University of Rhode Island

DigitalCommons@URI

Open Access Dissertations

2020

INVESTIGATION OF LANDSLIDE-INDUCED TSUNAMI GENERATION AND PROPAGATION THROUGH NUMERICAL MODELING

Lauren Schambach

University of Rhode Island, lschambach@my.uri.edu

Follow this and additional works at: https://digitalcommons.uri.edu/oa_diss

Recommended Citation

Schambach, Lauren, "INVESTIGATION OF LANDSLIDE-INDUCED TSUNAMI GENERATION AND PROPAGATION THROUGH NUMERICAL MODELING" (2020). *Open Access Dissertations*. Paper 1197.
https://digitalcommons.uri.edu/oa_diss/1197

This Dissertation is brought to you for free and open access by DigitalCommons@URI. It has been accepted for inclusion in Open Access Dissertations by an authorized administrator of DigitalCommons@URI. For more information, please contact digitalcommons-group@uri.edu.

INVESTIGATION OF LANDSLIDE-INDUCED TSUNAMI GENERATION
AND PROPAGATION THROUGH NUMERICAL MODELING

BY

LAUREN SCHAMBACH

A DISSERTATION SUBMITTED IN PARTIAL FULFILLMENT OF THE
REQUIREMENTS FOR THE DEGREE OF
DOCTOR OF PHILOSOPHY
IN
OCEAN ENGINEERING

UNIVERSITY OF RHODE ISLAND

2020

DOCTOR OF PHILOSOPHY DISSERTATION
OF
LAUREN SCHAMBACH

APPROVED:

Dissertation Committee:

Major Professor Stéphan T. Grilli

Jason M. Dahl

Tetsu Hara

Annette R. Grilli

Brenton DeBoef

DEAN OF THE GRADUATE SCHOOL

UNIVERSITY OF RHODE ISLAND

2020

ABSTRACT

This body of work consists of three manuscripts regarding numerical modeling of landslide-induced tsunamis.

Landslide Tsunami Hazard Along the Upper US East Coast: Effects of Slide Deformation, Bottom Friction, and Frequency Dispersion

Numerical simulations of Submarine Mass Failures (SMFs) are performed along the upper US East Coast to assess the effect of slide deformation on predicted tsunami hazard. Tsunami generation is simulated using the three-dimensional non-hydrostatic model NHWAVE. For rigid slumps, the geometry and law of motion are specified as bottom boundary conditions. Deforming slide motion is modeled using a depth-integrated bottom layer of dense Newtonian fluid, fully coupled to the overlying fluid motion. Once the SMFs are no-longer tsunamigenic, tsunami propagation simulations are performed using the Boussinesq wave model FUNWAVE-TVD, using nested grids of increasingly fine resolution towards shore and employing a one-way coupling methodology. Probable maximum tsunamis are simulated for Currituck SMF proxies sited in four areas of the shelf break slope that have enough sediment accumulation to cause large failures. Deforming slides have a slightly larger initial acceleration, but still generate a smaller tsunami than rigid slumps due to their spreading and thinning out during motion, which gradually makes them less tsunamigenic. Comparing the maximum envelope of surface elevations along a 5 m isobath, consistent with earlier work, the bathymetry of the wide shelf is found to strongly control the spatial distribution of tsunami inundation. Overall, tsunamis caused by rigid slumps are worst case scenarios, providing up to 50% more inundation than for deforming slides having a moderate level of viscosity set in the upper range of debris flows. Tsunamis from both types of SMFs are able to cause water withdrawal to the 5 m isobath or deeper. Bottom friction

effects are assessed by performing some of the simulations using two different Manning coefficients, one 50% larger than the other. With increased bottom friction, the largest tsunami inundations at the coast are reduced by up to 15%. Selected simulations are rerun by turning off dispersion in the model, which leads to moderate changes in maximum surface elevations nearshore (- 10 to + 5% changes), but to more significant effects in the far field (- 40 to 80% changes). Onshore, dispersion causes the appearance of short period undular bores that eventually break nearshore without significantly affecting inundation at the coast. However, these bores increase wave-induced maximum flow velocity and impulse forces, the latter by up to 40%, which may affect the design of coastal structures.

New simulations and understanding of the 1908 Messina tsunami for a dual seismic and deep submarine mass failure source

Over 100 years after the event, the mechanism of the 1908 Messina tsunami remains unresolved. The up to 12 m runups observed along the coasts of Sicily and Calabria cannot be explained by the coseismic tsunami, so recent studies have proposed a dual earthquake/submarine mass failure (SMF) mechanism. Here we propose a new dual source and use it to simulate tsunami generation with a three-dimensional non-hydrostatic model, coupled to a two-dimensional fully nonlinear and dispersive model, to simulate tsunami propagation to shore. We first reanalyze observations of tsunami arrival times from eyewitnesses acquired shortly after the 1908 event, and a tsunami record at a tide gauge in Malta. Similar to earlier work, this data is used to locate the likeliest tsunami source area by inverse wave ray tracing, but accounting for frequency dispersion effects on wave celerity, uncertainty in reported arrival times, and a time delay between the EQ and SMF triggering. Analyzing the seafloor morphology in this area, we identify a new SMF at the foot of the Fiumefreddo Valley, northeast of Mount Etna. The

general location is consistent with earlier studies, however our SMF is much smaller ($\sim 2 \text{ km}^3$) than, e.g., that of Billi et al. (2008) and is a fairly rigid-block-slump, rather than a translational SMF. We model the block motion and simulate tsunami generation from a dual EQ/SMF source, and its propagation to shore, in higher resolution grids and based on more accurate bathymetry and topography than in earlier work. Runups and travel times agree well with observations, except for runups on either side of the Messina Straits north of the SMF, which are still underpredicted. In the far field, simulations reproduce well the arrival time and initial wave amplitudes at the Malta tide gauge. Our newly parameterized SMF and modeling improve tsunami runups simulated near the SMF location and south of it. However, as with all previous modeling of this event, additional sources are required to explain runups in the northern Messina Straits, which we suggest might be smaller and shallower SMFs located in this area. These will be considered in future work.

Dual earthquake/landslide source modeling of the 2018 Palu tsunami generation and hazard

The Mw 7.5 earthquake that struck Central Sulawesi, Indonesia, on September 28, 2018, was rapidly followed by coastal landslides and destructive tsunami waves within Palu Bay. This earthquake was supershear and predominantly strike-slip, with most published mechanisms predicting limited seabed uplift/subsidence, which make it an unlikely source of the up to 10.5 m runups recorded in the southern portion of the bay. Scientific debate has continued over the tsunami mechanism; earthquake, coastal landslides, or a combination of both. Published research has been inconclusive, with some studies simulating an earthquake generated tsunami as explaining most observations, with others focusing solely on landslide sources. For the latter, most simulations are based on hypothetical landslides not identified

in post-tsunami onland field and bathymetric surveys. In this work, we simulate the tsunamis generated by the earthquake models of Jamelot et al. (2019), Socquet et al. (2019), and Ulrich et al. (2019), alone and in combination with seven coastal landslides that were confirmed by the field and bathymetric surveys (Liu et al., 2020; Takagi et al., 2019) which, from video evidence, produced significant waves. To generate and propagate the tsunamis, we use a combination of two numerical models, the 3D non-hydrostatic wave model NHWAVE and the 2D Boussinesq wave model FUNWAVE-TVD. Both models address the physics of wave frequency dispersion, which is important for modeling landslide tsunamis identified in the event. The coastal landslides are modeled in NHWAVE as granular material. Our combined earthquake and coastal landslide cases recreate the observed tsunami runups except for those in the southeast of the bay where they were most elevated (10.5 m). With regard to the timing of tsunami impact on the coast, results for the dual landslide/earthquake sources are in reasonable agreement with reconstructed time series at several locations around the bay, particularly using the model of Ulrich et al. (2019). In agreement with other studies, our work suggests an additional tsunami mechanism is necessary in the southeast of Palu Bay to explain observations there. Using partial information from bathymetric surveys, we site an additional landslide in the SE of Palu Bay and show that, when simulated together with the other slides and Ulrich et al. (2019)'s earthquake, results can better explain observations in the Southeast. This supports the need for future marine geology work in this area.

ACKNOWLEDGEMENTS

Getting a PhD in something totally rad has been a dream of mine for a really long time. I find myself immensely thankful that I've been able to work on such cool, interesting, and important projects over the last few years. I've been surrounded by such great and inspiring people on this journey, and I would like to take this opportunity to express my gratitude. I know acknowledgements are traditionally supposed to be kept to a page, but this is my dissertation and there are some things that need to be said on the record.

First, I would like to thank my advisor, Dr. Stéphan Grilli, for seeing potential in me and trusting me with this opportunity. He has been someone who I have looked up to since the first wave mechanics class I took with him back in the spring of 2011. When I was finishing my masters degree, we met in his office and he told me about the tsunami projects he was working on and asked if I was interested in doing a PhD with him. I agreed, he asked if he could count on me, I said yes, and then we high-fived. On my worst and most frustrated days I always thought back to that moment to help me carry on, because backing out of a high-five agreement is never an option.

I would also like to thank Professor Dave Tappin, who I met at 5 in the morning in the PVD airport on my first official day as a PhD student. Dave has patiently explained marine geology features to me over the last few years and never once complained about the amount of screenshots I've sent him of the seafloor with hand drawn circles around things I don't understand. He has been a pleasure to work with.

I must also thank Drs. Fengyan Shi, Cheng Zhang, and Jim Kirby at the University of Delaware. They have been so helpful in answering any questions about the numerical models I've had, and helping to troubleshoot any issues that

have come up.

Sometimes being a woman in engineering has felt particularly hard. I've been to classes, conferences, and meetings where I've looked around the room and felt like a total alien. Dr. Annette Grilli has been THE most influential mentor in my life. She is brilliant, an incredible problem solver, and also so passionate and so caring. Talking with her always reminds me that it's okay for me to be me, and I am so so grateful to her for that.

The entire OE department has had to deal with me for the past 12 years, and I'd like to thank all of the professors for teaching me some really cool stuff. I must also thank Gail Paolino, without whom I'm not sure any of us would be able to do anything within a reasonable amount of time. And Fred Pease has been giving me encouraging waves and smiles since I was an undergrad, which always brighten my day.

The sheets 223 crew, the hurricane modeling crew, and especially Dr. Boma and Dr. Perkins, I couldn't have gotten to this point with my sanity still (mostly) intact without you all. Dr. Adam Keen, thanks for the chats, especially during the COVID19 lockdown, and making me feel less alone.

Kerry, Amanda, John, Maggie, the entire DangerZone/ Activities crew, Newport Run and Chug and November Project, Drive Thru/Ev City, Tower Hill Hostel, and the Oasis yacht/patio crew - thank you for the endless encouragement and support.

To my family, the SCC, Kal, and especially my parents, Daryl and Lynn, and my siblings, Jenna and Brad, you are my rocks. Oceans of love, and "thank's guys, you're the best." To Gran and Grandad, I will "keep up the good work".

And finally I would like to acknowledge myself, for pushing myself to keep learning and growing and refusing to settle or give up.

DEDICATION

This dissertation is dedicated to the memory of Mary Y. Schambach, with oceans of love.

PREFACE

This work consists of three manuscripts with the overarching theme of numerically modeling landslide-induced tsunamis. The formatting of the manuscripts within this dissertation differ from those in or submitted for publication to fit the University of Rhode Island formatting guidelines, however, the content is identical.

The first manuscript is entitled, “Landslide Tsunami Hazard Along the Upper US East Coast: Effects of Slide Deformation, Bottom Friction, and Frequency Dispersion” and was published in *Pure and Applied Geophysics* (available online September 3, 2018; <https://doi.org/10.1007/s00024-018-1978-7>).

The second manuscript is entitled, “New simulations and understanding of the 1908 Messina tsunami for a dual seismic and deep submarine mass failure source” and was published in *Marine Geology* (available online December 15, 2019; <https://doi.org/10.1016/j.margeo.2019.106093>).

The third manuscript is entitled, “Dual earthquake/landslide source modeling of the 2018 Palu tsunami generation and hazard” and is in preparation for submission to *Frontiers*.

