (> 0).]

wave gages defined within a coarser grid, along the boundary of the finer grid used in the next level of nesting. Computations are fully performed in the coarser grid and then restarted in the finer grid using the station time series as boundary conditions. As these include both incident and reflected waves computed in the coarser grid, this method closely approximates open boundary conditions. It was found that a nesting ratio with a factor 3-4 reduction in mesh size allowed achieving good accuracy in tsunami simulations, which is the case for grids used here (Table 3.1).

Bathymetric/topographic data for both the 2 arc-min resolution G0 grid and the 0.6 arc-min G1 grid was interpolated from NOAA's 1 arc-min ETOPO-1 data. Bathymetry for the 270 m and 90 m resolution grids (G2 and G3) was based on the 3 arc-sec data provided for the coast of BC by NOAA's Marine Geology and Geophysics (MGG), wherever available; this higher-resolution data was also used in grids G0 and G1, instead of ETOPO-1 data, in the area overlapping with grid

G2. Another MGG 3 arc-sec dataset (the Northwest Pacific data set) was used for areas facing the US coast not covered by the BC bathymetry. Since the MGG BC dataset only included bathymetry, topography for grids G2 and G3 was based on ETOPO-1 data, which clearly is too coarse to accurately simulate coastal tsunami impact in these finer grids; this however is acceptable, since the present work focuses on detecting the tsunami offshore, at a significant distance away from the shoreline.

Except for close to shore, tsunamis are long waves that are well approximated by linear wave theory [28]; hence, tsunami currents and elevations will be on the order of,

$$\mathbf{U}_t \simeq \eta_t \sqrt{\frac{g}{h}} \frac{\mathbf{k}_t}{k_t} \quad \text{and} \quad \eta_t \simeq \eta_{t0} \left\{ \frac{c_0(h_0)}{c(h)} \right\}^{\frac{1}{2}} = \eta_{t0} \left\{ \frac{h_0}{h} \right\}^{\frac{1}{4}} \quad (3.1)$$

respectively, where $\eta_t \ll h(x, y)$ (the local depth) is surface elevation at locations (x, y) , $k_t(x, y) = |\mathbf{k}_t| = 2\pi/L_t$ is the tsunami wavenumber (with $L_t \gg 20h$ the characteristic tsunami wave length), and $\mathbf{k}_t = k_t (\cos \phi_t, \sin \phi_t)$ is the wavenumber vector, with ϕ_t the angle of the tsunami local direction of propagation with respect to the x axis ($g = 9.81 \text{ m/s}^2$, is gravitational acceleration). According to Eq. 3.1, assuming no refraction and linear long waves, the local tsunami elevation η_t can be predicted based on the initial deep water tsunami elevation η_{t0} using Green's law, where $c_0 = \sqrt{gh_0}$ is the tsunami phase speed in reference depth h_0 . It follows that $|\mathbf{U}_t| = U_t \propto h^{-3/4}$, and the tsunami current gradually increases as water depth decreases.

As detailed in [17], the proposed HF radar detection algorithm is applied along individual wave rays, which reflect site-specific refraction. These can be computed as a function of bathymetry and the ray's assumed incident direction in deep water ϕ_{t0} , independently of tsunami sources, using the geometric optics eikonal equation,

which using the long wave celerity (equal to the group velocity) reads,

$$\frac{\partial \phi_t}{\partial x} + \left\{ 1 - \frac{1}{2h} \frac{\partial h}{\partial x} \right\} \tan \phi_t \frac{\partial \phi_t}{\partial y} + \frac{1}{2h} \frac{\partial h}{\partial y} = 0 \quad (3.2)$$

and is solved for $\phi_t(x, y)$.

The pre-computed wave rays allow identifying radar cells located along specific rays (see details in next section); the tsunami propagation time between each pair of such cells (p, q) is then calculated as,

$$\Delta t_{pq} = t(\mathbf{R}_q) - t(\mathbf{R}_p) = \oint_{s(\mathbf{R}_p)}^{s(\mathbf{R}_q)} \frac{ds}{\sqrt{gh(s)}} \quad (3.3)$$

where $\mathbf{R}(x, y)$ denotes the radial position of cells in the radar grid and $s(\mathbf{R}(x, y))$ is the curvilinear abscissa along the selected wave ray, with $ds = dx \cos \phi_t + dy \sin \phi_t$.

3.2.2 Tsunami simulation results

Figure 3.3 shows maximum computed surface elevations during simulations in grid G0 (a) and G3 (b) for the SAFFR seismic source in the SSZ (Fig. 3.1a). As expected, Fig. 3.3a shows energy focusing on northern California and Oregon; however, one can also see significant tsunami elevations near Tofino. This is clearer in Fig. 3.3b, which shows that this tsunami would cause maximum elevations/runup of at least 2 m near Tofino. For the same case, Fig. 3.4a shows time series of surface elevation computed at 8 stations/numerical gauges located on the shelf off of Tofino (see locations in Fig. 3.2b). The sub-figures compare results obtained at the same locations in grids G2 and G3, showing a good agreement of the one-way coupling results for the incident tsunami; the agreement is less good once reflected waves have propagated back to the stations, as the interaction of waves with the coastline (including dissipation by breaking and bottom friction) is less accurately modeled in the coarser grid. At the two stations closest to shore near Tofino (stations 1 and 2), the incident tsunami is over 2 m height (trough to crest).

Figure 3.3. Simulations with FUNWAVE-TVD of M_w 9.1 SAFFR seismic source in SSZ (Fig. 3.1a). Maximum surface elevation (color scale in meters) during simulations in grids (Fig. 3.1): (a) G0 (only zoom-in area is shown); and (b) G3. TF marks HF radar deployment site in Tofino, BC.

Figure 3.4. Time series of (a) surface elevation at stations 1-8 (Fig. 3.2b) in grid G2 (dash lines) and G3 (solid lines); (b,c) spatially-averaged radial velocity U_r in radar cells 1-9 aligned along a given wave ray (Fig. 3.5b): (b) as a function of time; (c) shifted in time by the long wave propagation time t_{p1} from cell $p = 2, \dots, 9$ to cell 1.

Figure 3.5a shows examples of wave rays computed by solving Eq. 3.2 for the bathymetry in grid G3, off of Tofino (Fig. 3.2), assuming an incident direction from west, i.e., towards east ($\phi_{t0} = 0$); as expected, wave rays bend based on bathymetry, to become increasingly normal to bathymetric contours, in shallower water close to shore. Figure 3.4b shows time series of tsunami radial currents (i.e., projected in the radar direction $U_{tr} = \mathbf{U}_t \cdot \mathbf{R}/R$) computed at locations of 9 radar cells numbered 1-9 along a specific wave ray, in increasingly deeper water (Fig. 3.5b). As expected from Eq. 3.1, while radial velocity is over 0.4 m/s in the shallower water cell (1) in less than 50 m depth, it is less than 0.07 m/s at the deeper cell (9) close to the shelf break, in 500 m depth. The figure also shows that, independently of its magnitude, the pattern of time variation of the tsunami current repeats itself well from station 9 to 1. This is even more apparent in Fig. 3.4b where the same time series have been time- shifted by the long wave propagation time t_{p1} computed from cell $p = 2, \dots, 9$ to cell 1 using Eq. 3.3. As we shall see, such time-shifted currents are highly correlated in time, which was one of the main conclusions in the earlier work by [17], based on both idealized tsunami wave train and bathymetry; it thus appears from present results that this key property of tsunami currents for the viability of the proposed detection algorithm is confirmed to apply for realistic tsunami case studies.

3.3 Simulations of tsunami detection by HF radar

3.3.1 HF radar detection of tsunami currents

To simulate tsunami detection by HF radar, based on radial currents U_{tr} computed with FUNWAVE-TVD (e.g., Fig. 3.4), we use the HF radar simulator model developed and validated by Grilli *et al.* [17]; background information on the detection of coastal currents by HF radar can be found in this reference. Parameters of the simulator were set to match the characteristics of the WERA radar

Figure 3.5. (a) Examples of wave rays computed from an incident direction from west (green; $\phi_{t0} = 0$) and southwest (red; $\phi_{t0} = 45$) as a function of bathymetry in grid G3, over the Tofino HF radar sweep area (Fig. 3.2); (b) Radar cell grid with one selected wave ray (red line) intersecting the grid, and nine intersected cells numbered 1-9 as a function of increasing radar range and water depth.

deployed near Tofino, BC, which has a carrier electromagnetic wave (EMW) frequency $f_{EM} = 13.5$ MHz and a usable maximum range of 85 km. The radar sweep area is outlined in Fig. 3.1a and detailed in Fig. 3.5b; it is covered by radar cells within which the received radar signal is averaged, of length $\Delta R = 1.5$ km in the radial direction and angular opening $\Delta\phi_r = 1$ degree in the azimuthal direction; the detection sector of the sweep area is 120 degree, implying that cells are 1.48 km wide at a 85 km range and narrower closer to the radar (cell area: $\Delta S = R \Delta R \Delta\phi_r$ increases with range). The orientation of the radar antennas is such that one side

of the sweep area boundary is nearly parallel to the coastline southeast of Tofino (Fig. 3.5b).