TABLE OF CONTENTS

ABSTRACT	ii
ACKNOWLEDGEMENTS	vi
DEDICATION	viii
PREFACE	ix
TABLE OF CONTENTS	x
MANUSCRIPT	
1 Landslide Tsunami Hazard Along the Upper US East Coast: Effects of Slide Deformation, Bottom Friction, and Fre- quency Dispersion	1
1.1 Introduction	3
1.2 Modeling Methodology	13
1.2.1 Numerical Models	13
1.2.2 Methodology and Computational Grids	15
1.3 SMF Tsunami Generation with NHWAVE	19
1.3.1 SMF Geometry, Rheology, and Kinematics	19
1.3.2 Tsunami Generation	27
1.4 Onshore Tsunami Propagation and Coastal Impact	28
1.4.1 Instantaneous Propagation and Time Series	30
1.4.2 Maximum and Minimum Surface Elevations, and Bottom Friction Effects	34
1.5 Importance of Dispersive Effects on the SMF Tsunami Hazard	40
1.5.1 SMF Tsunami Generation	41

	Page
1.5.2 Far-Field Propagation in the Atlantic	44
1.5.3 Near-Field Propagation	48
1.6 Discussion	58
1.7 Conclusion	64
2 New Simulations and Understanding of the 1908 Messina Tsunami for a Dual Seismic and Deep Submarine Mass Fail- ure Source	80
2.1 Introduction	82
2.2 Rationale	89
2.3 Methods	92
2.3.1 Study area, computational grids, and bathymet- ric/topographic data	92
2.3.2 Observations	94
2.3.3 Inverse wave ray tracing	98
2.3.4 Siting of submarine mass failure	102
2.3.5 Tsunami generation and propagation simulations	104
2.4 Tsunami simulation results	114
2.4.1 Coseismic sources	114
2.4.2 Dual coseismic-SMF sources	114
2.5 Discussion	122
2.6 Conclusions	126
3 Dual earthquake/landslide source modeling of the 2018 Palu tsunami generation and hazard	144
3.1 Introduction	146
3.2 Materials and Methods	153

	Page
3.2.1 Tsunami Observations	153
3.2.2 Materials and methods	163
3.3 Results	176
3.4 Discussion	182

MANUSCRIPT 1

**Landslide Tsunami Hazard Along the Upper US East Coast: Effects of
Slide Deformation, Bottom Friction, and Frequency Dispersion**

by

Lauren Schambach, Stéphan T. Grilli, James T. Kirby, and Fengyan Shi

Schambach, L., Grilli, S.T., Kirby, J.T., and F. Shi (2019). Landslide Tsunami Hazard Along the Upper US East Coast: Effects of Slide Deformation, Bottom Friction, and Frequency Dispersion. Pure and Applied Geophysics, vol. 176, no. 7, pp. 3059-3098; <https://doi.org/10.1007/s00024-018-1978-7>

Abstract. Numerical simulations of Submarine Mass Failures (SMFs) are performed along the upper US East Coast to assess the effect of slide deformation on predicted tsunami hazard. Tsunami generation is simulated using the three-dimensional non-hydrostatic model NHWAVE. For rigid slumps, the geometry and law of motion are specified as bottom boundary conditions. Deforming slide motion is modeled using a depth-integrated bottom layer of dense Newtonian fluid, fully coupled to the overlying fluid motion. Once the SMFs are no-longer tsunamigenic, tsunami propagation simulations are performed using the Boussinesq wave model FUNWAVE-TVD, using nested grids of increasingly fine resolution towards shore and employing a one-way coupling methodology. Probable maximum tsunamis are simulated for Currituck SMF proxies sited in four areas of the shelf break slope that have enough sediment accumulation to cause large failures. Deforming slides have a slightly larger initial acceleration, but still generate a smaller tsunami than rigid slumps due to their spreading and thinning out during motion, which gradually makes them less tsunamigenic. Comparing the maximum envelope of surface elevations along a 5 m isobath, consistent with earlier work, the bathymetry of the wide shelf is found to strongly control the spatial distribution of tsunami inundation. Overall, tsunamis caused by rigid slumps are worst case scenarios, providing up to 50% more inundation than for deforming slides having a moderate level of viscosity set in the upper range of debris flows. Tsunamis from both types of SMFs are able to cause water withdrawal to the 5 m isobath or deeper. Bottom friction effects are assessed by performing some of the simulations using two different Manning coefficients, one 50% larger than the other. With increased bottom friction, the largest tsunami inundations at the coast are reduced by up to 15%. Selected simulations are rerun by turning off dispersion in the model, which leads to moderate changes in maximum surface elevations nearshore (- 10 to + 5% changes),

but to more significant effects in the far field (- 40 to 80% changes). Onshore, dispersion causes the appearance of short period undular bores that eventually break nearshore without significantly affecting inundation at the coast. However, these bores increase wave-induced maximum flow velocity and impulse forces, the latter by up to 40%, which may affect the design of coastal structures.

1.1 Introduction

This study is part of tsunami hazard assessment work carried out by the authors since 2010 to develop tsunami inundation maps for the US East Coast (USEC), under the auspices of the US National Tsunami Hazard Mitigation Program (NTHMP). While Probabilistic Tsunami Hazard Analyses (PTHAs) have been considered or performed in other areas of the US and the world (e.g., González et al. 2009; Heidarzadeh and Kijko 2011; Horspool et al. 2014; Geist and Lynett 2014; Grezio et al. 2017), the paucity of paleo-tsunami information makes it difficult at present to conduct such an analysis for the USEC. This absence of a coastal sedimentary record of the past tsunamis may be attributed to glaciers and frequent large tropical cyclones, which have caused significant erosion and geomorphic changes nearshore and onshore.

Consequently, in earlier NTHMP-USEC work, tsunami generation, propagation, and coastal impact were modeled for a series of extreme sources in the Atlantic Ocean basin causing Probable Maximum Tsunamis (PMTs), which were selected based on a variety of studies that can be found in references (Grilli et al. 2010, 2017a; Tehranirad et al. 2015). Hence, the resulting NTHMP inundation maps provide envelopes of maximum inundation from the combined coastal impact of these PMTs, without consideration of return periods (ECMAP 2017). In the earlier work, the generation and propagation of each PMT were simulated using three-dimensional (3D) non-hydrostatic (Ma et al. 2012) and two-dimensional

(2D) Boussinesq (Shi et al. 2012; Kirby et al. 2013) wave models, respectively. A series of nested spherical and Cartesian grids were used, with increasingly fine resolution towards the coast (down to 10-30 m meshes) and commensurately accurate bathymetric and topographic data.

This earlier work, as well as other studies, have shown that coastal hazard along the upper USEC is dominated by near-field tsunamis that could potentially be generated by large submarine mass failures (SMFs) (ten Brink et al. 2008, 2009a, b, 2014; Grilli et al. 2009, 2015, 2017a, b). The moderate seismicity typical of the region would not be expected to cause significant near-field coseismic tsunamis, but could trigger large SMFs where large amounts of sediment has accumulated over steep enough slopes, on or near the continental shelf break. This is particularly the case off of major estuaries such as Chesapeake Bay or the Hudson River (Fig. 1.1). In fact, the largest earthquake ever measured in the USEC area, with a Mw 7.2 magnitude in the the upper Laurentian Fan, was responsible for triggering the 1929 landslide and tsunami off of the Grand Banks (Piper et al. 1999; Fine et al. 2005; Løvholt et al. 2018). The large tsunami generated by this 100 km³ landslide caused a maximum runup of 13 m, 28 fatalities, and widespread destruction of coastal communities in Newfoundland, Canada. The SMF turned into a turbidity current that reached speeds of 17-28 m/s, breaking 12 underwater communication cables along its way [more details about this important historical case study can be found in Løvholt et al. (2018)]. Confirming that the 1929 landslide was not an isolated event, Chaytor et al. (2009) and ten Brink et al. (2014) mapped numerous paleo-SMFs on the US Atlantic continental shelf and margin, with the largest one being the Currituck slide complex (CSC) (Locat et al. 2009), off of Chesapeake Bay (Fig. 1.1). The CSC has been extensively studied from geological and slide triggering points of views (e.g., Locat et al. 2009, and

references herein). Tsunami generation from a reconstituted Currituck SMF was first modeled by Geist et al. (2009), using a simplified SMF tsunami generation model, and then by Grilli et al. (2015) using a more complete model. Both modeling studies showed that, if it had failed in present days, this paleo-SMF might have generated a destructive tsunami for the USEC, from the Carolinas to Massachusetts. Recent field work (Hill et al. 2017) has dated this old slide complex to 16-50 Ka, but Chaytor (personal communication, 2017) indicates that its age is likely to be on the younger end of this range. Considering when this failure occurred, the sea level would have been much lower than in present days.

Figure 1.1: Geography of study area [with marked state limits and names, and names of a few cities (red stars)]. Areas 1–4, identified by Grilli et al. (2015) as having high potential for large tsunamigenic SMFs, and the location of the historical Currituck slide complex are marked by yellow ellipses. Numerical gauge stations a–g (Table 1.3) are marked by black bullets. Color scale and contours show bathymetry in meters.

When considering coastal hazard assessment for SMF tsunamis, many associated sources of uncertainty should be considered, related to SMF triggering, source

characteristics, rheology/deformation, law of motion, epistemic uncertainties in the model, and others. With the goal to evaluate the first order SMF tsunami hazard along the USEC, Grilli et al. (2009) attempted to account for many of these sources of uncertainty by performing Monte Carlo Simulations (MCS) of slope instabilities triggered by seismicity and causing tsunami generation and coastal impact. This study, which involved many assumptions and simplified models, was initially carried out from southern New Jersey to Cape Cod, MA, but was later extended to southern Florida (Krauss 2011). Two main failure mechanisms were considered, translational slides or slumps, both assumed to be rigid (i.e., solid block motion), in noncohesive or cohesive sediment, respectively. Earlier surveys of the USEC margin by Booth et al. (2003) identified 43% of rotational failures (slumps) versus 57% of translational failures (slides). In cohesive sediment, the former can be considered as undertaking a mostly rigid motion; hence, slumps, although less frequent, are a realistic failure mode on the USEC. MCS results provided 100 and 500 year return period SMF tsunami runups along the entire USEC, as well as the respective percentages of slides and slumps, which were found to be consistent with field surveys Booth et al. (2003). MCS results predicted a 500 year maximum runup of 5-6 m north of Virginia and a significantly reduced runup (1-2 m) south of it. This study, however, did not consider whether there was sufficient sediment accumulation for large SMFs to occur off the coast, where large runup was predicted. To this effect, Eggeling (2012) performed geophysical and geotechnical analyses of seafloor data collected in regions facing coastal areas with the largest runup, which led to selecting 4 areas for further consideration and modeling (Areas 1-4 in Fig. 1.1). These areas typically have a large bottom slope and a sediment thickness sufficient to make a large tsunamigenic SMF possible, given sufficient seismicity. The likelihood for failure was confirmed by more detailed slope stability analyses per-

formed using SLIDE (2017). These yielded low factors of safety of bottom slopes in these areas in the presence of the maximum expected seismicity obtained using the USGS 2002 Interactive Deaggregation Tool (USGS 2002). Details can be found in Eggeling (2012), but overall, the mean factors of safety for transects in Areas 1-4 were found to be as low as 0.82-0.92 (with standard deviation of 0.11-0.12), yielding probabilities of failure of 73-88%.

In light of these results, and considering the many uncertainties on SMF mechanisms and the scarcity of geotechnical field data on the upper USEC shelf, the NTHMP modeling committee decided that the inundation mapping work for the USEC would be performed on the basis of SMF PMTs. Accordingly, Grilli et al. (2015) simulated tsunami generation from large SMFs sited in Areas 1-4 (Fig. 1.1), which were all parameterized based on the characteristics of the largest known historical failure in the region, the CSC, as a single large failure with a volume in the 128-165 km³ range estimated for the entire CSC by Locat et al. (2009).

For this reason, they were referred to as “Currituck SMF proxy” sources. In addition, to model worst case scenarios, each SMF was assumed to fail as a rigid slump, which, based on earlier work, was expected to maximize tsunami generation and coastal impact (Grilli and Watts 2005). Tsunami generation was computed using the 3D non-hydrostatic model NHWAVE (Ma et al. 2012) in which both the slump Gaussian-like geometry and rigid block law of motion were specified as bottom boundary conditions (details of this method, which is used in this paper to simulate rigid slumps, can be found in “Appendix”). The maximum SMF velocity used in the law of motion was that estimated by Locat et al. (2009) at $\simeq 35$ m/s, using a Bingham model. Once the slump stopped moving and tsunami generation was complete, simulations of tsunami propagation and coastal impact were continued using the 2D Boussinesq model FUNWAVE-TVD (Shi et

al. 2012) (referred to hereafter as FUNWAVE), in a series of nested grids. Results of this work were used to develop the NTHMP tsunami inundation maps currently released for the upper USEC region (ECMAP 2017).

To confirm that rigid slumps indeed lead to worst case scenarios for tsunami coastal hazard along the USEC, Grilli et al. (2017b) investigated effects of SMF deformation on tsunami generation. They used the recent two-layer version of NHWAVE by Kirby et al. (2016), in which deforming slides are simply modeled as a dense Newtonian fluid, in a depth-integrated lower layer two-way coupled to the upper water flow simulated with the standard multi-r-layer NHWAVE model. Earlier work based on simulations with the Bingham model BING concluded that it may not be necessary for proper tsunami generation to use a complex deforming slide rheology (Watts and Grilli 2003). Other work pointed out the importance of SMF mass, initial acceleration, and submergence for first-order tsunami generation, whereas SMF deformation should be second order (Grilli and Watts 2005). This justifies why more complex deformation models were not investigated here. Note, it is not the purpose of this paper to provide an exhaustive review of the various models and modeling methodologies that have been used to simulate underwater landslides and landslide tsunami generation, nor of historical landslide tsunami case studies. Fairly comprehensive reviews of such works to date can be found in Tappin et al. (2014), Kirby et al. (2016), and Grilli et al. (2015, 2017b). A broader scope recent review can be found in Yavari et al. (2016). Note that Løvholt et al. (2017) recently considered another mode of failure for giant slides, retrogressive failure, in which the landslide fails as a continuous sequence of blocks. They found that this failure mode could cause less tsunami generation than anticipated based on the slide size.

Grilli et al. (2017b) first revisited the simulation of the historical Currituck

SMF with this model, to estimate values of slide viscosity ν_s and bottom friction Manning coefficient n (between slide and substrate) yielding a slide center of mass motion $S(t)$ similar to that used by Grilli et al. (2015) to simulate rigid slumps. This ensured that maximum velocity during motion and runout S_f after a time of motion t_f were similar to those of the slumps and thus consistent with values derived from field observations by Locat et al. (2009). These simulations resulted in a fairly large viscosity, $\nu_s = 0.5 \text{ m}^2/\text{s}$, which was then used to simulate a Currituck SMF proxy in Area 1 (Hudson River canyon) as a deforming slide. For this case, maximum surface elevations simulated nearshore over the 5 m isobath were found to be smaller at most locations than those of the rigid slump, but nearly identical at some locations. Despite the fairly high viscosity, there was significant spreading of the deforming slide during its motion that led to a reduction of the acceleration during part of its initial motion and related maximum tsunami elevations. In addition, because the deforming slide flowed on its own (rather than having its motion prescribed), it followed the steepest bottom slope. Hence, the generated tsunami was more asymmetrical than for the slump, which affected both where the maximum tsunami impact occurred along the coast and the level of hazard. This preliminary study led Grilli et al. (2017b) to conclude that considering SMFs all failed as rigid slumps [as in Grilli et al. (2015)] was likely too conservative to estimate the coastal tsunami hazard along the USEC. Hence, modeling SMF tsunami generation and coastal hazard, with even a moderate level of deformation (i.e., here a high viscosity), should be more realistic, and also better cover the range of failures observed in the field (Booth et al. 2003; ten Brink et al. 2014).

Another aspect of the SMF tsunami hazard assessment relates to using proper physics in the wave generation and propagation models. In particular, one should assess whether using models that include frequency dispersion or non-hydrostatic

pressure effects is important for assessing tsunami coastal hazard. Earlier work addressed the importance of dispersion during the generation phase of SMF tsunamis when vertical seafloor accelerations as well as 3D effects may be significant (Grilli and Watts 1999, 2005; Grilli et al. 2002). This showed that the important parameter to consider is $\mu = kd$, where d denotes the water depth above the SMF center of mass initial location and $k = 2\pi/\lambda_0$, where λ_0 is the dominant wavelength of generated tsunami waves. The latter can be approximated as $\lambda_0 \simeq 2b$, with b the SMF downslope length. Thus, according to linear long-wave theory, initial SMF tsunami waves will be dispersive when $\mu > \pi/10$ or $d > b/10$. For such a case, for instance, Ma et al. (2012) simulated Enet and Grilli’s laboratory experiments for a rigid quasi-Gaussian shape slide moving down a slope using their 3D non-hydrostatic model NHWAVE, and showed a good agreement with time series of surface elevations measured at several wave gauges. Re-running simulations with the non-hydrostatic pressure terms turned off in the model led to large errors.

For the very large SMFs considered here, however, it will be shown that $d < b/10$ and, hence, the initially generated tsunami waves are not expected to be dispersive. During their subsequent propagation onshore and offshore, however, tsunami wave trains will experience increasing dispersive effects, the larger the magnitude of $\mu = kh$, here based on ocean depth h , with $k = 2\pi/\lambda$ and λ the local tsunami wavelength. The cumulative effects of dispersion will gradually become larger, the longer the propagation distance L . Assuming regular waves propagating over constant depth, Glimsdal et al. (2013) quantified dispersive effects by way of a “dispersion time” $\tau = 6h^2L/\lambda^3$ and showed that these typically become significant when $\tau > 0.1$. For landslide tsunamis, they pointed out as discussed above that the initial wavelength λ_0 is more significant to dispersion than depth or propagation distance [see also Løvholt et al. (2015) for additional considerations

and discussions]. Dispersive effects also affect long waves propagating shoreward, into shallow water, causing the eventual occurrence of undular bores around steep crests (e.g., Madsen et al. 2008; Glimsdal et al. 2013), which may locally nearly double maximum tsunami elevation and hence affect the coastal tsunami hazard. Glimsdal et al. (2013) indicated that undular bores may break before reaching the shore, as they are made of short and steep waves; hence, their effect on maximum elevation is more uncertain. However, breaking waves may increase flow velocity and impulse force. To be able to resolve such short waves in numerical simulations, tsunami models require fine resolution nearshore grids. With a 32 m resolution grid, Grilli et al. (2015) were barely able to simulate the initial stages of the appearance of undular bores in SMF tsunamis propagating into Chesapeake Bay. Using one-dimensional simulations, Grilli et al. (2013a) showed that to fully resolve such undular bores at this location, the resolution needed to be about 5 m.

Earlier work has shown a significant effect of bottom friction on tsunami propagation. Not only is tsunami inundation affected by land cover roughness (Jakeman et al. 2010; Kaiser et al. 2011), but for wide continental shelves, tsunami elevations can significantly decrease due to bottom friction (Geist et al. 2009; Tehranirad et al. 2015; Grilli et al. 2015). Indeed, based on linear long-wave theory, the depth averaged tsunami-induced velocity $U \propto h^{-3/4}$, for a depth h . Hence, energy dissipation due to bottom friction $D \propto C_d U^3$. Grilli et al. (2015) performed tsunami propagation simulations with FUNWAVE using a constant bottom friction coefficient $C_d = 0.0025$ (Shi et al. 2012); hence, bottom friction dissipation was $D \propto C_d h^{-9/4} \propto h^{-2.25}$. In more recent nearshore tsunami simulation work with FUNWAVE (Shelby et al. 2016), a more realistic depth-dependent friction coefficient was used based on Manning’s formula, $C_d = gn^2/h^{1/3}$, function of the Manning n parameter; this yields larger values of the bottom friction dissipation,

$D \propto h^{-31/12} \propto h^{-2.58}$, i.e., energy dissipation more than quadratically increases as depth decreases when long waves propagate onshore into shallower water.

In this paper, the landslide tsunami hazard is more realistically assessed along the upper USEC by considering effects of slide deformation on tsunami generation. Similar to Grilli et al. (2017b), tsunami generation and propagation are modeled for Currituck SMF proxies sited in Areas 1-4 (Fig. 1.1) off of the upper USEC. Tsunami generation is compared for SMFs failing as deforming slides or rigid slumps, with the latter being re-simulated to compare results for both types of SMFs in the same higher resolution coastal grids used here. Bottom friction is modeled using $n = 0.025 \text{ s}^2 / \text{m}^{2/3}$ throughout FUNWAVE simulations, which corresponds to a coarse sand substrate. However, since sandy and gravel/boulder substrates can be found in various areas of the USEC a limited comparison is performed for tsunami inundation computed with $n = 0.025$ or 0.0375 (which reflects the rougher substrate). This is done in one of the most impacted areas of the USEC, in northern New Jersey and western Long Island, NY. Finally, by re-running selected simulations without dispersion, the effects of dispersion are evaluated on tsunami propagation (both far- and near-field) and coastal hazard (maximum inundation height, velocity, and impulse force). Tsunami simulations for the four considered areas allow assessing the tsunami hazard from extreme SMFs along the USEC from Virginia to Cape Cod, MA. Results are provided as instantaneous surface elevations maps, time series of surface elevation at numerical wave gauges, envelopes of tsunami surface elevations for single or the combined SMFs, and percent differences between envelopes of results without and with dispersion.

Section 2 summarizes the modeling methodology. Section 3 presents simulations of SMF tsunami generation with NHWAVE and Sect. 4 of SMF tsunami

propagation and coastal impact with FUNWAVE. Section 5 is devoted to evaluating the importance of dispersive effects on the tsunami hazard and is followed by a discussion in Sect. 6, and conclusions in Sect. 7.

1.2 Modeling Methodology

1.2.1 Numerical Models

Tsunami generation by SMFs is modeled with the 3D non-hydrostatic model NHWAVE. For rigid SMFs, the geometry, velocity, and acceleration are specified as bottom boundary conditions, following the methodology of Ma et al. (2012) and Grilli et al. (2015) [see also, Grilli and Watts (2005), Watts et al. (2005) and Enet and Grilli (2007); a summary of this methodology is given in “Appendix”]. For deforming slides, the two-layer version of NHWAVE is used with the lower layer representing the SMF as a dense Newtonian fluid and the upper-layer water domain being divided into multiple σ layers (Kirby et al. 2016; Grilli et al. 2017b). In the bottom layer, equations of mass and momentum conservation are depth-integrated, similar to long-wave models. Both volumetric (i.e., viscous) and bottom friction (i.e., basal shear) dissipation terms are included in the equations. This is different from the standard depth integrated models, which typically only include a single bottom friction term. Modeling experiments for slides made of glass beads, Grilli et al. (2017b) observed that results were not very sensitive to slide viscosity, perhaps due to this parameterization of dissipation. Hence, they simulated a deforming Currituck SMF proxy in Area 1 using a fairly high viscosity $\nu_s = 0.5 \text{ m}^2/\text{s}$, which is in the range of suggested values for debris flows ($\nu_s = 0.2\text{-}0.6 \text{ m}^2/\text{s}$), and compared results obtained for various substrate friction coefficients, $n = 0.05, 0.10$, or $0.15 \text{ s}^2/\text{m}^{2/3}$ to those of the rigid slump (Grilli et al. 2015). They found that $n = 0.1$ led to approximately the same maximum runout of the slide center of mass at time $t_f = 11.9 \text{ min}$ as for rigid slumps, and a similar velocity of their center of mass

during the first 10 min of each SMF motion. [Note that Grilli et al. (2017b) used this ms value in their simulations, but wrongly listed the dynamic viscosity in their paper as $\mu_s = 500 \text{ kg}/(\text{m s})$, while it should have been $950 \text{ kg}/(\text{m s})$, based on bulk density.]

As in earlier work, the initial geometry of both rigid and deforming SMFs is modeled as a sediment mound of quasi-Gaussian cross sections with an elliptical footprint over the slope (see, Enet and Grilli (2007); Grilli et al. (2015), and Eqs. 1–3 in “Appendix”, for details). In all cases, NHWAVE uses a horizontal Cartesian grid of resolution $(\Delta x, \Delta y)$ and a boundary fitted σ -coordinate grid with N_σ layers in the vertical direction.

Once the tsunami is fully generated, simulations of wave propagation are pursued in the 2D fully nonlinear and dispersive Boussinesq model FUNWAVE (this will be detailed below). As in earlier work, these simulations are performed by one-way coupling in a series of nested grids of increasingly fine resolution in shallower water towards the coast [see, e.g., Grilli et al. (2013b, 2015, 2017a, b), Tappin et al. (2014) and Shelby et al. (2016), for details and examples of this approach]. When simulating tsunami propagation in regional grids of small geographic extent, the Cartesian coordinate version of FUNWAVE is used (Shi et al. 2012), whereas when simulating propagation in a grid at the scale of the northern Atlantic ocean, the spherical version of the model is used (Kirby et al. 2013).

NHWAVE and FUNWAVE are non-hydrostatic, i.e., dispersive, wave models, in which dispersion can be turned off as an option. As discussed in the introduction, earlier studies have shown the importance of including frequency dispersion effects to model SMF tsunami generation and propagation, particularly when the initial SMF footprint is small compared to depth (e.g., Ma et al. 2012). In this

case, the generated tsunami waves have a relatively small wavelength compared to depth (e.g., Grilli and Watts 1999, 2005; Glimsdal et al. 2013). Moreover, when simulating the long distance (far-field) propagation of any tsunami [i.e., for a large “dispersion time” τ in the sense of Glimsdal et al. (2013)], constructive- destructive wave-wave interactions will be affected by dispersion, potentially leading to significant differences in wave height and phase as compared to non-dispersive simulations (Kirby et al. 2013). Finally, when tsunamis propagate onshore in shallow water, dispersion leads to the development of undular bores near their wave crests. This was observed for many tsunamis: (1) experimentally (e.g., Matsuyama et al. 2007 and references listed in Grilli et al. 2017b; (2) in the field (e.g., Madsen et al. 2008; Tappin et al. 2014); or (3) numerically (e.g., Løvholt et al. 2008; Madsen et al. 2008; Abadie et al. 2012; Glimsdal et al. 2013; Grilli et al. 2013a, 2015; Tehranirad et al. 2015).

1.2.2 Methodology and Computational Grids

Table 1.1: Parameters of NHWAVE grids ($\Delta x = \Delta y = 500$ m resolution; $N_\sigma = 5$ σ layers; 1000 by 1000 cells), and locations (center) of SMFs of length $b = 30$ km, width $w = 20$ km, and thickness $T = 0.75$ km, in Areas 1-4 (Figs. 1.1, 1.2a). For rigid slumps, the average local slope is assumed to be $\alpha = 4^\circ$ (Fig. 1.3).

Grids and SMFs	Area 1	Area 2	Area 3	Area 4
SW corner (Lon., Lat.)	(-75.00, 36.90)	(-74.32, 37.47)	(-75.97, 36.12)	(-76.37, 35.80)
Center (x_0, y_0)	(-72.19, 39.19)	(-71.49, 39.76)	(-73.19, 38.41)	(-73.60, 38.09)
Azimuth θ (CW from N)	136 °	153 °	140 °	126 °

For rigid slumps, which are rotational, tsunami generation is completed when they stop moving upon reaching their maximum runout S_f at time t_f . By contrast, deforming slides keep “flowing” down the continental slope for a much longer time. However, they may no longer be tsunamigenic when becoming thin and/or reaching deep enough water. Here, no measurable wave generation was observed in NHWAVE for $t \geq 20$ min in all deforming slides simulations; hence, this time was

Figure 1.2: Horizontal footprints of computational grids, with (red) FUNWAVE $\Delta x = \Delta y = 500$ m resolution Cartesian grid G0 encompassing the SMFs and their grids, and (a) 4 color coded $\Delta x = \Delta y = 500$ m horizontal resolution nested Cartesian NHWAVE grids (with $N_\sigma = 5$ vertical σ -layers), used to simulate motion and tsunami generation for SMFs sited in Areas 1–4 (numbered ellipses; Table 1.1) and (b) 7 labeled $\Delta x = \Delta y = 120$ m resolution Cartesian FUNWAVE grids (OC, AC, NJ, NY, MO, RI, and NA from southwest to northeast; Table 1.2)

selected to initialize FUNWAVE. This model, however, assumes a steady bottom but, at $t \geq 20$ min, deforming slides are still slightly moving and thus imposing a time-dependent lower boundary condition to the upper water layer in NHWAVE. In

Table 1.2: Parameters of FUNWAVE computational grids (Figs. 1.2, 1.14a, 1.16a, 1.21a)

Grid name	Cells (M_x, N_y)	Resol.	SW corner (Lat., Lon.)
Base grid (G0)	(1600,1600)	500 m	(34.82N,76.37W)
Ocean City (OC)	(672,1104)	120 m	(37.68N, 75.60W)
Atlantic City (AC)	(752, 1104)	120 m	(38.70N, 75.04W)
New Jersey (NJ)	(720, 936)	120 m	(39.70N, 74.40W)
New York (NY)	(1212, 792)	120 m	(40.49N, 74.21W)
Montauk (MO)	(780, 744)	120 m	(40.60N, 72.60W)
Rhode Island (RI)	(816, 828)	120 m	(40.90N, 71.65W)
Nantucket (NA)	(828, 832)	120 m	(40.81N, 70.67W)
North Atlantic (ATL)	(5236, 3136)	1 arc min	(5.903N, 88.154W)
Seaside heights (NJ30)	(984, 984)	30 m	(39.92N, 74.17W)
Ocean City (OC30)	(984, 984)	30 m	(38.26N, 75.23W)
Seaside heights (NJ7.5)	(1200,1200)	7.5 m	(39.91N, 74.13W)

their simulations of a deforming slide in Area 1, Grilli et al. (2017b) observed that this boundary condition mismatch caused the generation of spurious (rebound) waves in FUNWAVE, which they eliminated by filtering out velocities over the slide location before initializing simulations. Because at this time, the slide was far offshore and no-longer tsunamigenic, and this filtering did not affect simulations of the tsunami onshore propagation (as was verified by performing a longer simulation with NHWAVE). A similar method is applied here, in combination with a larger grid (details of this procedure are given later).