Near-surface ocean currents are inferred from EMW interactions with ocean surface waves, based on the Bragg scattering property that the diffracted radar signal is maximum when it interacts with ocean waves whose wavelength is half the EWM wavelength,

$$L_B = \frac{\lambda_{EM}}{2} = \frac{gT_B^2}{2\pi} \quad \text{with} \quad \lambda_{EM} = \frac{c_{EM}}{f_{EM}}; \quad \text{thus} \quad f_B = \sqrt{\frac{g f_{EM}}{\pi C_{EM}}} \quad (3.4)$$

with $c_{EM} = 299,700 \text{ km/s}$ the speed of light in the air and f_B the Bragg frequency. For the WERA radar, we find $L_B \simeq 10.3 \text{ m}$; assuming deep ocean waves the first Eq. 3.4 further yields the wave period, $T_B = 2.57 \text{ s}$ and $f_B = 0.389 \text{ Hz}$. Based on a Pierson Moskowitz spectrum, wind waves of this period and length are present in the ocean for low wind speeds (at a 10 m elevation), $U_{10} \geq 0.318\sqrt{gL_B} = 3.2 \text{ m/s}$ (or 6.4 knot); hence, they are widespread.

Tsunami radial currents $\pm U_{tr}$ cause a Doppler effect on surface waves, which causes a shift of the Bragg frequency in the radar signal Doppler spectrum proportional to it, $\Delta f_B = \pm U_{tr}/L_B$. The magnitude of radial currents, $\bar{\tilde{U}}_{tr}(\mathbf{R}, t)$ can thus be inferred (inverted) from this shift, once the radar signal Doppler spectrum is computed; these are currents averaged (overbar) over a radar cell of area ΔS (for a monostatic configuration such as here), centered at $\mathbf{R}(x, y)$, and a measuring (or integration) time interval T_i (tilde) (here 120 s). To accurately compute the spectrum, the radar cells' spatial dimensions must be sufficiently large to include a statistically meaningful sample of ocean surface waves of various wavelengths, and particularly of length L_B . The frequency resolution of the Doppler spectrum near its peak is, $\Delta f_D = 1/T_i$ and that of the inverted current $\Delta U_{tr} = L_B/T_i$; hence, to accurately infer surface currents based on a Doppler shift, the measuring time interval must be sufficiently long, typically at least 2 min for a 13.5 MHz

(such as used here), yielding $\Delta U_{tr} = 0.086$ m/s. Because of the oscillatory nature of tsunami currents, however, T_i cannot be increased too much to improve resolution, as this would gradually reduce the cell- and time-averaged currents, until they have a nearly zero average and become undetectable. As concluded by Grilli *et al.* [17], the limited resolution of inverted currents combined with their rapidly decreasing magnitude with radar range (and increasing depth; Eq. 3.1) implies that tsunami detection algorithms, such as proposed by Lipa *et al.* [14], based on “inverting” Doppler spectral shifts would only be reliable nearshore, over the continental shelf, where tsunami currents would be sufficiently larger than background currents (e.g., $> 0.15 - 0.20$ m/s). By contrast, the new algorithm proposed by Grilli *et al.* [17], which is tested here on a realistic case study, takes advantage of the high correlation of time-shifted tsunami currents along a wave ray (Fig. 3.4c), which is also observed for the corresponding time-shifted time series of radar signals, to detect tsunami arrival in deeper water by observing a change in the radar signal correlation pattern, and hence does not require tsunami currents to reach large values to be detectable. Applying this algorithm for idealized tsunami wave trains and bathymetry, but in the presence of noise and background current, Grilli *et al.* [17] showed that the arrival of tsunami currents as low as background values of 0.05-0.1 m/s could be inferred, and thus tsunami detection can take place in deeper water, beyond the continental shelf.

3.3.2 HF radar simulator

We simulate tsunami detection by HF radar using the backscattering model (HF radar simulator) of Grilli *et al.* [17], which accounts for the presence of a time varying surface current in a random sea state. A summary of the main first-order equations of the model is given below; details and second-order equations can be found in reference.

The total surface current over the radar sweep area is assumed to be the sum of: (i) a spatially variable, but nearly stationary at the time scale of radar data acquisition ($> \mathcal{O}(T_i)$) residual (mesoscale) current, $\mathbf{U}_0(\mathbf{R})$; and (ii) a spatially and temporally varying current, $\mathbf{U}_t(\mathbf{R}, t)$ induced by the tsunami wave train (e.g., Eq. 3.1), computed here with FUNWAVE-TVD); hence, $\mathbf{U}(\mathbf{R}, t) = \mathbf{U}_0(\mathbf{R}) + \mathbf{U}_t(\mathbf{R}, t)$. The residual current, although stationary, is spatially variable in a way that depends on local and synoptic environmental oceanic conditions; in a specific case such as off of Vancouver Island, this current could be obtained from a regional ocean model, but this will not even be necessary to apply the proposed tsunami detection algorithm. Because the radar signal is simulated over cells of varying sizes (Fig. 3.5b), the tsunami-induced current computed over the finest FUNWAVE grid G3 is spatially averaged over each radar cell (e.g., Fig. 3.2c) before being used in the radar simulator.

Assuming a small steepness, the surface elevation of random ocean waves is represented by a second-order perturbation expansion, $\eta(\mathbf{R}, t) = \eta_1(\mathbf{R}, t) + \eta_2(\mathbf{R}, t)$, with,

$$\eta_1(\mathbf{R}, t) = \sum_{\epsilon=\pm 1} \int a^\epsilon(\mathbf{K}) e^{i(\mathbf{K} \cdot \mathbf{R} - \epsilon \Omega(\mathbf{K}, \mathbf{R}, t)t)} d\mathbf{K}, \quad (3.5)$$

where the integration is carried out over the wavenumber vectors, $\mathbf{K} = (K_x, K_y) = K(\cos \theta, \sin \theta)$, and wave harmonic amplitudes are given by,

$$a^\epsilon(\mathbf{K}) = \frac{1}{\sqrt{2}} \sqrt{\Psi(\epsilon \mathbf{K})} Z^\epsilon(\mathbf{K}), \quad (3.6)$$

with Ψ the directional wave energy density spectrum and $Z^\epsilon(\mathbf{K})$ a complex normal variable (with unit variance and zero mean), independent for each wave harmonic. The angular frequency of each wave component, $\Omega(\mathbf{K}, \mathbf{R}, t)$, is modulated by the surface current $\mathbf{U}(\mathbf{R}, t)$. Assuming that the tsunami current is slowly varying in time at the scale of ocean waves, i.e., the tsunami characteristic period, $T_t \gg T_p$,

the peak spectral wave period, and that waves are in the deep water regime, we have,

$$\Omega(\mathbf{K}, \mathbf{R}, t) t = (\Omega_g + \mathbf{K} \cdot \mathbf{U}_0(\mathbf{R})) t + \int_0^t \mathbf{K} \cdot \mathbf{U}_t(\mathbf{R}, \tau) d\tau, \quad (3.7)$$

where the integral is a memory term representing the cumulative effects of the tsunami current on the instantaneous wave angular frequency, and, $\Omega_g = \sqrt{gK}$ the linear angular wave frequency in deep water. Details of $\eta_2(\mathbf{R}, t)$ can be found in reference.

Here, we simulate fully developed sea states represented by a Pierson-Moskowitz (PM) directional wave energy density spectrum $\Psi(K_x, K_y)$, parametrized as a function of wind velocity at 10m elevation, U_{10} , and with a standard asymmetric angular spreading function, allowing us to model a fraction ξ of wave energy associated with waves propagating in the direction opposite to the dominant wind direction (see details in [17]). For instance, for $U_{10} = 10$ m/s, $s = 5$, and $\xi = 0.1$, we find a sea state with significant wave height, $H_s = 1.71$ m, peak spectral wavelength $L_p = 127.4$ m and, assuming deep water, peak period $T_p = 9.04$ s.

In a monostatic radar configuration, radar cells are identified by their range vector $\mathbf{R}$ center on the radar (or range R and radar steering angle ϕ_r). The Bragg vector, $\mathbf{K}_B$ is defined to point in the radar direction of observation, with $K_B = (2\pi/L_B)$. Up to second-order, the unattenuated backscattered radar signal is denoted by, $S(t) = S^1(t) + S^2(t)$, with,

$$S^1(t) = \sqrt{2} K_B^2 \sum_{\epsilon=\pm 1} \sqrt{\Psi(\epsilon \mathbf{K}_B)} e^{-i\epsilon \Omega_B t} Z^\epsilon(\mathbf{K}_B) \quad (3.8)$$

where Z^ϵ again denotes a complex normal variable (with unit variance and zero mean), the factor $\sqrt{2} K_B^2$ ensures consistency with the Doppler spectrum definition, and Ω_B is obtained from the wave dispersion relationship in the presence of a current (Eq. 3.7). The expression for the second-order signal $S^2(t)$ can be found

in reference. Accounting for effects of attenuation with range and environmental noise, the radar signal received from each cell is finally modeled as,

$$V(t) = \mathcal{A}S(t) + \mathcal{N}(t) \quad \text{with} \quad \mathcal{A}(R) = |F(R)|^2 R^{-2} \sqrt{\Delta \mathbb{S}}, \quad (3.9)$$

a geometric attenuation factor function of range R and $\mathcal{N}$ the environmental noise, detailed below. F represents the EMW attenuation by the ocean surface, which is computed here with the GRwave model, for the WERA radar frequency [29]. Environmental noise is modeled in each cell as independent complex Gaussian distributions with constant standard deviation σ_N ,

$$\mathcal{N}(t) = \sigma_N \{ \mathcal{G}_t^R(0, 1) + i \mathcal{G}_t^I(0, 1) \}, \quad (3.10)$$

where t indicates that different Gaussian random values with unit standard deviation and zero mean, $[\mathcal{G}_t^R(0, 1), \mathcal{G}_t^I(0, 1)]$, are being generated for each time level t . Since noise is not affected by range, Eq. 3.9 implies that the radar signal-to-noise ratio (SNR) gradually decreases with range, until the signal becomes undetectable from noise, which sets the effective radar measuring range (here 85 km). Here, we use the same σ_N value as in [17], which was based on HF radar experiments done in the Mediterranean sea, for normal temperature and pressure conditions; in future work, σ_N will be adjusted using site-specific values of the SNR for the WERA radar deployed offshore of Tofino, once these have been measured.

For an integration time T_i , the (non-normalized) radar Doppler spectrum is calculated at time t_s (with $\Delta t_s \leq T_i$) as the mean square of the modulus of the Fourier transform of the received radar signal $V(t)$, centered on its mean, over a finite time window $[t_s - T_i/2, t_s + T_i/2]$, that is,

$$I(f_D, t_s) = \frac{1}{T_i} \left| \int_{t_s - \frac{T_i}{2}}^{t_s + \frac{T_i}{2}} V(\tau) e^{2i\pi f_D \tau} d\tau \right|^2, \quad (3.11)$$

with f_D denoting a set of discrete Doppler frequencies (with $\omega_D = 2\pi f_D$). If the

received radar signal is simulated/(recorded) at a constant temporal sampling rate $\Delta t = T_i/N$, Eq. 3.11 can be easily computed as a summation from $-N/2$ to $N/2$.

Given a directional wave energy density spectrum $\Psi(\mathbf{K})$ and sets of random functions $Z^\epsilon(\mathbf{K})$ (representing random wave phases) and $\mathcal{G}_t$ (used to simulate noise in each cell), time series of the received signal and corresponding Doppler spectra can be simulated in each radar cell, by applying Eqs. 3.8 to 3.11, in the presence of cell-averaged radial surface currents $\overline{U}_r(\mathbf{R}, t)$. More details can be found in [17].

3.3.3 Application of algorithms for tsunami detection to SSZ tsunami

We apply the radar simulator Eqs. 3.5-3.10 to the WERA radar sweep area off of Tofino (Fig. 3.5b), assuming a local wind speed $U_{10} = 10$ m/s, no background current to start with, and using time series of cell-averaged radial tsunami currents $\overline{U}_{tr}(\mathbf{R}, t)$ calculated for the SSZ tsunami from FUNWAVE-TVD's computations (e.g., Fig. 3.4b). We simulate time series of received radar signal $V_p(t)$, in each radar cell p , and based on these time-dependent Doppler spectra $I(f_D, t_s)$ (Eq. 3.11).

Figure 3.6 shows snapshots of both tsunami surface current and corresponding Doppler spectra computed for cells aligned along the wave ray shown in Fig. 3.5b, between 1h42' and 2h16' (6,120 to 8,160 sec) into the event. These results show that a direct inversion of the propagating tsunami currents based on the oscillatory shift induced in Doppler spectra (referred to by Grilli *et al.* [17] as Tsunami Detection Algorithm 1; TDA1) would become effective at about a 45 km range. This corresponds to cell 5 in Fig. 3.5b, for which Fig. 3.4b shows maximum cell-averaged currents are just above 0.2 m/s; however, the integration time $T_i = 120$ sec used to compute Doppler spectra means the radar signal is based on currents that are also averaged over this time, $\tilde{\overline{U}}_{tr}(\mathbf{R}, t)$, which reduces current magnitude (here, just below 0.2 m/s). Assuming the tsunami is detected immediately upon reaching

Figure 3.6. Case study for SAFFR seismic source in SSZ. Doppler spectra (first-order) simulated for $T_i = 120$ s (top panels; color scale in Db) and tsunami radial current (bottom panels), in cells aligned along the same wave ray (solid black line in bottom panels and see Fig. 3.5b) in radar sweep area, at $t =$ (a) 1h42'; (b) 1h49'; (c) 1h56'; (d) 2h03'; (e) 2h10'; (f) 2h17'. Only one side (with positive frequencies) of Doppler spectra is shown, centered on the Bragg frequency $f_B = 0.375$ Hz. Tsunami currents shift this frequency by $\Delta f_B = \pm |U_{tr}|/L_B$.

Station 5, from this location of detection, the tsunami would reach the shore within 20-25 min, which only offers very low warning time. This confirms the conclusion of the idealized study of [17], that TDA1 is only reliable for $|\tilde{U}_{tr}| > 0.15 - 0.20 \text{ m/s}$.

We now evaluate the performance of the Tsunami Detection Algorithm 2 (TDA2) proposed by Grilli *et al.* [17] on the basis of time-shifted correlations of radar signal. As already pointed out, once shifted by the long wave propagation time Δt_{pq} between two cells p and q located along the same wave ray, Fig. 3.4c showed that tsunami radial currents appear to be highly correlated. This is verified here by computing the correlation of time series of time-shifted currents (subscript r was dropped for simplicity),

$$\text{corr}\{\bar{U}_{tq}(t - \Delta\tau), \bar{U}_{tp}(t)\} = \frac{1}{T_c} \int_{t-\frac{T_c}{2}}^{t+\frac{T_c}{2}} \bar{U}_{tq}(\tau - \Delta\tau) \bar{U}_{tp}(\tau) d\tau \quad (3.12)$$

as a function of an additional time lag Δt , where T_c is the correlation time (here 300 sec), which unlike T_i can be $\gg T_t$. Fig. 3.7a shows that such correlations, calculated between stations $q = 4.5$ (i.e., between stations 4 and 5; 39 km range) and $p = 1, \dots, 9$ (Figs. 3.5b), are maximum near the zero time lag and drop on either side of it (for positive or negative time lags). Because the radar signal is modulated by surface currents, we expect to observe similarly high correlations near the zero lag time for the time-shifted received radar signal,

$$\text{corr}\{V_q(t - \Delta\tau), V_p(t)\} = \left| \frac{1}{T_i} \int_t^{t+T_i} V_q(\tau - \Delta\tau) V_p^*(\tau) d\tau \right|, \quad (3.13)$$

due to the presence of a highly correlated tsunami current. By contrast these correlations should be flat as a function of time lag in the absence of a tsunami current (and this will hold true in the presence of uncorrelated background currents). This pattern is indeed observed in Figs. 3.7b and 3.7c, for correlations at stations 1 to 9, with and without current, respectively. [Note that, to reduce high frequency noise in correlations, these are computed on the analytical radar signals, which

are calculated, for simulated or measured signals, by applying a Fourier transform (FT) to the signal, removing the negative frequencies, and applying an inverse FT; see details in [17]. Correlations averaged over cells 1 to 9 are also plotted in Fig. 3.7 as thick red lines; in Figs. 3.7a and b, these are clearly peaked near the zero time lag in the presence of a tsunami current, but flat in Fig. 3.7c in the absence of a current.

Due to lack of space, we do not show results in the presence of a random background current but these will be shown at the conference, and confirm that currents from both a spatially varying (but nearly stationary at the considered scales) mesoscale current and local effects of environmental conditions (e.g., wind), have no correlation between two cells selected on a wave ray, more particularly when shifted in time by Δt_{pq} , and hence do not affect correlations in Eqs. 3.12 and 3.13. Thus, only the spatially coherent surface current caused by a tsunami affects correlations of the radar signal shifted by the long wave propagation time. This property is well supported in our numerical simulations and justifies why a much weaker, but spatially coherent, tsunami current can be detected by this algorithm, even in the presence of a background current of similar or even larger magnitude. This change in average correlation pattern over a series of cells can be dynamically evaluated in processing the radar signal in real time and used to detect the arrival of a tsunami.

It should be finally emphasized that, unlike with TDA1, with TDA2 we do not need to estimate currents by inverting the Doppler spectra and hence that limitations in such inversion in terms of integration time $T_i \ll T_t$, resolution $\Delta U_r \propto 1/T_i$, and the need for currents to be $> 0.15 - 0.20$ m/s do not apply. We see for instance that correlations between cell 4.5 and 1 to 9 in Fig. 3.7 all have a similar pattern, whereas the current magnitude is less than 0.05 m/s at station 9, in a 500 m

Figure 3.7. Time correlations, as a function of additional time lag Δt , between radar cell $q = 4.5$ (between 4 and 5; 39 km range) and $p = 1, \dots, 9$ (3 to 72 km range) aligned along the same wave ray (Fig. 3.4b), time-shifted by the long wave propagation time $t_{p4.5}$ from cell p to 4.5 of: (a) spatially-averaged radial tsunami currents U_r (Fig 3.4c); and analytical radar signals simulated with (b) and without (c) tsunami current. Red lines are cell/range averages of each correlation. Correlations use $T_C = 600s$, centered on tsunami front arrival time at cell 4.5, $t = 6,240s$.

depth and reaches nearly 0.3 m/s in station 2, in a 50 m depth. This confirms the conclusions of Grilli *et al.* [17], but here this is based on a realistic case study.