Tsunami generation from each SMF sited in Areas 1–4 is computed with NHWAVE, using four distinct Cartesian grids with $\Delta x = \Delta y = 500$ m horizontal resolution and $N_\sigma = 5$ σ -layers centered around each SMF initial location (x_0, y_0) (Fig. 1.2a; Table 1.1). Each grid covers a 500 by 500 km footprint (hence, it has 1000 by 1000 cells; see grid SW corner coordinates in Table 1.1) and is nested within FUNWAVE’s $\Delta x = \Delta y = 500$ m Cartesian grid G0 (Fig. 1.2). This first-level (or base) grid, in which simulations of tsunami propagation are continued, covers an 800 km by 800 km area (hence, it has 1600 by 1600 cells; see details in

Table 1.2). To eliminate reflection along offshore boundaries, 125 km-thick sponge layers are specified on both the eastern and southern boundaries of grid G0. At $t = 20$ min, the free surface elevation, bottom bathymetry, and horizontal velocities (u, v) [at the required $0.531h$ depth to ensure extended dispersion properties in FUNWAVE; see Shi et al. (2012)] computed in NHWAVE are interpolated onto G0 and used as initial conditions to continue simulations for each SMF. For the onshore tsunami propagation, simulations are then continued with FUNWAVE by one-way coupling in seven nested Cartesian grids of $\Delta x = \Delta y = 120$ m resolution (Fig. 1.2b; Table 1.2). To study dispersive effects nearshore, additional finer nested grids are used in FUNWAVE, with 30 m (NJ30, OC30) and 7.5 m (NJ7.5) resolutions (Table 1.2; see details later). Likewise, to study dispersive effects on far-field tsunami propagation, simulations are also continued with FUNWAVE in the 1 arc-min (about 1800 m) resolution spherical grid ATL, which covers the North Atlantic ocean; this large grid has 5236 by 3136 cells in the longitudinal and latitudinal directions, respectively (Table 1.2), and uses 150 km-thick sponge layers along the outer boundaries. Simulations in the ATL grid are also initialized using NHWAVE results computed at $t = 20$ min (details of this grid are given later as well).

The bathymetry used in the 500 and 120 m resolution grids (Table 1.2) was interpolated from the 3-arc-s (~ 90 m) resolution Northeast Atlantic and Southeast Atlantic US Coastal Relief Models (NGDC 1999a, b), wherever available (mostly up to the shelf break); otherwise, data were obtained from the 1 arcmin resolution ETOPO-1 Global Relief Model (Amante and Eakins 2009). The bathymetry of the 30 m NJ30 and 7.5 m NJ7.5 resolution grids, centered on Seaside Heights, NJ, was interpolated from the 1/9 arc-s (~ 3 m) post Hurricane Sandy survey data (NCEI 2014) and that of the 30 m OC30 grid, centered on Ocean City, MD, from Grothe

et al. (2010) (Table 1.2). For each Cartesian grid, the bathymetry/topography was constructed using a Mercator transverse projection centered on each grid center. The ETOPO1 bathymetry in the study area is plotted in Fig. 1.1, as both a color scale and contour lines. North of North Carolina, it shows the typical pattern of a wide shallow USEC shelf (depth $h \leq 100$ m), a steep shelf break slope, and a deep abyssal plain ($h = 2000\text{--}5000$ m). Similar to the Currituck slide complex, the four study Areas 1–4 are all located on the shelf break slope.

1.3 SMF Tsunami Generation with NHWAVE

1.3.1 SMF Geometry, Rheology, and Kinematics

Tsunami generation from Currituck SMF proxies sited in Areas 1–4 off of the USEC (Fig. 1.1) is modeled with NHWAVE, assuming each SMF fails either as a rigid slump [similar to Grilli et al. (2015)] or as a deforming slide [similar to that considered in Area 1 by Grilli et al. (2017b)]. Consistent with earlier work (Grilli et al. 2015), each SMF is assumed to have a quasi-Gaussian geometry with an initial elliptical footprint of downslope length $b = 30$ km, width $w = 20$ km, and thickness $T = 0.75$ km (Enet and Grilli 2007; Grilli et al. 2015, 2017b). [These values were inferred by Grilli et al. (2015) from field data of the CSC, for which the total material in the two main failure lobes identified by Locat et al. (2009) were lumped into a single initial failure originating in the headwall scar located up the slope.] Using a shape coefficient $\epsilon = 0.717$, this yields a SMF volume $V_s = 158 \text{ km}^3$, which is in the range of the total volume estimated for the entire Currituck slide complex (see details in “Appendix”). The initial SMF geometry is specified below current seafloor in the model (Fig. 1.3), with the elliptical footprint centered at longitude–latitude (x_0, y_0) and its major axis oriented downslope in an initial azimuthal direction of motion θ (Table 1.1). Figure 1.3 shows vertical cross sections in direction θ through each SMF sited in Areas 1–4 (both seafloor

bathymetry and initial/later SMF profiles). The SMFs are all initially located on the continental shelf slope, with the water depth above their initial center of mass location varying between $d = 1000$ and 1800 m. Hence, $d < b/10 = 3$ km and the dominant SMF tsunami waves that will be generated are expected to be in the long-wave regime. The initial elliptical footprints of each SMF are plotted at scale in Fig. 1.4; the rigid slumps are all specified to move in the direction of the major axis of each ellipse.

As in earlier work (Grilli and Watts 2005; Grilli et al. 2015, 2017b), for rigid slumps, the bottom boundary conditions are specified analytically as a function of SMF geometry and other parameters, assuming a pendulum-like motion parallel to slope, $S(t)$. [Note, it is not the scope of this paper to revisit this earlier work on solid block motion, whose details can be found in reference; a summary is provided in “Appendix” (see Eq. 4).] In this motion, maximum runout S_f and duration t_f are computed as a function of slump parameters: beside geometry, these include bulk sediment density $\rho_s = 1900$ kg/m³, water density $\rho_w = 1025$ kg/m³, average local slope α , radius of (pendulum) motion R , and (hydrodynamic) added mass coefficient C_M . In the transects of Fig. 1.3, the average seafloor slopes are 3.2deg, 3.0deg, 4.0deg, and 3.1deg, in Areas 1–4 SMFs, respectively. However, considering that there are steeper parts in each transect, $\alpha = 4$ deg is used for each case, which also yields identical maximum runout and time of motion in each case. As in Grilli et al. (2015, 2017b), $C_M = 1$ is used and the radius is estimated as, $R \sim b^2/(8T) = 150$ km. With these parameters, Eqs. (4)–(8) in “Appendix” yield $S_f = 15.8$ km and $t_f = 715$ s (11.9 min). Figure 1.3 shows the final profile of each rigid slump in Areas 1–4 after they have moved a distance S_f down the slope (for $t = t_f$).

The motion of deforming slides is modeled with the two-layer NHWAVE, which computes it together with tsunami generation based on the same initial geometry

Figure 1.3: Vertical bathymetric transects in each NHWAVE grid (Fig. 2a), through Areas 1–4 (plots (a)–(d)) SMF centers (x_0, y_0) , in azimuthal direction α (Table 1.1): (solid black) current bathymetry; (dash blue) initial SMF profile; (solid red) final slump profile after displacement (runout) $S_f = 15.8$ km at $t_f = 715$ s (11.9 min); (solid green/dash green) deforming slide profiles at $t = 715$ and 1200 s (20 min) [note, slides continue moving beyond this time but are no longer significantly tsunamigenic.] The x-axis measures distances from each SMF center; vertical exaggeration is 25 times.

and bulk sediment density as for slumps and on a specified SMF viscosity (assuming a dense Newtonian fluid) and substrate-to-slide friction. Following the above discussions, the kinematic viscosity is set to $\nu_s = 0.5 \text{ m}^2/\text{s}$ and slide-to-substrate Manning friction coefficient to $n = 0.1 \text{ s}^2/\text{m}^{2/3}$, as in earlier work by Grilli et al. (2017b). This will make deforming slide kinematics consistent with that of rigid slumps, allowing to compare tsunami generation and coastal impact for both SMF types. Figure 1.5 shows the kinematics of each SMF center of mass, i.e., motion $S(t)$, velocity $U(t)$, and acceleration $A(t)$, computed during NHWAVE simulations, for rigid slumps and deforming slides. For the former, equations of the specified kinematics are listed in “Appendix”; for the latter, kinematics is derived from calculating each deforming slide center of mass location as a function of time in NHWAVE results.

Figure 1.3 shows vertical cross sections through SMFs computed in Areas 1–4, in their initial direction of motion θ . Profiles are shown at time $t_f = 715 \text{ s}$ when rigid slumps stop moving, and at $t = 1200 \text{ s}$ (20 min) when NHWAVE’s solution is moved into FUNWAVE as an initial condition. As expected, the deforming slides’ center of mass locations at $t = t_f$ appears to be similar to those of the rigid slumps. However, while rigid slumps have preserved their geometry and thickness all the way to the end of their motion, the deforming slides have flowed both downslope and sideways, leading to a significant reduction in thickness. For $t > t_f$, the deforming slides continue flowing down the slope, some by a considerable distance, but due to both their reduced thickness and the large depth, they are no longer significantly tsunamigenic. Figure 1.4 confirms this behavior by showing, for Areas 1–4, the final footprints of the rigid slumps at 715 s and those of the deforming slides at 20 min. These different behaviors during motion are further detailed in Fig. 1.6, which show the instantaneous 3D geometry at $t = 5, 10$, and 15

Figure 1.4: Free surface elevations (color scale in m) simulated at $t = 800$ s (13.3 min) with NHWAVE, in 500 m grids (Fig. 1.2a; Table 1.1), for Currituck SMF proxies sited in Areas 1–4 (from top to bottom; Fig. 1.1), modeled as: (a, c, e, g) rigid slumps (similar to Grilli et al. 2015); or (b, d, f, h) deforming slides (with $\nu_s = 0.5$ m/s², $n = 0.1$). Black ellipses mark: (solid) initial footprints of each SMF (Table 1.1); (dashed) SMF footprints at 715 s for the slumps and 20 min for the slides. All SMFs have a volume $V_s = 158$ km³, density $\rho_s = 1900$ kg/m³, and similar runout S_f at $t_f = 715$ s (when the slumps stop moving)

Figure 1.5: Kinematics of Currituck SMF proxies sited in Areas 1–4 (Fig. 1.1), during tsunami generation. SMF center of mass: a motion; b velocity; and c acceleration, specified for rigid slumps (black) (based on Eqs. 4–8 in “Appendix”), and computed for deforming slides (for $\nu_s = 0.5 \text{ m}^2/\text{s}$ and $n = 0.1$), in Area: (orange) 1; (green) 2; (brown) 3; (blue) 4

min, of the Currituck SMF proxy modeled in Area 1 with NHWAVE, either as a deforming slide or a rigid slump. The SMFs both start moving from the same initial

Figure 1.6: Instantaneous 3D geometry (grey volume) of Currituck SMF proxy in Area 1 (Fig. 1.1), simulated with NHWAVE in a 500 m resolution grids with $N_\sigma = 5$ vertical layers (Fig. 1.2a), modeled as a: (a, c, e) deforming slide; or (b, d, f) rigid slump, at $t =$ (a, b) 5, (c, d) 10, and (e, f) 15 min. The vertical axis denotes depth in meter. Note, the rigid slump reaches its maximum runout at $t = 11.9$ min (hence this is the situation for f), while the slide keeps moving beyond $t = 15$ min (see Fig. 1.3)

elliptical footprint and geometry below seafloor, on the West side of the Hudson canyon apron (visible on the figures), but as time increases the deforming slide continuously flows down the steepest slope, while the rigid slump undertakes its forced pendulum motion in direction θ . In the case considered here, it is apparent that the deforming slide sediment flows asymmetrically, more towards the west side, whereas the rigid slump stays symmetrical with respect to its axis of motion.

This will lead to different alongshore tsunami generation in each case (see below). Finally, as the slump stops moving at $t_f = 11.9$ min, there is not much additional motion between $t = 10$ and 15 min visible in Fig. 1.6d, f.

The SMF kinematics displayed in Fig. 1.5 are consistent with these observations. The theoretical laws of motion of rigid slumps are plotted as a reference. These are identical for Areas 1–4, since slumps are all similarly parameterized to reach a maximum runout and stop moving at $t_f = 715$ s. They reach a maximum velocity $U_{max} = 34.7$ m/s at mid-course and have a maximum acceleration $A_0 = 0.153$ m/s² at $t = 0$, with an identical negative value at the end of their rigid motion. By contrast, the center of mass kinematics of deforming slides computed during NHWAVE simulations is site specific to each area and thus different for each slide. Figure 1.5c shows that, up to $t = 60$ – 100 s depending on the considered area, their acceleration is larger than that of rigid slumps. In Areas 1 and 2, whose seafloor features a simple fairly convex continental shelf slope (Fig. 1.3a, b), deforming slides reach a maximum velocity during motion and runout at t_f similar to those of rigid slumps (Fig. 1.5a, b). This is expected from their parameterization, consistent with the work of Grilli et al. (2017b). In Areas 3 and 4, however, whose shelf slope is concave, despite a motion similar to the other two areas for $t \leq 350$ s, the slides slow down quicker and reach a smaller runout at time t_f . This is clearly seen in the transects of Fig. 1.3c, d, where the slide material ends up filling the bottom of the initial cavity, which is below the deeper part of the shelf slope; hence, only a smaller part of each slide continues moving more slowly down the slope. It will be shown that this pattern of velocity and acceleration of deforming slides will lead to the generation of smaller onshore propagating tsunami waves, but larger offshore propagating tsunami waves compared to those of rigid slumps.

1.3.2 Tsunami Generation

Figure 1.4 shows free surface elevations simulated with NHWAVE at $t = 800$ s (13.3 min) for Currituck SMF proxies of Areas 1–4 (Fig. 1.1), whose motion is modeled as: (left panels) rigid slumps, or (right panels) deforming slides. For each area, computations are performed in individual grids with 500 m horizontal resolution and $N_\sigma = 5$ vertical layers (Fig. 1.2a; Table 1.1). By design, rigid slumps stop moving at $t_f = 715$ s after covering a downslope runout distance $S_f = 15.8$ km (Fig. 1.3), and at this time, deforming slides have approximately reached the same runout (see Fig. 1.5). As a result of their similar kinematics, waves generated at $t = 800$ s by deforming slides have a similar horizontal pattern as those generated by rigid slumps. However, they have different elevations due to the lateral spreading and thickness reduction of deforming slides late in their motion (Figs. 1.3, 1.6). In all cases, onshore propagating tsunamis feature a 10–15 m-deep leading depression wave followed by an equally large crest; these leading waves are larger for the rigid slumps. Deforming slides generate more asymmetric waves (alongshore) than rigid slumps, as a result of the effects of bathymetry on their motion. In all cases in Fig. 1.4, the offshore propagating tsunamis feature concentric (cylindrical) waves, with a 5–10 m leading crest followed by a deeper trough. The size of this leading wave is function of the SMF initial acceleration, which is slightly larger for deforming slides (Fig. 1.5c); hence, the leading crest elevation is larger for deforming slides than for rigid slumps.

Although the rigid slumps stop moving and no longer generate waves at $t_f = 715$ s, simulations were performed with NHWAVE for both types of SMFs up to $t = 1200$ s (20 min), before continuing simulations in FUNWAVE, to make sure 3D effects had become negligible in the generated wave trains. At this time, Fig. 1.3 shows that deforming slides are quite thin (i.e., have a large length-to-

thickness ratios) and have reached a large depth (> 1500 m); hence, these are no longer significantly tsunamigenic. As indicated before, flow velocities are filtered out of NHWAVE results above each slide location to prevent triggering numerical instabilities when initializing simulations in FUNWAVE, due to a mismatch in bottom boundary condition.

1.4 Onshore Tsunami Propagation and Coastal Impact

Results of simulations with FUNWAVE of the onshore propagation and coastal inundation of tsunamis generated by rigid slumps and deforming slides sited in Areas 1–4 are detailed in the following. Simulations initialized at $t = 20$ min based on NHWAVE results are first performed in the 500 m resolution grid G0; then, these are continued by one-way coupling in 120 m resolution nearshore nested grids (Fig. 1.2b; Table 1.2). Results of additional simulations in higher resolution grids of the deforming slide in Area 1 are presented in the next Section, to analyze dispersive effects during nearshore propagation. Specifically, we use: (1) two 30 m resolution grids nested within their respective 120 m grids, NJ30 off of Seaside Heights, NJ and OC30 off Ocean City, MD and then (2) one 7.5 m resolution grid, NJ7.5, nested within NJ30 (Table 1.2). These results will also allow assessing the convergence with grid resolution of the computed maximum surface elevations near the coastline.

Bottom friction was neglected in NHWAVE simulations of tsunami generation, except for slide-to-substrate friction, which was modeled with a Manning $n = 0.1$ in the bottom layer of the model. This was acceptable since wave generation took place over short times in fairly large depth. In the FUNWAVE simulations of the onshore tsunami propagation over the wide USEC shelf, bottom friction effects, which are more significant, are specified as a function of a Manning coefficient. In the absence of site specific data, a coarse sand substrate was assumed in most

Figure 1.7: Spatial variation of friction coefficient $C_d = gn^2/h^{1/3}$ (color scale) used in FUNWAVE simulations, based on local depth h (contour lines in meter) using a Manning coefficient $n =$ (a) 0.025 over grid G0 (Fig. 1.2); (b, c) 0.025 and 0.0375 $\text{s}^2/\text{m}^{2/3}$ over grid NJ, respectively (Fig. 1.2b; Table 1.2). Note, a minimum depth of 0.1 m is assumed in the C_d calculations

simulations, for which $n = 0.025 \text{ s}^2/\text{m}^{2/3}$. However, the seafloor of the upper USEC shelf, whose geology is post-glacial, features many areas covered with gravels and boulders (McMaster 1960), for which the Manning coefficient would be larger, up to $n = 0.0375 \text{ s}^2/\text{m}^{2/3}$ or so. In other shallow water areas exposed to high wave energy from storms, fine particles could be preferentially transported offshore, leaving coarser sediment nearshore (Twichell et al. 1981). All other parameters being equal, using $n = 0.0375$ instead of 0.025 more than doubles the bottom

friction force and related energy dissipation. In the following, the sensitivity of coastal inundation to the magnitude of bottom friction (in relation with changes in substrate) is assessed by repeating simulations of the deforming slide sited in Area 1 in grids G0 and NJ using $n = 0.0375 \text{ s}^2/\text{m}^{2/3}$ (Table 1.2). The NJ grid was selected for this sensitivity study among the 120 m resolution grids, since it encompasses highly impacted areas in northern New Jersey and western Long Island (Fig. 1.2b). Figure 1.7 shows maps of C_d values computed over grid G0 using $n = 0.025$ and grids G0 and NJ, using $n = 0.025$ and 0.0375 . A minimum depth of 0.1 m was specified in the model, which based on Manning’s formula limits the value of C_d to $C_d^{max} = 21.14n^2$, i.e., 0.013 and 0.030 for $n = 0.025$ and 0.0375 , respectively. For the smaller n value (Fig. 1.7a, b), C_d is fairly low in deeper water (less than 0.0015), but increases over the shelf to more than 0.002 (reaching 0.003–0.005 in less than 10 m depth); using the larger n value (Fig. 1.7c), C_d is 2.25 times larger than in the first case.

1.4.1 Instantaneous Propagation and Time Series

Figure 1.8 shows a sequence of instantaneous surface elevations computed at $t = 10\text{--}150$ min (by steps of 20 min) for a Currituck SMF proxy sited in Area 1, assuming a deforming slide. [Four video files, with side-by-side animations of simulated surface elevations for each Currituck SMF proxy sited in Areas 1–4, modeled as a rigid slump or a deforming slide, propagating in grid G0, are provided as online supplementary material: EtaSMF-A1-SDnear.mp4, EtaSMF-A2-SDnear, EtaSMF-A3-SDnear, and EtaSMF-A4-SDnear (color scale in m).] Tsunami generation is modeled with NHWAVE (upper left panel) and the subsequent wave propagation in grid G0 is modeled with FUNWAVE (other panels). As time increases, the initially onshore propagating tsunami waves continue their propagation towards the nearest shores. At the same time, the northern and southern parts of

the offshore propagating waves gradually refract over the shelf break bathymetry and eventually orient themselves parallel to local isobaths, propagating onshore towards the upper and lower parts of the coastline. Although there is some decrease in elevation up and down the coast due to energy spreading, the onshore propagating waves remain quite large (several meters) over the entire study area. At 110 min, Fig. 1.8f shows a nearly continuous elevation or depression wave, spanning from south to north and propagating towards the coast. This wave approximately reaches the 20 m isobath at the same time at most locations (the 7 black bullets denote wave gauge stations a–g on the 20 m isobath; Fig. 1.1; Table 1.3). This implies that the SMF tsunami generated in the Hudson river canyon in Area 1 ends up affecting the coast over the entire study area. This observation also applies to tsunamis generated by SMFs sited in other Areas 2–4 (as can be seen in the videos provided as supplementary material).

Note that in FUNWAVE simulations of all SMF cases, it was observed that outbound waves were not fully absorbed by the 125 km wide sponge layers specified along the eastern and southern boundaries of grid G0, which caused some reflection. Faced with the same problem, Grilli et al. (2017b) had filtered out the outbound wave surface elevations and horizontal velocities, before continuing simulations in a finer resolution nested grid. Such filtering, however, affects the ability of outbound waves to properly refract up and down the shelf break and contribute to the coastal tsunami hazard. Here, to prevent affecting the tsunami impact up and down the coast, the footprint of grid G0 was extended by 400 km in both the east and south directions and simulations of each tsunami were pursued up to $t = 50$ min in this larger grid (stage of the third panel in Fig. 1.8). At this time, the outbound waves, which had propagated beyond the offshore boundary of grid G0, were simply truncated and simulations restarted with these waves in grid G0, up

Table 1.3: Name, location, and depth of deep and shallow water wave gauge stations (Figs. 1.1, 1.14a, 1.16a), where time series of surface elevations (Figs. 1.9, 1.15, 1.18 are computed in 500 m G0, 1 arc min ATL, 30 m NJ30, and 30 m OC30 grids (Figs. 1.2, 1.14a, 1.16a, 1.21a).

Number	Label	Lat. (deg. N)	Lon. (deg. E)	Depth (m)
a		37.678	-75.303	20
b		38.832	-74.682	20
c		39.744	-74.000	20
d		40.522	-73.391	20
e		40.807	-72.489	20
f		41.316	-71.492	20
g		41.313	-70.674	20
1	DART 44401	37.592	-50.030	5494
2	DART 44402	39.298	-70.659	2649
3	DART 41424	33.000	-72.660	5250
4	DART 41420	23.508	-67.287	5743
5	DART 41421	23.380	-64.112	5845
6		51.812	-13.077	985
7		45.163	-4.652	4556
8		38.675	-10.181	3349
9		29.713	-19.715	4676
10		39.856	-74.069	15
11		39.919	-74.062	15
12		39.990	-74.051	15
13		40.037	-74.037	15
14		38.2910	-75.0960	10
15		38.3565	-75.0648	10
16		38.4208	-75.0469	10
17		38.4784	-75.0452	10

to $t = 187$ min (3h7').

Figure 1.9 shows time series of surface elevations computed with FUNWAVE at wave gauge Stations a– g in grid G0 (Fig. 1.1) for tsunamis generated by the 8 Currituck SMF proxies sited in Areas 1–4 (both slumps and deforming slides). Note that these stations are all located at a 20 m depth on the boundary of one of the finer nested grids, but results shown in Fig. 1.9 were computed in the coarse 500 m resolution grid G0. Hence, higher frequency waves that might be

generated are not resolved here. In a following section, similar results computed at shallow water wave gauge stations located in finer resolution nested grids will be shown and discussed. Consistent with the instantaneous surface elevations shown in Fig. 1.8, depending on the SMF case and considered station, incoming tsunami waves feature either a leading depression or elevation wave, which in most cases has a N-wave shape. For the deforming slide in Area 1 (Fig. 1.8), for instance, the tsunami arrives at all stations as an elevation wave, except at station d, which is located most directly onshore of the source. Earlier work has shown that such long N-wave shaped waves typically develop undular bores in shallow water (a.k.a. dispersive shock waves), made of shorter oscillations occurring at wave crests and/or behind wave troughs (Matsuyama et al. 2007; Madsen et al. 2008; Grilli et al. 2013a, 2015). Here, grid G0 is too coarse to resolve undular bores, but these will appear in results of finer resolution grids presented later. Finally, depending on the considered SMF and station, the period of the leading wave varies between 6 and 29 min, i.e., from quite short to much longer, and these waves are followed by a train of multiple waves, which can be shorter or longer with many shorter waves riding on top of them.

1.4.2 Maximum and Minimum Surface Elevations, and Bottom Friction Effects

Figure 1.10a, b shows the combined envelope of maximum surface elevations computed for the 4 deforming and 4 rigid slump Currituck SMF proxies sited in Areas 1-4, respectively. These figures are based on results of simulations in grid G0 and in the 7 nested 120 m grids, whichever is available, using a Manning friction coefficient $n = 0.025$. While offshore surface elevations significantly differ for the two different SMF rheologies, particularly over their generation areas (in accordance with Fig. 1.4), patterns and directionality of maximum nearshore

Figure 1.9: Instantaneous free surface elevations (color scale in m) simulated with NHWAVE and FUNWAVE in grid G0 (zoom-in of Fig. 1.2) for the deforming Currituck SMF proxy in Area 1 (Figs. 1.1, 1.4b) at $t =$ (a) 10, (b) 30, (c) 50, (d) 70, (e) 90, (f) 110, (g) 130, and (h) 150 min. Black bullets mark locations of wave gauge stations a–g (Fig. 1.1; Table 1.3). Black contours mark depth in meters.

surface elevations are similar for both SMF types. This results from the strong control of the shelf bathymetry on the focusing and defocusing of long tsunami waves, which start significantly refracting in fairly deep water (Tehrani et al. 2015). This bathymetric control effect is more apparent in Fig. 1.11a, which shows

Figure 1.10: Combined envelope of maximum surface elevations (color scale in m) computed with FUNWAVE in 500 m grid G0 or 120 m nested grids wherever available (Fig. 1.2b; Table 1.2), for the four SMFs sited in Areas 1–4 (Fig. 1.1): (a) deforming slides; or (b) rigid slumps. The white lines mark the 5 m isobath along which maximum and minimum wave heights are computed (see, Fig. 1.12). Black contours mark depth in meter

the envelope of maximum surface elevation computed in the 120 m resolution grid NJ with $n = 0.025$ (Fig. 1.2b; Table 1.2), for the deforming Currituck SMF proxy

sited in Area 1 (Fig. 1.1), which is the same case as in Fig. 1.8. The Hudson River canyon bathymetry causes incident tsunami waves, which are nearly parallel to the local isobath ahead of the canyon in 100 m depth (Fig. 1.8, second panel), to gradually refract away from the canyon and the entrance of New York Harbor. As a result, waves are focused towards both the northern New Jersey and western Long Island areas. In particular, Fig. 1.11a shows maximum surface elevations of more than 4 and 2.5 m in the highly developed barrier beaches of Seaside Height, NJ and Long Beach, NY, respectively, with the lagoons being fully flooded behind these. This defocusing-focusing phenomenon was already pointed out in earlier work performed in this area by Grilli et al. (2010, 2015) and Shelby et al. (2016).