3.4 Conclusion

The detection of tsunamis by HF radars, based on a direct inversion of tsunami currents from the radar signal Doppler spectra (referred to as algorithm TDA1), is typically limited to areas where such currents are large enough as compared to background currents (i.e., $> 0.15 - 0.2$ m/s), hence, to shallow water and the continental shelf. To overcome this limitation, Grilli *et al.* [17] proposed a new detection algorithm (TDA2) based on observing changes in the pattern of correlations between two radar cells, of time series of radar signal shifted by the tsunami propagation time between the cells; TDA2 was validated on case studies with idealized tsunami wave trains and bathymetry. Here, we confirmed that TDA2 is also applicable to realistic tsunami case studies performed off of Vancouver Island, BC, for an incident tsunami from a Semidi SSZ M_w 9.1 event. Time-shifted correlations of radar signal between pairs of radar cells aligned along the same wave ray, computed for simulated tsunami currents, showed that TDA2 can detect the effects on correlations of radar signal of currents as low as 0.05 m/s; hence, this makes tsunami detection possible in deeper water, beyond the continental shelf.

TDA2 can be easily implemented in a radar system, in a real time tsunami detection mode (rather than simulation mode) for which the radar signal is continuously measured (rather than computed with a radar simulator), processed in all the radar cells (Fig. 3.5b), and time-shifted correlations are dynamically calculated between all pair of cells located along a large number of pre-computed wave rays (Fig. 3.5b). To detect tsunamis from expected (e.g., seismic) or unknown (SMF) tsunami sources, a series of wave rays can be pre-computed for tsunamis incident from a range of potential directions, based on bathymetry, and

used in the algorithm. Applying TDA2, the appearance of a peaked correlation between time series of time-shifted radar signals, in pairs of cells located along the same wave ray (for a single pair of cell or averaged over a few cells, from offshore to onshore), will indicate that a tsunami is approaching the radar. In the range of periods/time scales that are considered here, there is indeed no other geophysical phenomenon that can create long wave trains that are spatially coherent, with a current magnitude sufficient to cause measurable modulations in the HF radar signal. By computing signal correlations in all relevant pairs of cells along one or many wave rays, one can thus track the progression in time of an incoming tsunami by following the locations (front) of peaked correlations (e.g., Fig. 3.7; this was verified in simulations). In the absence of a spatially coherent current, signal correlations are independent of time lag (i.e., are flat); therefore, a marked difference in correlation pattern around the theoretical long wave propagation time (zero time lag) can be used to specify a tsunami detection threshold for the algorithm.

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List of References

- [1] M. Ioualalen, J. Asavanant, N. Kaewbanjak, S. T. Grilli, J. T. Kirby, and P. Watts, “Modeling the 26th December 2004 Indian Ocean tsunami: Case study of impact in Thailand,” *J. Geophys. Res.*, vol. 112, p. C07024, 2007.
- [2] S. Grilli, J. Harris, T. Tajalli-Bakhsh, T. Masterlark, C. Kyriakopoulos, J. Kirby, and F. Shi, “Numerical simulation of the 2011 Tohoku tsunami based on a new transient FEM co-seismic source: Comparison to far- and near-field observations,” *Pure and Applied Geophysics*, vol. 170, pp. 1333–1359, 2013.
- [3] I. Fine, A. Rabinovich, B. Bornhold, R. Thomson, and E. Kulikov, “The Grand Banks landslide-generated tsunami of November 18, 1929: preliminary analysis and numerical modeling,” *Mar Geol*, vol. 215, pp. 45–57, 2005.

- [4] D. R. Tappin, P. Watts, and S. T. Grilli, "The Papua New Guinea tsunami of 1998: anatomy of a catastrophic event," *Natural Hazards and Earth System Sciences*, vol. 8, pp. 243–266, 2008.
- [5] S. Grilli, C. O'Reilly, J. Harris, T. Tajalli-Bakhsh, B. Tehranirad, S. Banihashemi, J. Kirby, C. Baxter, T. Eggeling, G. Ma, and F. Shi, "Modeling of SMF tsunami hazard along the upper US East Coast: Detailed impact around Ocean City, MD," *Natural Hazards*, vol. 76, no. 2, pp. 705–746, 2015.
- [6] R. E. Thompson, A. B. Rabinovich, I. Fine, D. Sinnott, A. McCarthy, N. A. S. Sutherland, and L. K. Neil, "Meterological tsunamis on the ccoast of british columbia and washington," *Pyisics of Chemistry of the Earth Parts A/B/C*, vol. 34, no. 17, pp. 971–988, 2009.
- [7] D. E. Barrick, "A coastal radar system for tsunami warning," *Remote Sensing of Environment*, vol. 8, no. 4, pp. 353–358, 1979.
- [8] B. J. Lipa, D. E. Barrick, J. Bourg, and B. B. Nyden, "HF radar detection of tsunamis," *Journal of Oceanography*, vol. 62, no. 5, pp. 705–716, 2006.
- [9] M. L. Heron, A. Prytz, S. F. Heron, T. Helzel, T. Schlick, D. J. Greenslade, E. Schulz, and W. J. Skirving, "Tsunami observations by coastal ocean radar," *International Journal of Remote Sensing*, vol. 29, no. 21, pp. 6347–6359, 2008.
- [10] A. Dzvonkovskaya, K.-W. Gurgel, T. Pohlmann, T. Schlick, and J. Xu, "Simulation of tsunami signatures in ocean surface current maps measured by HF radar," in *OCEANS 2009-EUROPE*. IEEE, 2009, pp. 1–6.
- [11] K.-W. Gurgel, A. Dzvonkovskaya, T. Pohlmann, T. Schlick, and E. Gill, "Simulation and detection of tsunami signatures in ocean surface currents measured by HF radar," *Ocean Dynamics*, vol. 61, no. 10, pp. 1495–1507, 2011.
- [12] H. Hinata, S. Fujii, K. Furukawa, T. Kataoka, M. Miyata, T. Kobayashi, M. Mizutani, T. Kokai, and N. Kanatsu, "Propagating tsunami wave and subsequent resonant response signals detected by HF radar in the Kii Channel, Japan," *Estuarine, Coastal and Shelf Science*, vol. 95, no. 1, pp. 268–273, 2011.
- [13] B. Lipa, D. Barrick, S.-I. Saitoh, Y. Ishikawa, T. Awaji, J. Largier, and N. Garfield, "Japan tsunami current flows observed by HF radars on two continents," *Remote Sensing*, vol. 3, no. 8, pp. 1663–1679, 2011.
- [14] B. Lipa, J. Isaacson, B. Nyden, and D. Barrick, "Tsunami arrival detection with high frequency (HF) radar," *Remote Sensing*, vol. 4, no. 5, pp. 1448–1461, 2012.
- [15] A. Dzvonkovskaya, "Ocean surface current measurements using HF radar during the 2011 Japan tsunami hitting Chilean coast," in *Geoscience and Remote*

Sensing Symposium (IGARSS), 2012 IEEE International. IEEE, 2012, pp. 7605–7608.

- [16] L. R. Benjamin, P. Flament, K. F. Cheung, and D. S. Luther, “The 2011 Tohoku tsunami south of Oahu: high-frequency Doppler radio observations and model simulations of currents,” *J. Geophys. Res. (published online)*, pp. 1–29, 2016.
- [17] S. T. Grilli, S. Grosdidier, and C.-A. Guérin, “Tsunami detection by High Frequency Radar beyond the continental shelf. I. Algorithms and validation on idealized case studies,” *Pure and Applied Geophysics*, pp. 1–40 (published online 10/28/15), 2015.
- [18] F. Shi, J. T. Kirby, J. C. Harris, J. D. Geiman, and S. T. Grilli, “A high-order adaptive time-stepping TVD solver for Boussinesq modeling of breaking waves and coastal inundation,” *Ocean Modelling*, vol. 43–44, pp. 36–51, 2012.
- [19] J. Kirby, F. Shi, T. B., J. Harris, and S. Grilli, “Dispersive tsunami waves in the ocean: Model equations and sensitivity to dispersion and Coriolis effects,” *Ocean Modelling*, vol. 62, pp. 39–55, 2013.
- [20] T. Insua, A. R. Grilli, S. T. Grilli, M. Shelby, K. Wang, D. Gao, J. Cherniawsky, J. C. Harris, M. Heesemann, S. McLean, and K. Moran, “Preliminary tsunami hazard assessment in British Columbia, Canada,” *EOS Trans. AGU*, vol. 96(52), pp. Fall Meet. Suppl., Abstract NH23C–1890, 2015.
- [21] S. Kirby, D. Scholl, R. Von Huene, and R. Wells, “Alaska earthquake source for the SAFRR tsunami scenario,” U.S. Geological Survey Open-File . <http://pubs.usgs.gov/of/2013/1170/b/>, The SAFRR (Science Application for Risk Reduction) Tsunami Scenario, Tech. Rep. USGS Report, 2013, 2013.
- [22] S. Grilli, M. Ioualalen, J. Asavanant, F. Shi, J. Kirby, and P. Watts, “Source constraints and model simulation of the December 26, 2004 Indian Ocean tsunami,” *Journal of Waterway, Port, Coastal, and Ocean Engineering*, vol. 133, no. 6, pp. 414–428, 2007.
- [23] S. T. Grilli, S. Dubosq, N. Pophet, Y. Pérignon, J. Kirby, and F. Shi, “Numerical simulation and first-order hazard analysis of large co-seismic tsunamis generated in the Puerto Rico trench: near-field impact on the north shore of Puerto Rico and far-field impact on the US East Coast,” *Natural Hazards and Earth System Sciences*, vol. 10, pp. 2109–2125, 2010.
- [24] D. R. Tappin, S. T. Grilli, H. J. C., G. R. J., M. Timothy, K. J. T., F. Shi, G. Ma, K. Thingbaijam, and P. Maig, “Did a submarine landslide contribute to the 2011 Tohoku tsunami ?” *Marine Geology*, vol. 357, pp. 344–361, 2014.

- [25] S. Abadie, J. Harris, S. Grilli, and R. Fabre, “Numerical modeling of tsunami waves generated by the flank collapse of the Cumbre Vieja Volcano (La Palma, Canary Islands): Tsunami source and near-field effects,” *J. Geophys. Res.*, vol. 117, no. CH03050, 2012.
- [26] B. Tehranirad, J. Harris, A. Grilli, S. Grilli, S. Abadie, J. Kirby, and F. Shi, “Far-field tsunami hazard in the North Atlantic basin from large scale flank collapses of the Cumbre Vieja Volcano, La Palma,” *Pure Appl. Geophys.*, p. 28 pps, 2015.
- [27] B. Tehranirad, F. Shi, J. Kirby, J. Harris, and S. Grilli, “Tsunami benchmark results for fully nonlinear Boussinesq wave model FUNWAVE-TVD, version 1.0,” Center for Applied Coastal Research, University of Delaware, Tech. Rep., 2011, research Report No. CACR-11-02.
- [28] R. Dean and R. Dalrymple, *Water wave mechanics for engineers and scientists*, ser. Advanced Series on Ocean Engineering. Prentice-Hall, 1991, vol. 4.
- [29] S. Grosdidier, P. Forget, Y. Barbin, and C.-A. Guerin, “HF bistatic ocean Doppler spectra: Simulation versus experimentation,” *IEEE Trans. Geosci. and Remote Sens.*, vol. 52, no. 4, pp. 2138–2148, 2014.

BIBLIOGRAPHY

- “Japan quake’s economic impact worse than first feared,” reuters, published/accessed 14 April 2011. [Online]. Available: <http://www.reuters.com/article/japan-economy-idUSL3E7FC09220110412>
- “Estimates of monthly and annual net discharge, in cubic feet per second, of Hudson River at New York, N.Y. (mouth),” 2010, U.S. Department of the Interior, US Geological Survey. [Online]. Available: http://ny.water.usgs.gov/projects/dialer_plots/Hudson_R_at_NYC_Freshwater_Discharge.htm
- Abadie, S., Harris, J., Grilli, S., and Fabre, R., “Numerical modeling of tsunami waves generated by the flank collapse of the Cumbre Vieja Volcano (La Palma, Canary Islands): Tsunami source and near-field effects,” *J. Geophys. Res.*, vol. 117, no. CH03050, 2012.
- Abadie, S., Morichon, D., Grilli, S., and Glockner, S., “VOF/Navier-Stokes numerical modeling of surface waves generated by subaerial landslides,” *La Houille Blanche*, vol. 1, pp. 21–26, Feb 2008.
- Abadie, S., Morichon, D., Grilli, S., and Glockner, S., “Numerical simulation of waves generated by landslides using a multiple-fluid Navier-Stokes mode,” *Coastal Engineering*, vol. 57, pp. 779–794, 2010.
- Assier-Rzadkiewicz, S., Mariotti, C., and Heinrich, P., “Numerical simulation of submarine landslides and their hydraulic effects,” *J Waterw Port Coast Ocean Eng*, vol. 123, pp. 149–157, 1997.
- Ataie-Ashtiani, B. and Najafi-Jilani, A., “Prediction of submerged landslide generated waves in dam reservoirs: an applied approach,” *Dam Eng*, vol. 17, no. 3, pp. 135–155, 2006.
- Ataie-Ashtiani, B. and Najafi-Jilani, A., “A higher-order Boussinesq-type model with moving bottom boundary: applications to submarine landslide tsunami waves,” *Int J Numer Meth Fluid*, vol. 53, no. 6, pp. 1019 – 1048, 2007.
- Ataie-Ashtiani, B. and Najafi-Jilani, A., “Laboratory investigations on impulsive waves caused by underwater landslide,” *Coastal Engineering*, vol. 55, no. 12, pp. 989–1004, 2008.
- Ataie-Ashtiani, B. and Shobeyri, G., “Numerical simulation of landslide impulsive waves by incompressible smoothed particle hydrodynamics,” *Intl J Num Meth Fluids*, vol. 56, no. 2, pp. 209 – 232, 2008.

- Barkan, R., ten Brink, U., and Lin, J., “Far field tsunami simulations of the 1755 Lisbon earthquake: Implication for tsunami hazard to the U.S. East Coast and the Caribbean.” *Marine Geology*, vol. 264, no. 1–2, pp. 109–122, 2009.
- Barrick, D. E., “A coastal radar system for tsunami warning,” *Remote Sensing of Environment*, vol. 8, no. 4, pp. 353–358, 1979.
- Benjamin, L. R., Flament, P., Cheung, K. F., and Luther, D. S., “The 2011 Tohoku tsunami south of Oahu: high-frequency Doppler radio observations and model simulations of currents,” *J. Geophys. Res. (published online)*, pp. 1–29, 2016.
- Bouchut, F. and Boyaval, S., “Unified derivation of thin-layer reduced models for shallow free-surface gravity flows of viscous fluids,” *European Journal of Mechanics - B/Fluids*, vol. 55, Part 1, pp. 116 – 131, 2016. [Online]. Available: <http://www.sciencedirect.com/science/article/pii/S0997754615001120>
- Chaytor, J., ten Brink, U., Solow, J., and Andrews, B., “Size distribution of submarine landslides along the U.S. Atlantic Margin,” *Mar Geol*, vol. 264, pp. 16–27, 2009.
- Day, S., Llanes, P., Silver, E., Hoffmann, G., Ward, S., and Driscoll, N., “Submarine landslide deposits of the historical lateral collapse of Ritter Island, Papua New Guinea,” *Marine and Petroleum Geology*, vol. 67, pp. 419–438, 2015.
- Day, S., Watts, P., Grilli, S., and Kirby, J., “Mechanical models of the 1975 Kalapana, Hawaii earthquake and tsunami,” *Marine Geology*, vol. 215, pp. 59–92, 2005.
- Dean, R. and Dalrymple, R., *Water wave mechanics for engineers and scientists*, ser. Advanced Series on Ocean Engineering. Prentice-Hall, 1991, vol. 4.
- Dzvonkovskaya, A., “Ocean surface current measurements using HF radar during the 2011 Japan tsunami hitting Chilean coast,” in *Geoscience and Remote Sensing Symposium (IGARSS), 2012 IEEE International*. IEEE, 2012, pp. 7605–7608.
- Dzvonkovskaya, A., Gurgel, K.-W., Pohlmann, T., Schlick, T., and Xu, J., “Simulation of tsunami signatures in ocean surface current maps measured by HF radar,” in *OCEANS 2009-EUROPE*. IEEE, 2009, pp. 1–6.
- Eilers, H., “Die viskosität von emulsionen hochviskoser stoffe als funktion der konzentration,” *Kolloid-Zeitschrift*, vol. 97, no. 3, pp. 313–321, 1941. [Online]. Available: <http://dx.doi.org/10.1007/BF01503023>
- Enet, F. and Grilli, S., “Tsunami landslide generation: Modeling and experiments,” in *Proc 5th Intl on Ocean Wave Measurement and Analysis (WAVES 2005)*. Madrid, Spain: IAHR Pub, 2005, pp. 88–98.

- Enet, F. and Grilli, S., "Experimental study of tsunami generation by three-dimensional rigid underwater landslides," *J Waterw Port Coast Ocean Eng*, vol. 133, no. 6, pp. 442 – 454, 2007.
- Enet, F., Grilli, S., and Watts, P., "Laboratory experiments for tsunamis generated by underwater landslides: Comparison with numerical modeling," in *Proc 13th Offshore and Polar Engng Conf (ISOPE03)*, Honolulu, USA, 2003, pp. 372–379.
- Region II Storm Surge Project - Coastal Terrain Processing Methodology*, Federal Emergency Management Agency (FEMA), Department of Homeland Security, 500 C Street, SW Washington DC, 20472, September 2014.
- Fine, I., Rabinovich, A., Bornhold, B., Thomson, R., and Kulikov, E., "The Grand Banks landslide-generated tsunami of November 18, 1929: preliminary analysis and numerical modeling," *Mar Geol*, vol. 215, pp. 45–57, 2005.
- Fine, I., Rabinovich, A., Kulikov, E., and et al., "Numerical modelling of landslide generated tsunamis with application to the skagway harbor tsunami of november 3, 1994," in *Proc Int Conf on Tsunamis*, Paris, 1998, pp. 211–223.
- Frankel, N. and Acrivos, A., "On the viscosity of a concentrated suspension of solid spheres," *Chem Eng Sci*, vol. 22, pp. 847–853, 1967.
- Fritz, H., Hager, W., and H.-E., M., "Lituya Bay case: rock slide impact and wave runup," *Sc Tsunami Haz*, vol. 19, pp. 3–22, 2001.
- Fritz, H., Hager, W., and Minor, H.-E., "Nearfield characteristics of landslide generated impulse waves," *J Waterw Port Coast Ocean Eng*, vol. 130, pp. 287–302, 2004.
- Fryer, G., Watts, P., and Pratson, L., "Source of the great tsunami of 1 April 1946: a landslide in the upper Aleutian forearc," *Mar Geol*, vol. 203, pp. 201–218, 2004.
- Fuhrman, D. and Madsen, P., "Tsunami generation, propagation, and run-up with a high-order Boussinesq model," *Coastal Engng*, vol. 56, pp. 747–758, 2009.
- Fujii, Y., Satake, K., Sakai, S., Shinohara, M., and Kanazawa, T., "Tsunami source of the 2011 off the Pacific coast of Tohoku, Japan earthquake," *Earth, Planets, and Space*, vol. 63, pp. 815–820, 2011.
- Geist, E., Lynett, P., and Chaytor, J., "Hydrodynamic modeling of tsunamis from the Currituck landslide," *Marine Geology*, vol. 264, pp. 41–52, 2009.
- Gica, E., Spillane, M., Titov, V., Chamberlin, C., and Newman, J., "Development of the forecast propagation database for NOAA's short-term inundation forecast for tsunamis," National Oceanographic and Atmospheric Administration, NOAA Tech. Memo., 2008, OAR PMEL-139.