Figure 1.11: Envelopes of maximum surface elevations (color scale in m) computed with FUNWAVE in 120 m grid NJ (Fig. 1.2; Table 1.2), for the deforming Currituck SMF proxy sited in Area 1 (Fig. 1.1), with a Manning $n =$ (a) 0.025; or (b) 0.0375 $\text{s}^2/\text{m}^{2/3}$ (Fig. 1.7b, c). Black contours mark depth in meter

Bottom friction effects on tsunami coastal impact are assessed next by comparing results of simulations of the deforming SMF in Area 1 using a larger Manning friction coefficient $n = 0.0375$ to the above results using $n = 0.025$ (Fig. 1.7b, c). Figure 1.11a, b shows the respective envelopes of maximum surface elevation computed in the 120 m resolution grid NJ. While the patterns of wave focusing–defocusing caused by refraction are not significantly affected when increasing

Figure 1.12: (a), (b) Maximum and (c) minimum surface elevation computed in 120 m grids (Fig. 1.2b), along the 5 m isobath (Fig. 1.10) for tsunamis generated by: (blue) rigid slump in a Area 1, or b, c the superposition of slumps in Areas 1–4; and (red) deforming slide in a Area 1, or b, c the superposition of all deforming slides in Areas 1–4. All FUNWAVE simulations are performed with $n = 0.025 \text{ s}^2/\text{m}^{2/3}$; green line in plot a is deforming slide in Area 1 for $n = 0.0375 \text{ s}^2/\text{m}^{2/3}$ (Fig. 1.7). The distance s is the curvilinear distance along the 5 m isobath measured from its southern end. Labels mark entrance to: *OC* Ocean City, MD, *DB* Delaware Bay, *AC* Atlantic City, NJ, *SH* Seaside Heights, NJ, *NYH* New York Harbor, *MT* Montauk, NY, *LIS* Long Island Sound, *PJ* Point Judith, RI, *NB* Narragansett Bay, *BB* Buzzards Bay

n there is an overall reduction in maximum nearshore surface elevation, particularly along the barrier beaches skirting the coastline. Figure 1.12 better illustrates focusing/defocusing and bottom friction effects on the coastal tsunami hazard by plotting maximum surface elevations computed along the 5 m isobath (marked as a white line in Fig. 1.10) in the seven 120 m resolution grids for the two SMF types, as a function of the curvilinear distance s measured along the isobath (from Chincoteague Island, VA, south of Ocean City, MD to Cape Cod, MA).

Figure 1.12a first shows results for the Currituck SMF proxy sited in Area 1 simulated in grid NJ: (1) rigid slump with $n = 0.025$ and (2) deforming slide with either $n = 0.025$ (Fig. 1.11a) or $n = 0.0375$ (Fig. 1.11b). As mentioned before, the pattern of higher and lower maximum surface elevations is nearly identical for the three cases. However, as already noted by Grilli et al. (2017b), given the same friction coefficient $n = 0.025$ and a single high resolution grid, the maximum elevations caused by the deforming slide are nearly half those of the rigid slump, and there is another small reduction in elevation when using the larger friction coefficient in the former case. Figure 1.12b shows similar results, but for the combined envelope of the rigid slumps or deforming slides sited in Areas 1–4, simulated in the seven 120 m resolution nested grids (as in Fig. 1.10). Here, the picture appears more complex: due to wave refraction and wave-guiding effects over the complex geometry of the shelf break, as discussed above, all SMF sources can generate tsunamis that affect the entire upper USEC in some manner. Due to the larger onshore propagating waves, they generate, rigid slumps cause the largest inundation at all sites, with a 11.5 m maximum around the eastern end of Long Island near Montauk, NY (Fig. 1.1). [This maximum impact occurs onshore of a marked ridge in the shelf bathymetry that causes high wave focusing (see, e.g., Fig. 1.1)]. However, because deforming slides generate larger offshore propagating

waves, which upon refracting on the shelf break can affect distant areas, the reduction in coastal impact shown in Fig. 1.12a for a single slide and a single small grid is overall less marked (except along the southern coast of Long Island west of Montauk). For deforming slides, the maximum inundation is reduced to 7 m in Montauk and is largest, 4.5–8.5 m, between Atlantic City and Seaside Heights, NJ (Fig. 1.1). In these combined envelopes, the pattern of wave focusing and defocusing repeats itself, with the lowest impact occurring in or near Bays and estuaries (e.g., Delaware Bay, New York Harbor, Narragansett Bay, Buzzards Bay).

Finally, Fig. 1.12c shows the envelope of minimum surface elevations computed along the 5 m isobath for the same cases as in Fig. 1.12b. Such results would be important, for instance, when considering the tsunami impact on the fresh water intakes of a power plant, of which there are a few in this area (e.g., nuclear power plants in Millstone, CT near the mouth of Long Island Sound or Oyster Creek, NJ, just south of Seaside Heights). Here, results appear quite similar for the rigid slump and deforming slide cases; at a few instances, in particular near Atlantic City and Montauk, the minimum elevation would reach down to the local seafloor (i.e., 5 m, with some minor irregularities related to discretization and the moving shoreline algorithm). This similarity in minimum values is expected, since the minimum coastal elevation is typically caused by the depression wave that first arrives on the nearest shores facing Areas 1–4 (Figs. 1.1, 1.9), and it is a result of the initial motion of each SMF, i.e., at short time when deformation has not yet played an important role, but instead SMF mass, initial acceleration, and local slope control wave generation (Grilli and Watts 2005).

1.5 Importance of Dispersive Effects on the SMF Tsunami Hazard

In earlier work, frequency dispersion effects were shown to significantly affect tsunami generation by SMFs and subaerial slides when their downslope length b

is relatively small with respect to the initial water depth d : as discussed in the introduction, when $b/d < 10$ (e.g., Grilli and Watts 2005; Ma et al. 2012; Abadie et al. 2012). In the far field, dispersion causes changes in surface elevations and lag times that increase with propagation distance (Madsen et al. 2008; Glimsdal et al. 2013; Kirby et al. 2013), and are particularly significant for initially dispersive SMF tsunamis (e.g., Løvholt et al. 2008, 2015; Abadie et al. 2012; Tehranirad et al. 2015). In onshore propagating tsunamis, dispersive effects cause the appearance of undular bores made of shorter waves that may significantly increase nearshore wave elevations and affect the coastal tsunami hazard (Madsen et al. 2008; Glimsdal et al. 2013).

In the present case, $b > 10d$ for the SMFs sited in Areas 1–4 and the waves are not expected to be significantly dispersive initially. However, as tsunamis propagate in the far field over long distances across the Atlantic Ocean, dispersive effects may potentially become significant. On the nearshore side, the N-wave-shaped waves simulated in grid G0 at 20 m depth (Fig. 1.9) may lead to the occurrence of undular bores, given a sufficiently fine grid resolution.

In the following, the magnitude of dispersive effects is assessed during tsunami generation, and for both the far- and near-field propagations of the deforming slide sited in Area 1 (Table 1.1). This will be done by comparing results of additional simulations, with regular dispersive effects or these being turned off in FUNWAVE, in finer nearshore nested grids (NJ30, OC30, NJ7.5; Fig. 1.2) and in one coarser farfield grid (ATL; Table 1.2). Without dispersive terms, FUNWAVE becomes a standard Nonlinear Shallow Water Wave equations solver.

1.5.1 SMF Tsunami Generation

The importance of dispersive effects on tsunami generation is first assessed with NHWAVE using the same grid and parameters as discussed above. Fig-

Figure 1.13: Dispersive effects in simulations with NHWAVE of tsunami generation for the deforming Currituck SMF proxy sited in Area 1 (Fig. 1.1), propagating in the corresponding 500 m NHWAVE grid (Fig. 1.2a): (a) relative difference (color scale in %) of maximum surface elevations in non-dispersive versus dispersive computations at $t = 1200$ s; black lines mark the SMF initial elliptical footprint as well as traces of cross- (n) and along- (f) slope transects. (b), (c) Instantaneous surface elevation along cross- and along-slope transects marked in a, respectively, at $t =$ (dashed) 300, (chained) 600, and (solid) 800 s, for: (blue) dispersive; and (red) non-dispersive simulations

Figure 1.13a shows the relative difference (in percent) computed between the non-dispersive and dispersive results at $t = 1200$ s (the time when these waves are passed onto FUNWAVE), and Fig. 1.13b, c shows surface elevations computed along cross- and along slope transects marked in (a), at $t = 300, 600$ and 800 s, in

the non-dispersive and dispersive results.

In the cross-slope transect of Fig. 1.13b, surface elevations are almost identical when turning off dispersion for onshore propagating waves, but differences start growing with time/distance for offshore propagating waves. This is consistent with results that could be expected based on long-wave theory. At $t = 300$ or 600 s, the dominant wavelength in the offshore propagating tsunami is $\lambda_0 \simeq 47.5$ or 60 km (the latter being $\simeq 2b$), in an average depth $d = 1.7$ or 1.9 km, respectively (Fig. 1.3a); hence, $\mu = 0.22$ or $0.2 < \pi/10$, which is in the long-wave regime. For the onshore propagating waves at $t = 300$ s, $\lambda_0 \simeq 24.5$ km in depth $d = 0.7$ km, so $\mu = 0.18$, which is also in the long-wave regime. Hence, this is indeed a very large SMF generating a long tsunami, for which dispersive effects are not significant. Consistent with these observations, in the along-slope transects of Fig. 1.13c, differences between dispersive and non-dispersive results are negligible at all times in the first 30 km on either side of the SMF initial location, but start slightly growing with time/distance, with larger differences occurring on the eastern side, maybe due to the effect of the Hudson River canyon.

According to the dispersion time criterion of Glimsdal et al. (2013), dispersive effects start being significant for $\tau \geq 0.1$, and a propagation distance $L \geq \tau \lambda_0^3 / (6h^2)$. In the near field, based on the initial depth and wavelength of the slide in Area 1, this yields $L = 500$ km, but even larger values as waves are propagating over the shallower shelf; for waves propagating offshore, this distance becomes $L > 618$ km. Consistent with these L values, Fig. 1.13a shows that, at $t = 1200$ s, differences between the nondispersive and dispersive results only become significant at a large distance from the source. The annular-shaped area with large relative differences ($\pm 100\%$) in instantaneous elevations is a result more of increasing phase shifts between the two wave trains than from large absolute

differences in surface elevation.

1.5.2 Far-Field Propagation in the Atlantic

Figure 1.14: Dispersive effects in FUNWAVE simulations of far-field tsunami propagation and hazard, for deforming Currituck SMF proxy sited in Area 1 (Fig. 1.1), propagating in 1 arc-min Atlantic grid ATL (Table 1.2): (a) boundary of ATL grid with locations of wave gauge stations 1–9 (Table 1.3) where time series of surface elevation are compared (Fig. 1.15); (b) relative difference (color scale in %) of maximum surface elevations in non-dispersive versus dispersive computations (note, areas with dispersive elevations less than 0.05 m have been avoided for clarity); (c), (d) envelope of maximum surface elevation computed without/with dispersion

Figure 14 compares the non-dispersive and dispersive results for the tsunami far-field propagation. In both cases, FUNWAVE is initialized with the same NHWAVE results, computed at $t = 800$ s with dispersion for the deforming slide in Area 1 (Fig. 1.4b). Results discussed in the previous section indicate that, at this time, differences with a non-dispersive generation are very small (Fig. 1.13). Tsunami propagation is simulated in the 1 arc-min resolution (about 1800 m)

spherical coordinate ATL grid defined over the Northern Atlantic Ocean (Fig. 1.14a; Table 1.2). [A video file, with side-by-side animations of simulated surface elevations for the deforming Currituck SMF proxy sited in Area 1 propagating in grid ATL, modeled without and with dispersion, is provided as online supplementary material: EtaSMFA1-Dispfar.mp4 (color scale in m).]

Figure 14c, d shows envelopes of maximum surface elevations computed without and with dispersive effects. These have fairly identical patterns and, as is typical in transoceanic tsunami propagation (e.g., Kirby et al. 2013; Tehranirad et al. 2015), both envelopes feature narrow stripes of increased surface elevations due to wave-guiding effects by seafloor topographic features. These cause maximum surface elevations to reach 1–2 m in the far field at some locations, such as the north shores of most of the Caribbean islands, the Dominican Republic, Puerto Rico, and the Leeward Islands. One prominent and most distant location with increased impact is a shore-parallel area off of Western Sahara (Morocco; Lat. 23°), south of the Canary Islands, where maximum elevations reach 2 m over 20–30 m-deep shallow water shoals. Figure 14b shows the relative difference in percent computed between these envelopes (note that areas with dispersive maximum elevations less than 0.05 m have been avoided in this figure, for clarity). As tsunami waves propagate away from the source, positive relative changes in elevation first occur, which gradually grow with radial distance, to reach about 50% in all directions. Then, as stripes of increased elevations start appearing in the envelopes, relative changes in elevation become heterogeneous. In far-field areas spanning the S-to-SE directions, where waves propagate towards South America and South Africa, maximum elevations without dispersion increase by up to 80% within the stripes and do not change or slightly decrease in between. In far-field areas spanning the SE-to-NE directions, however, this pattern reverses itself with

relative amplitude differences mostly decreasing by 20–40% along the upper West coast of Africa, and the coasts of Portugal and Spain, implying that dispersion increases far-field tsunami elevations in these areas.