- Glicken, H., “Rockslide-debris avalanche of May 18, 1980, Mount St. Helens Volcano, Washington,” US Geological Survey, Reston, VA, USGS Open-file Report 96-677, 1996, 90 pp.
- Glimsdal, S., Pedersen, G., Harbitz, C., and F, L., “Dispersion of tsunamis: does it really matter ?” *Nat. Hazards Earth Syst. Sci.*, vol. 13, pp. 1507–1526, 2013.
- Grilli, A. and Grilli, S., “Modeling of tsunami generation, propagation and regional impact along the U.S. East Coast from the Azores Convergence Zone,” 2013a, research Report no. CACR-13-04, 20 pps. [Online]. Available: <http://personal.egr.uri.edu/grilli/grilli-grilli-cacr-13-04>
- Grilli, A. and Grilli, S., “Modeling of tsunami generation, propagation and regional impact along the upper U.S East Coast from the Puerto Rico trench,” 2013b. [Online]. Available: <http://personal.egr.uri.edu/grilli/grilli-grilli-cacr-13-02>
- Grilli, S. T., Grosdidier, S., and Guérin, C.-A., “Tsunami detection by High Frequency Radar beyond the continental shelf. I. Algorithms and validation on idealized case studies,” *Pure and Applied Geophysics*, pp. 1–40 (published online 10/28/15), 2015.
- Grilli, S., Grilli, A., Tehranirad, B., and Kirby, J., “Modeling tsunami sources and their propagation in the Atlantic Ocean for coastal tsunami hazard assessment and inundation mapping along the US East Coast,” in *Proc. 2015 COPRI Solutions to Coastal Disasters Conf.* Boston, USA: American Soc. Civil Eng., September 2015, p. 12.
- Grilli, S., Harris, J., Tajalli-Bakhsh, T., Masterlark, T., Kyriakopoulos, C., Kirby, J., and Shi, F., “Numerical simulation of the 2011 Tohoku tsunami based on a new transient FEM co-seismic source: Comparison to far- and near-field observations,” *Pure and Applied Geophysics*, vol. 170, pp. 1333–1359, 2013.
- Grilli, S., Ioualalen, M., Asavanant, J., Shi, F., Kirby, J., and Watts, P., “Source constraints and model simulation of the December 26, 2004 Indian Ocean tsunami,” *Journal of Waterway, Port, Coastal, and Ocean Engineering*, vol. 133, no. 6, pp. 414–428, 2007.
- Grilli, S., O’Reilly, C., Harris, J., Tajalli-Bakhsh, T., Tehranirad, B., Banihashemi, S., Kirby, J., Baxter, C., Eggeling, T., Ma, G., and Shi, F., “Modeling of SMF tsunami hazard along the upper US East Coast: Detailed impact around Ocean City, MD,” *Natural Hazards*, vol. 76, no. 2, pp. 705–746, 2015.
- Grilli, S. and Watts, P., “Modeling of waves generated by a moving submerged body. applications to underwater landslides,” *Engng Analys Boundary Elements*, vol. 23, pp. 645–656, 1999.

- Grilli, S. T., Dubosq, S., Pophet, N., Pérignon, Y., Kirby, J., and Shi, F., “Numerical simulation and first-order hazard analysis of large co-seismic tsunamis generated in the Puerto Rico trench: near-field impact on the north shore of Puerto Rico and far-field impact on the US East Coast,” *Natural Hazards and Earth System Sciences*, vol. 10, pp. 2109–2125, 2010.
- Grilli, S. T., Taylor, O.-D. S., Baxter, C. D., and Marezki, S., “Probabilistic approach for determining submarine landslide tsunami hazard along the upper East Coast of the United States,” *Marine Geology*, vol. 264, no. 1–2, pp. 74–97, 2009.
- Grilli, S. T., Vogelmann, S., and Watts, P., “Development of a 3d Numerical Wave Tank for modeling tsunami generation by underwater landslides,” *Engineering Analysis with Boundary Elements*, vol. 26, no. 4, pp. 301–313, 2002.
- Grilli, S. T. and Watts, P., “Tsunami generation by submarine mass failure Part I : Modeling, experimental validation, and sensitivity analysis,” *Journal of Waterway Port Coastal and Ocean Engineering*, vol. 131, no. 6, pp. 283–297, 2005.
- Grosdidier, S., Forget, P., Barbin, Y., and Guerin, C.-A., “HF bistatic ocean Doppler spectra: Simulation versus experimentation,” *IEEE Trans. Geosci. and Remote Sens.*, vol. 52, no. 4, pp. 2138–2148, 2014.
- Gurgel, K.-W., Dzvonkovskaya, A., Pohlmann, T., Schlick, T., and Gill, E., “Simulation and detection of tsunami signatures in ocean surface currents measured by HF radar,” *Ocean Dynamics*, vol. 61, no. 10, pp. 1495–1507, 2011.
- Harbitz, C., Pedersen, G., and Gjevik, B., “Numerical simulations of large water waves due to landslides,” *J Hydraul Eng*, vol. 119, no. 1, pp. 1325–1342, 1993.
- Harbitz, C., “Model simulations of tsunamis generated by the Storegga slides,” *Mar Geol*, vol. 105, pp. 1–21, 1992.
- Heinrich, P., “Nonlinear water waves generated by submarine and aerial landslides,” *J Waterw Port Coast Ocean Eng*, vol. 118, pp. 249–266, 1992.
- Heron, M. L., Prytz, A., Heron, S. F., Helzel, T., Schlick, T., Greenslade, D. J., Schulz, E., and Skirving, W. J., “Tsunami observations by coastal ocean radar,” *International Journal of Remote Sensing*, vol. 29, no. 21, pp. 6347–6359, 2008.
- Hinata, H., Fujii, S., Furukawa, K., Kataoka, T., Miyata, M., Kobayashi, T., Mizutani, M., Kokai, T., and Kanatsu, N., “Propagating tsunami wave and subsequent resonant response signals detected by HF radar in the Kii Channel, Japan,” *Estuarine, Coastal and Shelf Science*, vol. 95, no. 1, pp. 268–273, 2011.

- Horrrillo, J., Wood, A., Kim, G., and Parambath, A., "A simplified 3-d navier-stokes numerical model for landslide-tsunami: Application to the gulf of mexico," *J Geophys Res*, vol. 118, pp. 6934–6950, 2013.
- Insua, T., Grilli, A. R., Grilli, S. T., Shelby, M., Wang, K., Gao, D., Cherniawsky, J., Harris, J. C., Heesemann, M., McLean, S., and Moran, K., "Preliminary tsunami hazard assessment in British Columbia, Canada," *EOS Trans. AGU*, vol. 96(52), pp. Fall Meet. Suppl., Abstract NH23C–1890, 2015.
- Ioualalen, M., Asavanant, J., Kaewbanjak, N., Grilli, S. T., Kirby, J. T., and Watts, P., "Modeling the 26th December 2004 Indian Ocean tsunami: Case study of impact in Thailand," *J. Geophys. Res.*, vol. 112, p. C07024, 2007.
- Jiang, L. and LeBlond, P., "The coupling of a submarine slide and the surface waves which it generates," *J Geophys Res*, vol. 97, no. C8, pp. 12 731–12 744, 1992.
- Kirby, J. T., Shi, F., Nicolsky, D., and Misra, S., "The 27 April 1975 Kitimat, British Columbia submarine landslide tsunami: A comparison of modeling approaches," *Landslides*, 2016, published online Feb. 26, 2016.
- Kirby, J., Shi, F., B., T., Harris, J., and Grilli, S., "Dispersive tsunami waves in the ocean: Model equations and sensitivity to dispersion and Coriolis effects," *Ocean Modelling*, vol. 62, pp. 39–55, 2013.
- Kirby, S., Scholl, D., Von Huene, R., and Wells, R., "Alaska earthquake source for the SAFRR tsunami scenario," U.S. Geological Survey Open-File . <http://pubs.usgs.gov/of/2013/1170/b/>, The SAFRR (Science Application for Risk Reduction) Tsunami Scenario, Tech. Rep. USGS Report, 2013, 2013.
- Kowalik, Z. and Proshutinsky, A., "Tsunami-tide interactions: A Cook Inlet case study," *Continental Shelf Research*, vol. 30, no. 6, pp. 633–642, 2010.
- Kowalik, Z., Proshutinsky, T., and Proshutinsky, A., "Tide-tsunami interactions," *Science of Tsunami Hazards*, vol. 24, no. 4, pp. 242–256, 2006.
- Lipa, B., Barrick, D., Saitoh, S.-I., Ishikawa, Y., Awaji, T., Largier, J., and Garfield, N., "Japan tsunami current flows observed by HF radars on two continents," *Remote Sensing*, vol. 3, no. 8, pp. 1663–1679, 2011.
- Lipa, B., Isaacson, J., Nyden, B., and Barrick, D., "Tsunami arrival detection with high frequency (HF) radar," *Remote Sensing*, vol. 4, no. 5, pp. 1448–1461, 2012.
- Lipa, B. J., Barrick, D. E., Bourg, J., and Nyden, B. B., "HF radar detection of tsunamis," *Journal of Oceanography*, vol. 62, no. 5, pp. 705–716, 2006.

- Liu, P.-F., Wu, T.-R., Raichlen, F., Synolakis, C., and Borrero, J., “Runup and rundown generated by three-dimensional masses,” *J Fluid Mech*, vol. 536, pp. 107–144, 1995.
- Locat, J., Lee, H., ten Brink, U., Twichell, D., Geist, E., and Sansoucy, M., “Geomorphology, stability and mobility of the Currituck slide,” *Mar Geo*, 2009.
- Løvholt, F., vholt, F., Pedersen, G., and Gisler, G., “Oceanic propagation of a potential tsunami from the la palma island,” *Journal of Geophysical Research: Oceans*, vol. 113, no. C9, pp. n/a–n/a, 2008, c09026. [Online]. Available: <http://dx.doi.org/10.1029/2007JC004603>
- Lynett, P. and Liu, P.-F., “A numerical study of submarine landslide generated waves and runup,” *Proc R Soc London*, vol. A458, pp. 2885–2910, 2002.
- Lynett, P. and Liu, P.-F., “A numerical study of the run-up generated by three-dimensional landslides,” *J Geophys Res*, vol. 110, no. C3, pp. 2156–2202, 2005.
- Ma, G., Shi, F., and Kirby, J., “Shock-capturing non-hydrostatic model for fully dispersive surface wave processes,” *Ocean Modelling*, vol. 43–44, pp. 22–35, 2012.
- Ma G, Kirby JT, H. T. and Shi, F., “A two-layer granular landslide model for tsunami wave generation: theory and computation,” *Ocean Model*, vol. 93, pp. 40–55, 2015.
- Maeno, F. and Imamura, F., “Tsunami generation by a rapid entrance of pyroclastic flow into the sea during the 1883 Krakatau eruption, Indonesia,” *J. Geophys. Res.*, vol. 116, no. B9, 2011.
- McMurtry, G. M., Tappin, D. R., Sedwick, P. N., Wilkinson, I., Fietzke, J., and Sellwood, B., “Elevated marine deposits in bermuda record a late quaternary megatsunami,” *Sedimentary Geology*, vol. 200, no. 3–4, pp. 155 – 165, 2007, sedimentary Features of Tsunami Deposits - Their Origin, Recognition and Discrimination: An Introduction. [Online]. Available: <http://www.sciencedirect.com/science/article/pii/S0037073806002831>
- Mendoza, C. I. and Santamaría-Holek, I., “The rheology of hard sphere suspensions at arbitrary volume fractions: An improved differential viscosity model,” *The Journal of Chemical Physics*, vol. 130, no. 4, 2009. [Online]. Available: <http://scitation.aip.org/content/aip/journal/jcp/130/4/10.1063/1.3063120>
- Mohammed, F. and Fritz, H. M., “Physical modeling of tsunamis generated by three-dimensional deformable granular landslides,” *Journal of Geophysical Research: Oceans*, vol. 117, no. C11, pp. n/a–n/a, 2012, c11015. [Online]. Available: <http://dx.doi.org/10.1029/2011JC007850>

- Moore, J. G., Clague, D. A., Holcomb, R. T., Lipman, P. W., Normark, W. R., and Torresan, M. E., “Prodigious submarine landslides on the Hawaiian Ridge,” *Journal of Geophysical Research: Solid Earth*, vol. 94, no. B12, pp. 17 465–17 484, 1989. [Online]. Available: <http://dx.doi.org/10.1029/JB094iB12p17465>
- Mueller, S., Llewellyn, E. W., and Mader, H. M., “The rheology of suspensions of solid particles,” *Proceedings of the Royal Society of London A: Mathematical, Physical and Engineering Sciences*, vol. 466, no. 2116, pp. 1201–1228, 2010. [Online]. Available: <http://rspa.royalsocietypublishing.org/content/466/2116/1201>
- Murty, T. S., “Submarine slide-generated water waves in Kitimat Inlet, British Columbia,” *Journal of Geophysical Research: Oceans*, vol. 84, no. C12, pp. 7777–7779, 1979. [Online]. Available: <http://dx.doi.org/10.1029/JC084iC12p07777>
- Nakada, S., Hayashi, M., Koshimura, S., Yoneda, S., and Kobayashi, E., “Tsunami simulation generated by the greatest earthquake scenario along the Nankai Trough under consideration of tidal currents in a large bay,” in *Proceedings of the Twenty-fifth (2015) International Ocean and Polar Engineering Conference*. International Society of Offshore and Polar Engineers, 2015, pp. 811–816.
- NOAA-NGDC, “National geophysical data center, U.S. coastal relief mode,” 2013, retrieved December 2013. [Online]. Available: <http://www.ngdc.noaa.gov/mgg/coastal/crm.html>
- Okada, Y., “Surface deformation due to shear and tensile faults in a half space,” *Bull. Seismol. Soc. America*, vol. 75, no. 4, pp. 1135–1154, 1985.
- Piper, D. J. W., Cochonat, P., and Morrison, M. L., “The sequence of events around the epicentre of the 1929 Grand Banks earthquake: initiation of debris flows and turbidity current inferred from sidescan sonar,” *Sedimentology*, vol. 46, no. 1, pp. 79–97, 1999. [Online]. Available: <http://dx.doi.org/10.1046/j.1365-3091.1999.00204.x>
- Prior, D. B., Doyle, E. H., and Neurauter, T., “The currituck slide, mid-atlantic continental slope - revisited,” *Marine Geology*, vol. 73, no. 1, pp. 25 – 45, 1986. [Online]. Available: <http://www.sciencedirect.com/science/article/pii/002532278690109X>
- Quemada, D., “Rheology of concentrated disperse systems and minimum energy dissipation principle,” *Rheologica Acta*, vol. 16, no. 1, pp. 82–94, 1977. [Online]. Available: <http://dx.doi.org/10.1007/BF01516932>

- Ramalho, R. S., Winckler, G., Madeira, J., Helffrich, G. R., Hipólito, A., Quartau, R., Adena, K., and Schaefer, J. M., “Hazard potential of volcanic flank collapses raised by new megatsunami evidence,” *Science Advances*, vol. 1, no. 9, 2015. [Online]. Available: <http://advances.sciencemag.org/content/1/9/e1500456>
- Romano, F., Piatanesi, A., Lorito, S. and D’Agostino, N., Hirata, K., Atzori, S., Yamazaki, Y., and Cocco, M., “Clues from joint inversion of tsunami and geodetic data of the 2011 Tohoku-oki earthquake,” *Scientific Reports*, vol. 2, no. 385, 2012. [Online]. Available: <http://dx.doi.org/10.1038/srep00385>
- Satake, K., Fujii, Y., Harada, T., and Namegaya, Y., “Time and Space Distribution of Coseismic Slip of the 2011 Tohoku Earthquake as Inferred from Tsunami Waveform Data,” *The Bulletin of the Seismological Society of America*, vol. 103, pp. 1473–1492, May 2013.
- Schnyder, J. S., Kirby, J. T., Shi, F., Tehranirad, B., Eberli, G. P., Mulder, T., and Ducassou, E., “Potential for tsunami generation by submarine slope failures along the western Great Bahama Bank,” *AGU Fall Meeting Abstracts*, Dec. 2013.
- Shelby, M., Grilli, S. T., and Grilli, A. R., “Tsunami hazard assessment in the hudson river estuary based on dynamic tsunami–tide simulations,” *Pure and Applied Geophysics*, pp. 1–39, 2016. [Online]. Available: <http://dx.doi.org/10.1007/s00024-016-1315-y>
- Shi, F., Kirby, J. T., Harris, J. C., Geiman, J. D., and Grilli, S. T., “A high-order adaptive time-stepping TVD solver for Boussinesq modeling of breaking waves and coastal inundation,” *Ocean Modelling*, vol. 43–44, pp. 36–51, 2012.
- Shi, F., Kirby, J., and Tehranirad, B., “Tsunami benchmark results for spherical coordinate version of FUNWAVE-TVD (version 1.1),” Center for Applied Coastal Research, Univ. of Delaware, Tech. Rep., 2012b, research Report no. CACR-12-02.
- Song, C., P, W., and Makse, H., “A phase diagram for jammed matter,” *Nature*, pp. 629–632, 2008.
- Stammer, D., Ray, R. D., Andersen, O. B., Arbic, B. K., Bosch, W., Carrere, L., Cheng, Y., Chinn, D. S., Dushaw, B. D., Egbert, G. D., Erofeeva, S. Y., Fok, H. S., Green, J. A. M., Griffiths, S., King, M. A., Lapin, V., Lemoine, F. G., Luthcke, S. B., Lyard, F., Morison, J., Muller, M., Padman, L., Richman, J. G., Shriver, J. F., Shum, C. K., Taguchi, E., and Yi, T., “Accuracy assessment of global barotropic ocean tide models,” *Reviews of Geophysics*, vol. 52, no. 3, pp. 243–282, 2014.