To better assess dispersive effects on tsunami wave trains, Fig. 1.15 shows time series of surface elevations computed without and with dispersion at deep water wave gauge stations 1–9 specified in the ATL grid, the first 5 being collocated with existing DART buoys (deep water tsunami sensors operated by NOAA; Table 1.3; Fig. 1.14a). In all cases, differences between these wave trains and, hence, dispersive effects, increase with the propagation distance from the SMF source. Dispersion causes a gradual reduction in leading wave elevation, with a corresponding increase in the second wave elevation, which becomes the largest wave in the train, followed by an increasingly long dispersive wave tail. [The appearance of an increasingly long tail of leading waves in the dispersive simulations is quite apparent in the provided online video.] Note that a shorter tail of smaller waves also appears in the non-dispersive simulations due numerical dispersion in the model. While at the nearest stations (e.g., 3, 4, and 5), the maximum elevations of non-dispersive waves are larger than those of dispersive waves, at the more distant stations (e.g., 1, 6, 7, and 8), these are either about equal or the trend reverses itself, with dispersive waves being larger (e.g., 8). Finally, as noted in earlier work (Kirby et al. 2013; their Fig. 1.14), at all stations, there is a time lag between the leading nondispersive wave in the trains, whose phase speed is faster, and the leading dispersive wave, which becomes increasingly large at the most distant gauges (e.g., 6, 7.5, and 6 min at stations 6, 8, and 9, respectively). Such time lags further confirm the significance of frequency dispersion effects during far-field tsunami propagation.

Figure 1.15: Case of Fig. 1.14. Time series of surface elevation computed in grid ATL at deep water wave gauge stations 1–9 (Table 1.3; Fig. 1.14a) in: (red) non-dispersive; and (blue) dispersive simulations

1.5.3 Near-Field Propagation Directly Onshore of the Source

Dispersive effects are first assessed on the onshore wave propagation by simulating tsunami propagation without and with dispersion in the 500 m G0 and 120 m resolution NJ grids, which are directly onshore of Area 1 (Table 1.2; Fig. 1.2b). Then, to more accurately resolve the shorter waves that dispersive effects might generate and evaluate the effect of grid resolution on tsunami runup, tsunami propagation to shore is simulated without and with dispersion in two additional nested grid levels: a 30 m resolution grid NJ30 and a 7.5 m resolution grid NJ7.5, defined within grid NJ and centered on Seaside Heights (Table 1.2; Fig. 1.16a). [A video file, with side-by-side animations of simulated surface elevations for the deforming Currituck SMF proxy sited in Area 1 propagating in grids G0, NJ, NJ30 and NJ7.5, modeled without and with dispersion, is provided as online supplementary material: EtaSMF-A1-Dispnear.mp4 (color scale in m).]

Figures 16c, d and 17b, c show envelopes of maximum surface elevations computed in grids NJ30 and NJ7.5, without and with dispersion, respectively, and Figs. 16b and 17a show their relative difference in percent, respectively. Results in Fig. 1.16c, d in the 30 m grid NJ30 show fairly identical patterns for both envelopes, with moderate maximum elevation changes when turning off dispersive effects: (1) offshore, a decrease by up to 10% in a nearly shore parallel area, in 20–25 m depth and (2) nearshore, an increase (over and beyond the barrier island) of 2–5%, likely due to shallow water steepening effects (Fig. 1.16b). In the finer resolution, 7.5 m grid NJ7.5, however, Fig. 1.17a–c, show larger relative differences and more different patterns for both envelopes of surface elevations when turning off dispersive effects: (1) offshore, a decrease by up to 40% in a nearly shore-parallel area, in depth greater than 10–15 m, and (2) nearshore, a slight increase (over and beyond the barrier island) of 2–5%, likely due to shallow water steepening effects (Fig.

Figure 1.16: Dispersive effects in FUNWAVE simulations of nearshore tsunami propagation and hazard, for deforming SMF sited in Area 1 (Fig. 1.1) propagating in 30 and 7.5 m resolution nested grids NJ30 and NJ7.5 (Table 1.2): (a) boundaries of NJ30 and NJ7.5 grids sited within 120 m NJ grid, with location of wave gauge stations 10–13 (Table 1.3); (b) relative difference (color scale in %) of maximum surface elevations in NJ30 nondispersive versus dispersive computations; (c), (d) envelope of maximum surface elevation computed without/with dispersion in NJ30

1.17a).

Figure 1.17d shows maximum surface elevations computed in grid NJ7.5 at the 5 m isobath with and without dispersion, compared to those computed in the 30 and 120 m resolution grids NJ30 and NJ (the latter results being identical to those of Fig. 1.12a, b). Consistent with the above observations, there is a small increase in maximum elevation in the 7.5 m grid NJ7.5 when turning off dispersion;

Figure 1.17: Case of Fig. 1.16: (a) relative difference (color scale in %) of maximum surface elevations in NJ7.5 non-dispersive versus dispersive computations; (b), (c) envelope of maximum surface elevation computed without/with dispersion in NJ7.5; (d) maximum surface elevation along the 5 m isobath computed in: (solid) dispersive versus (dashed) non-dispersive simulations in: (blue) 120 m NJ; (red) 30 m NJ30; and (green) 7.5 m NJ7.5, grids (as a function of the distance along the contour, from S to N)

however, results are almost identical with or without dispersion in the 30 and 120 m resolution grids, in which dispersive effects are underresolved. The figure also

clearly illustrates the important effect of grid resolution on the maximum computed nearshore surface elevations, showing a large increase in elevation when resolution is increased from 120 to 30 m, and a smaller additional increase from 30 to 7.5 m. The latter confirms that tsunami inundation maps should be computed using at least a 30 m resolution grid, but preferably a 10 m resolution [as done by the authors along the USEC as part of NTHMP; see ECMAP (2017)].

These observations are further confirmed and detailed in Fig. 1.18, which shows time series of surface elevations computed without and with dispersion in the 30 m resolution NJ30 grid, at wave gauge stations 10–13, all located in 15 m depth (Fig. 1.16a; Table 1.3). For comparison and validation, time series computed at the same locations in the 120 m resolution NJ grid (with dispersion) are also shown. Overall, the latter agree well with the higher resolution results, except that the steeper and higher parts of wave crests are not as sharply captured as in the finer resolution grid. In the NJ30 grid results, in all cases, there are no noticeable differences between dispersive and non-dispersive results at the front of the wave trains (up to $t \simeq 2.6\text{--}2.7$ h), where waves are higher and steeper; these are waves that have only propagated a short distance from the onshore part of the SMF tsunami source. Small differences between dispersive and non-dispersive results start appearing at all stations later in the wave train ($t > 2.6\text{--}2.7$ h). When dispersion is turned off, waves' crests become slightly higher and steeper, and trough slightly deeper. The appearance of these dispersive effects likely results from the longer propagation distance of these waves, from their generation area.

Figure 1.19 shows time series of surface elevations computed at stations 11 and 12 in grids NJ30 (dispersive case) and NJ7.5 (dispersive and nondispersive cases). While there is a good overall agreement between simulations in the 30 and 7.5 m grids, larger discrepancies can be seen around the steeper wave crests and

Figure 1.18: Case of Fig. 1.16. Time series of surface elevation computed in: (solid) 30 m NJ30; (dashed) 120 m NJ, grid at shallow water wave gauge stations 10–13 (Fig. 1.16a; Table 1.3), for: (red) non-dispersive; and (blue) dispersive simulations

Figure 1.19: Case of Fig. 1.16. Time series of surface elevation computed in: (blue) 30 m NJ30; and (green) 7.5 m NJ7.5, grids at shallow water wave gauge stations 11–12 (Fig. 1.16a; Table 1.3), for: (dashed) non-dispersive; and (solid) dispersive simulations

some troughs. Specifically, in the finer grid, well-developed undular bores made of shorter oscillatory waves of $T \simeq 20$ s period appear at both gauges near $t = 2.2$ h. These waves are much shorter than the 3–5 min shortest period waves in the tsunami wave train. By contrast, in the 30 m resolution grid, undular bores are absent, which confirms the need for a high-resolution nearshore grid to properly model them. [Undular bores are quite apparent in the provided online video.] Undular bores cause surface elevations to be much larger off the coast in the dispersive simulations, which is consistent with the relative differences shown in Fig. 1.17a, for $h > 10$ m or so. In shallower water, however, perhaps, because the undular bore waves break and dissipate their energy nearshore, the maximum surface elevations in the dispersive and non-dispersive simulations become nearly identical, consistent with Fig. 1.17d.

While surface elevations near the coast and at the shoreline do not significantly differ in the nondispersive and dispersive results of grid NJ7.5 (Figs. 1.16, 1.17, 1.18, 1.19), Fig. 1.20 shows that this is not the case for the horizontal velocity module $U = |\mathbf{U}|$ (m/s) and impulse force $F_I = \rho_w(h + \eta)U^2$ (N/m; with $\rho_w = 1025$ kg/m³, the average sea water density, and η the instantaneous surface elevation). Figure 1.20a, b shows the maximum envelopes of these parameters computed with dispersion in grid NJ7.5 and Fig. 1.20c, d shows the relative differences (%) of non-dispersive versus dispersive results. Consistent with Fig. 1.17a, large decreases in both parameters occur offshore of the 5 m isobath. Onshore of this depth, the pattern becomes more complex but, unlike surface elevations, near the most impacted stretch of coastline around Seaside Heights, turning off dispersion decreases both U and F_I at many locations (see zoom-ins of Fig. 1.20e, f). In particular, near the shoreline and in the overland flow slightly onshore of it, maximum velocities and corresponding impulse forces are underpredicted by up to 20 and 40%, respectively.

In other areas, turning off dispersion still leads to overpredicting these parameters. It should be pointed out that a significant underprediction of the maximum F_I and U in build-up areas highly impacted by the tsunami-induced overland flow would have important implications for coastal hazard assessment if these results were used to design coastal structures. [A video file, with side-by-side animations of simulated impulse forces for the deforming Currituck SMF proxy sited in Area 1 propagating in grids NJ30 and NJ7.5, modeled without and with dispersion, is provided as online supplementary material: IFSMF-A1-Dispnear.mp4 (color scale in N/m).]

Towards Coastlines Away from the Source

As shown in Fig. 1.8b–g, due to wave-guiding effects by the shelf break and slope, tsunami waves generated by the deforming SMF in Area 1 “open-up” and end up propagating onshore along the entire length of the shelf in grid G0. In particular, some of these waves propagate onshore towards Ocean City, MD, which is located about 325 km west of the SMF source (Fig. 1.1). Since such propagation takes place over longer distances than for waves propagating directly onshore in grid NJ30, albeit over the shallow shelf, we evaluate whether this may cause increased dispersive effects. To do so, nearshore wave propagation is simulated in a 30 m resolution grid OC30 centered around Ocean City, nested within the 120 m grid OC (Table 1.2; Figs. 1.2b, 1.21a). As before, Fig. 1.21b–d shows envelopes of maximum surface elevations and their relative difference computed in grid OC30 without and with dispersion. The SMF tsunami impact on Ocean City is almost as severe as in Seaside Heights, with a maximum inundation reaching over 6.5 m and overtopping the barrier beach. Both envelopes, however, are fairly similar with almost no change in surface elevation nearshore and an up to 8% decrease in elevation offshore when turning off dispersion in the model. This is confirmed by

Figure 1.20: Case of Fig. 1.16 in 7.5 m grid NJ7.5: (a), (b) maximum velocity module (color scale in m/s)/impulse force (color scale in N/m) computed with dispersion; (c), (d) relative difference of non-dispersive versus dispersive computations in (a), (b) (color scale in %); (e), (f) zoom-in of (c), (d) in maximum impacted area around Seaside Heights. Black bullets mark locations of wave gauge stations 11 (bottom) and 12 (top)

Figure 1.21: Dispersive effects in FUNWAVE simulations of nearshore tsunami propagation and hazard, for deforming SMF sited in Area 1 (Fig. 1.1) propagating in nested 30 m grid OC30 (Table 1.2): (a) boundary of OC30 grid sited within 120 m OC grid, with location of wave gauge stations 14–17 (Table 1.3); (b) relative difference (color scale in %) of maximum surface elevations in non-dispersive versus dispersive computations; (c), (d) envelope of maximum surface elevation computed without/with dispersion

time series of surface elevations at 4 gauges stations (14–17) located in 10 m depth shown in Fig. 1.22, for a 1 h time period. At grid OCE30's resolution, similar to results in NJ30, only small changes in instantaneous surface elevation occur when turning off dispersion, and these all correspond to a slight increase and a steepening of the wave crest fronts. It is equally clear here that the first wave generated by this large SMF is quite long, with a 30 min period, and becomes N-wave-shaped

in shallow water. For such a long wave propagating from its source in shallow water, one would not have expected significant dispersive effects, as confirmed by the results presented here. However, as for simulations around Seaside Heights and in view of the waves' steep fronts, it is likely that further increasing the grid resolution would cause the appearance of undular bores in numerical results. Such additional simulations are not performed here for the sake of brevity.