- Tajali-Bakhsh, T. S., Grilli, S. T., and Grilli, A. R., "Dynamic tide effects on tsunami coastal hazard in large estuaries: Case of the Chesapeake Bay/James River, USA," Center for Applied Coastal Research, Dept. of Civil and Environmental Engineering, University of Delaware, Tech. Rep., 2014, research Report no. CACR-15-09, 42 pps.
- Tappin, D. R., Grilli, S. T., C., H. J., J., G. R., Thimothy, M., T., K. J., Shi, F., Ma, G., Thingbaijam, K., and Maig, P., "Did a submarine landslide contribute to the 2011 Tohoku tsunami ?" *Marine Geology*, vol. 357, pp. 344–361, 2014.
- Tappin, D. R., Watts, P., and Grilli, S. T., "The Papua New Guinea tsunami of 1998: anatomy of a catastrophic event," *Natural Hazards and Earth System Sciences*, vol. 8, pp. 243–266, 2008.
- Tappin, D., Watts, P., McMurtry, G., Lafoy, Y., and Matsumoto, T., "The Sissano, Papua New Guinea tsunami of July 1998 - offshore evidence on the source mechanism," *Marine Geology*, vol. 175, pp. 1 – 23, 2001. [Online]. Available: <http://www.sciencedirect.com/science/article/pii/S0025322701001311>
- Tehrani-rad, B., Banihashemi, S., Kirby, J. T., Callahan, J. A., and Shi, F., "Tsunami inundation mapping for Ocean City, MD NGDC DEM," Center for Applied Coastal Research, Department of Civil and Environmental Engineering, University of Delaware, Tech. Rep., 2014, research Report No. CACR-14-04.
- Tehrani-rad, B., Harris, J., Grilli, A., Grilli, S., Abadie, S., Kirby, J., and Shi, F., "Far-field tsunami hazard in the North Atlantic basin from large scale flank collapses of the Cumbre Vieja Volcano, La Palma," *Pure Appl. Geophys.*, p. 28 pps, 2015.
- Tehrani-rad, B., Shi, F., Kirby, J., Harris, J., and Grilli, S., "Tsunami benchmark results for fully nonlinear Boussinesq wave model FUNWAVE-TVD, version 1.0," Center for Applied Coastal Research, University of Delaware, Tech. Rep., 2011, research Report No. CACR-11-02.
- ten Brink, U., Twichell, D., Geist, E., Chaytor, J., Locat, J., Lee, H., Buczkowski, B., Barkan, R., Solow, A., Andrews, B., Parsons, T., Lynett, P., Lin, J., and Sansoucy, M., "Evaluation of tsunami sources with the potential to impact the U.S. Atlantic and Gulf Coasts," USGS Administrative report to the U.S. Nuclear Regulatory Commission, Tech. Rep., 2008, 300 pps.
- ten Brink, U. S., Chaytor, J. D., Geist, E. L., Brothers, D. S., and Andrews, B. D., "Assessment of tsunami hazard to the US Atlantic margin," *Marine Geology*, vol. 353, pp. 31–54, 2014.

- ten Brink, U. S., Lee, H. J., Geist, E. L., and Twichell, D., “Assessment of tsunami hazard to the U.S. East Coast using relationships between submarine landslides and earthquakes,” *Marine Geology*, vol. 264, pp. 65 – 73, 2009, tsunami hazard along the U.S. Atlantic coast. [Online]. Available: <http://www.sciencedirect.com/science/article/pii/S0025322708001710>
- ten Brink, U., Barkan, R., Andrews, B., and Chaytor, J., “Size distributions and failure initiation of submarine and subaerial landslides,” *Earth and Planetary Science Letters*, vol. 287, pp. 31 – 42, 2009. [Online]. Available: <http://www.sciencedirect.com/science/article/pii/S0012821X09004464>
- Thompson, R. E., Rabinovich, A. B., Fine, I., Sinnott, D., McCarthy, A., Sutherland, N. A. S., and Neil, L. K., “Meterological tsunamis on the ccoast of british columbia and washington,” *Pyisics of Chemistry of the Earth Parts A/B/C*, vol. 34, no. 17, pp. 971–988, 2009.
- Tinti, S., Manucci, A., Pagnoni, G., Armigliato, A., and Zaniboni, F., “The 30 December 2002 landslide-induced tsunamis in Stromboli: sequence of the events reconstructed from the eyewitness accounts,” *Natural Hazards and Earth System Sciences*, vol. 5, no. 6, pp. 763–775, 2005. [Online]. Available: <http://www.nat-hazards-earth-syst-sci.net/5/763/2005/>
- Tolkova, E., “Tide-tsunami interaction in the Columbia River, as implied by historical data and numerical simulations,” *Pure and Applied Geophysics*, vol. 170, no. 6–8, pp. 1115–1126, 2013.
- Tolkova, E., Tanaka, H., and Roh, M., “Tsunami observations in rivers from a perspective of tsunami interaction with tide and riverine flow,” *Pure and Applied Geophysics*, vol. 172, no. 3–4, pp. 953–968, 2015.
- Twichell, D. C., Chaytor, J. D., ten Brink, U. S., and Buczkowski, B., “Morphology of late quaternary submarine landslides along the U.S. Atlantic continental margin,” *Marine Geology*, vol. 264, pp. 4 – 15, 2009, tsunami hazard along the U.S. Atlantic coast. [Online]. Available: <http://www.sciencedirect.com/science/article/pii/S0025322709000255>
- Viroulet, S., Cébron, D., Kimmoun, O., and Kharif, C., “Shallow water waves generated by subaerial solid landslides,” *Geophysical Journal International*, vol. 193, no. 2, pp. 747–762, 2013. [Online]. Available: <http://gji.oxfordjournals.org/content/193/2/747.abstract>
- von Huene, R., Kirby, S., Miller, J., and Dartnell, P., “The destructive 1946 unimak near-field tsunami: New evidence for a submarine slide source from reprocessed marine geophysical data,” *Geophysical Research Letters*, vol. 41, no. 19, pp. 6811–6818, 2014, 2014GL061759. [Online]. Available: <http://dx.doi.org/10.1002/2014GL061759>

- Ward, S. and Day, S., “Cumbre Vieja Volcano potential collapse at La Palma, Canary Islands,” *Geophysical Research Letter*, vol. 28, no. 17, pp. 3397–3400, 2001.
- Ward, S. and Day, S., “Ritter Island Volcano - lateral collapse and the tsunami of 1888,” *GEOPHYSICAL JOURNAL INTERNATIONAL*, vol. 154, no. 3, pp. 891–902, SEP 2003.
- Watts, P., Grilli, S. T., Kirby, J. T., Fryer, G. J., and Tappin, D. R., “Landslide tsunami case studies using a boussinesq model and a fully nonlinear tsunami generation model,” *Natural Hazards and Earth System Sciences*, vol. 3, no. 5, pp. 391–402, 2003. [Online]. Available: <http://www.nat-hazards-earth-syst-sci.net/3/391/2003/>
- Watts, P. and Grilli, S., “Underwater landslide shape, motion, deformation, and tsunami generation,” in *Proc 13th ISOPE*, Honolulu, Hawaii, 2003, pp. 364–371.
- Watts, P., Grilli, S. T., Tappin, D. R., and Fryer, G. J., “Tsunami generation by submarine mass failure. ii: predictive equations and case studies,” *Journal of Waterway, Port, Coastal, and Ocean Engineering*, vol. 131, no. 6, pp. 298–310, 2005. [Online]. Available: <http://nora.nerc.ac.uk/504819/>
- Wei, G., Kirby, J., Grilli, S., and Subramanya, R., “A fully nonlinear Boussinesq model for free surface waves. part i: Highly nonlinear unsteady waves,” *Journal of Fluid Mechanics*, vol. 294, pp. 71–92, 1995.
- Wei Y., Newman A., H. G. T. V. and Tang, L., “Tsunami forecast by joint inversion of real-time tsunami waveforms and seismic or gps data: application to the tohoku 2011 tsunami,” *Pure Appl Geophys*, vol. 171, no. 12, pp. 3281–3305, 2014.
- Weiss, R., Fritz, H. M., and Wünnemann, K., “Hybrid modeling of the mega-tsunami runup in lituya bay after half a century,” *Geophysical Research Letters*, vol. 36, no. 9, pp. n/a–n/a, 2009, 109602. [Online]. Available: <http://dx.doi.org/10.1029/2009GL037814>
- Yavari-Ramshe, S. and Ataie-Ashtiani, B., “Numerical modeling of subaerial and submarine landslide generated tsunami waves - recent advances and future challenges,” *Landslides*, 2016, (in press).
- Yeh, H., Tolkova, E., Jay, D., Talke, S., and Fritz, H., “Tsunami hydrodynamics in the Columbia River,” *J. Disaster Res.*, vol. 7, no. 5, pp. 604–608, 2012.
- Zhang, Y., Witter, R., and Priest, G., “Tsunami-tide interaction in 1964 Prince Williams Sound tsunami,” *Ocean Modelling*, vol. 40, no. 3–4, pp. 246–259, 2011.