Figure 1.22: Case of Fig. 1.21. Time series of surface elevation computed in 30 m OC30 grid at shallow water wave gauge stations 14–17 (Fig. 1.21a, Table 1.3), for: (red) non-dispersive; and (blue) dispersive simulations

1.6 Discussion

In their probabilistic Monte Carlo Simulations (MCS) of the landslide tsunami hazard along the USEC, Grilli et al. (2009) concluded that the 500 year tsunami runup along the upper USEC was largest from Virginia to Cape Cod (about 6–7 m maximum). By performing additional MCS and geotechnical analyses, Krauss (2011) and Eggeling (2012) identified 4 areas (Fig. 1.1), where large SMFs were both possible due to sufficient sediment thickness and most probable due to low safety factors in slope stability analyses. In light of these conclusions and considering Probable Maximum Tsunamis (PMTs), Grilli et al. (2015) simulated extreme SMFs sited in these 4 areas, which had the lumped characteristics of the historical Currituck slide complex (i.e., “Currituck SMF proxies”; 30 km long by 20 m wide by 0.75 km maximum thickness; with a 158 km³ volume). To maximize coastal impact, these SMFs were all assumed to fail as rigid slumps. By combining the coastal inundation caused by these SMF tsunamis with that of other PMTs from extreme sources identified in the Atlantic Ocean, a first generation of tsunami inundation maps was developed for the most exposed areas of the USEC to tsunami hazard, under the auspice of the National Tsunami Hazard Mitigation Program (NTHMP; see ECMAP 2017). Using a newer model, in which deforming slides are modeled as a dense viscous fluid layer (Kirby et al. 2016), Grilli et al. (2017b) had simulated deforming SMFs of various rheology sited in Area 1 and briefly assessed how the nearshore tsunami hazard was affected. In the present work, following a similar methodology, we more thoroughly assessed how the earlier conservative assumption that all SMFs failed as rigid slumps affected coastal tsunami hazard along the upper USEC, in terms maximum inundation, current velocity, and impulse forces. In addition, effects on these results of including bottom friction and frequency dispersion in the simulations were evaluated.

As in earlier work, both slumps and slides were assumed to all have the Currituck SMF proxy characteristics, with a quasi-Gaussian initial geometry and bulk density $\rho_s = 1900 \text{ kg/m}^3$. Slides were assumed to have a fairly large viscosity, in the upper range of debris flows ($\nu_s = 0.5 \text{ m/s}^2$), and a substrate-to-slide Manning friction coefficient $n = 0.1$. To more accurately compare nearshore results and evaluate dispersive effects in tsunami wave trains, simulations were performed at much higher resolution than before [in nested grids down to a 7.5 m resolution versus a 120 m resolution in Grilli et al. (2017b)]. Tsunami generation was simulated using the 3D non-hydrostatic model NHWAVE, either with a single layer for slumps (Ma et al. 2012) or with two layers for deforming slides (Kirby et al. 2016). For rigid slumps, both time varying geometry and law of motion were specified as bottom boundary conditions (Grilli and Watts 2005; Watts et al. 2005; Grilli et al. 2015), and for deforming slides, SMF motion and deformation were directly modeled as a depth-integrated bottom layer of dense Newtonian fluid, fully coupled to the overlying fluid motion modeled with the standard σ -layer NHWAVE. Once SMFs were no-longer tsunamigenic (i.e., the slumps had stopped moving or the slides were deep and thin enough), tsunami propagation simulations were continued using the 2D fully nonlinear and dispersive longwave model FUNWAVE. Simulations of onshore tsunami propagation and inundation along the entire stretch of USEC considered here (from Virginia to Cape Cod) were performed in nested 500 and 120 m resolution grids, with a standard one-way coupling methodology. To assess dispersive effects, selected simulations were similarly performed nearshore in 30 and 7.5 m resolution nested grids (off of Seaside Heights, NJ and Ocean City, MD), and offshore in a 1 arc-min spherical coordinate grid covering the Northern Atlantic Ocean.

Results of tsunami generation simulations showed that deforming slides, while

having a slightly larger initial acceleration, generated smaller onshore propagating tsunamis than rigid slumps. Because deforming slide parameters were selected to ensure that the center of mass motion of both types of SMFs was quite similar, at least during the first half of their motion (see Fig. 1.5), these observed differences in onshore tsunami waves should essentially be associated with slide deformation. This was also the conclusion of Grilli and Watts (2005), who modeled deforming slides by superimposing a rate of deformation on top of the center of mass law of motion of rigid (block) slides. As shown in Figs. 1.4 and 1.6, when deforming slides move down the slope, their material spreads and thins out, which gradually makes them less tsunamigenic; by contrast, rigid slumps keep their specified shape during their pendulum-like motion. In addition, since they flow down the slope following the terrain, deforming slides cause more asymmetric tsunami waves alongshore, with respect to their initial direction of motion, than rigid slumps (Fig. 1.6). The offshore propagating tsunami waves caused by slides were usually as large as or even larger in some cases than for slumps (Fig. 1.4).

Effects of SMF deformation on the coastal tsunami impact were evaluated by comparing the combined maximum envelope of surface elevations caused nearshore and along the shore, and over the 5m isobath, by tsunamis generated by the 4 rigid slumps or deforming slides, in seven 120 m resolution grids covering the coast (Fig. 1.12). Although a 120 m resolution is not sufficient to accurately compute coastal inundation, it is adequate to perform this comparison. Consistent with earlier work (Tehrani et al. 2015, 2017; Grilli et al. 2017b), the bathymetry of the wide shelf was found to have strong, first-order, control on the magnitude of tsunami coastal inundation, as it induces wave focusing and defocusing effects. In addition, tsunami propagation and refraction over the shelf and shelf break, both north and south of each source area, causes nontrivial variations in surface elevation and

coastal inundation. This implies that a given SMF can generate tsunami waves causing a significant coastal inundation far alongshore from their source area (see, e.g., in Fig. 1.21, the large inundation caused at Ocean City, MD, by the SMF sited in Area 1 in the Hudson River Canyon).

Overall, as expected from earlier work (Grilli and Watts 2005; Grilli et al. 2017b), tsunamis caused by rigid slumps were found to be worst case scenarios, causing the largest maximum inundation at all sites (with a maximum of about 11.5 m around Montauk, NY), and up to 50% more inundation directly onshore than for the slides considered here, which only had a moderate level of deformation (Fig. 1.12). However, the slightly larger offshore propagating tsunamis generated by deforming slides could affect far-distant nearshore areas, through shelf refraction, thus increasing the relative coastal impact of such tsunamis and reducing the difference in maximum inundation with that of tsunamis generated by rigid slumps. For rigid slumps, the maximum computed inundation of 8–11.5 m at the 5 m isobath is larger than the 500 year runup of 6–7 m estimated by Grilli et al. (2009), also based on rigid slumps (or slides) assumptions. This could be expected, since the Currituck SMF proxies considered here were aimed at causing PMTs for the upper USEC, with potentially thousands of years return periods (the estimated age of the Currituck slide complex is about 16Ka). By contrast, the maximum inundation of 8.5 m computed here for deforming slides (around Atlantic City) is more consistent with the 500 year runup estimated in earlier work based on MCS. This could indicate that the return period of or the maximum runup caused by the largest events were underestimated in the MCS of Grilli et al. (2009). Regarding minimum elevations at the coast, which affect power plant intakes, tsunamis from both types of SMFs were shown to be able to cause water withdrawal to the 5 m isobath or slightly deeper.

The effects of bottom friction on tsunami coastal inundation were assessed by performing simulations for the Area 1 deforming slide, in the 120 m NJ grid, using two different Manning coefficients, one 50% larger than the other (from $n = 0.025$ – 0.375). Using the increased friction in simulations, i.e., considering a gravel/boulder rather than a sandy substrate, led to a reduction of the largest maximum tsunami inundation at the coast, in some cases, by up to 15%. Smaller inundation levels were less noticeably affected (Fig. 1.12a).

Dispersive effects were assessed during tsunami generation, for the deforming Currituck SMF proxy sited in Area 1, by comparing NHWAVE simulations with and without non-hydrostatic pressure terms in the model. Similarly, dispersive effects were assessed on both near- and far-field tsunami propagations by comparing FUNWAVE simulations with and without dispersion. For the very large SMFs considered here, which generate tsunamis having a very long initial wavelength compared to the local depth ($\mu < \pi/10$), non-hydrostatic (dispersive) effects in NHWAVE were found to be negligible for tsunami generation (Fig. 1.13). During tsunami propagation (Figs. 1.14, 1.15, 1.16, 1.17, 1.18, 1.19, 1.20, 1.21, 1.22), overall, turning off dispersion in FUNWAVE only led to moderate changes in maximum inundation at the coast (2–5% changes), but to more significant effects in the far field (- 40 to 80% changes). In addition, in far-field dispersive wave trains, the largest wave, which is initially the leading crest, gradually becomes the second wave in the train and is associated with a significant dispersive wave tail. In non-dispersive wave trains the largest wave is always the leading crest and the number of significant waves in the tail is more limited. Finally, there is an increasing time lag with propagation distance of the dispersive waves, as compared to the non-dispersive case.

The nearshore simulations in finer grids clearly confirmed the need to perform

tsunami inundation mapping using at least a 30 m resolution, or preferably a 10 m resolution grid as recommended by NTHMP, to have sufficiently converged maximum inundation results (see Fig. 1.17; note a 10 m resolution grid was not used here, but the 7.5 m grid is deemed close enough to it to draw this conclusion). However, equally important, the 7.5 m resolution results with dispersion showed the appearance nearshore of undular bores made of shorter (~ 20 s period) waves, which significantly affect the dynamics of incoming tsunami waves; such shorter waves are absent in the coarser resolution grids, hence further justifying the need for using high-enough nearshore resolution grids (here 7.5 m). While these bores likely break near or at the coast and thus have little effect on maximum inundation, they significantly increase maximum flow velocity and impulse forces, particularly in the overland flow, which both increase the tsunami hazard on civil infrastructures. Without dispersive effects, the undular bores do not occur in model results at any resolution, and maximum flow velocities and corresponding impulse forces can be underestimated at the coast, by up to 20 and 40%, respectively (note, this conclusion is only based on limited results around the highly impacted area of Seaside Heights, NJ; see Fig. 1.20).

Because onshore surface elevations with or without dispersion are almost identical, the large underprediction of impulse forces mostly results from the underprediction of velocities. When undular bores are present, flow velocities are larger near the longwave crests, including when the steep individual bores end up breaking in very shallow water [breaking undular bores riding longer tsunami waves can be seen, for instance, in pictures of the Indian Ocean tsunami approaching the coast reported by Madsen et al. (2008); see also the various discussion regarding undular bores in this paper]. Hence, incident undular bores intensify the dynamics and kinematics of the overall flow. This is an important finding for tsunami

hazard assessment, particularly in view of the recently released structural design code ASCE 7–16, which recommends using impulse forces and maximum speed of transported debris for designing tsunami resistant structures (Chock 2016).

These results imply that while using a NSW model may be adequate to predict maximum coastal inundations, this may not capture the required physics to model flow velocities and corresponding impulse forces. This conclusion is consistent with those of a recent tsunami velocity benchmarking workshop (Lynett et al. 2017).

1.7 Conclusion

Unless considering critical coastal facilities (such as a nuclear power plant), which require considering the most extreme PMTs, it appears adequate in coastal tsunami hazard assessment to simulate more realistic moderately deforming SMFs, whose center of mass may follow a similar kinematics as that of rigid slumps, but whose deformation leads to a reduced coastal inundation. Using a moderate level of SMF deformation is also consistent with paleo-SMF observations made on the upper East coast continental shelf and margin (Booth et al. 2003; ten Brink et al. 2014). Considering its marked effect on maximum inundation, values of the bottom friction coefficient should be carefully selected as a function of the local substrate, particularly nearshore, and onshore land cover. In the absence of detailed information, assuming that a coarse sand substrate should be conservative. The need to use a dispersive model for landslide tsunami generation depends on the relative size of the SMF and waves it is generating, as compared to the local depth. For the very large SMFs considered here, a non-dispersive model would have been adequate. For tsunami propagation, using a dispersive model such as a long-wave Boussinesq model is important to accurately simulate both far-field and near-field tsunami propagation, together with sufficiently high-resolution model grids. In the near field, even if the initial SMF tsunami is made of non-dispersive long waves (as

in the present case), this is necessary to accurately model the undular bores made of shorter dispersive waves that may occur in shallow water near the steep fronts of incoming long waves. Such bores cause rapid changes in surface elevation and intensify wave dynamics in a way that may significantly increase maximum forces on structures. However, using a NSW model is adequate if maximum onshore inundation is the only parameter of interest in the simulations. While in future tsunami inundation mapping work, it would be desirable to perform a PTHA, as is currently being done in other regions of the US (e.g., West Coast), the current lack of sufficient information on return periods of tsunamigenic SMF sources precludes considering this approach for the USEC. However, should a sufficient number of past tsunamigenic SMF sources be dated, this could provide relevant hazard information, making it possible to perform a PTHA.

Acknowledgements. This work was supported by the National Tsunami Hazard Mitigation Program (NTHMP), NOAA, through Grants NA-15-NWS4670029 and NA-16NWS4670034 to the University of Delaware (with subaward to the University of Rhode Island). Additional support at the University of Rhode Island and the University of Delaware came from Grants CMMI-15-35568 and CMMI-15-37568 from the Engineering for Natural Hazards Program, National Science Foundation, respectively. Numerical simulations reported in this work used HPC resources, as part of the Extreme Science and Engineering Discovery Environment (XSEDE) (project BCS170006), which is supported by the National Science Foundation (NSF) Grant number ACI-1548562. FUNWAVE-TVD is open source software, available at <http://github.com/fengyanshi/FUNWAVE-TVD/>. NHWAVE is open source software, available at <http://github.com/jimkirby/nhwave/>. Finally, the authors acknowledge anonymous reviewers for their thorough and constructive reviews of this work.

Appendix: SMF Geometry and Slump Law of Motion

For rigid slumps, kinematics is specified based on the analytical laws developed by Grilli and Watts (1999, 2005); Watts et al. (2005). Additionally, as in Enet and Grilli (2007), the SMF geometry is idealized as having a “Quasi-Gaussian” shape (below seafloor) of elevation $\zeta(x, y)$ whose steepness is controlled by a shape parameter ϵ (here $\epsilon = 0.717$), and elliptical footprint of downslope length b , width w , and maximum thickness T defined as (Fig. 1.23)

$$\zeta(x, y) = \frac{T}{1 - \epsilon} \max\{0, \text{sech}(k_b \xi) \text{sech}(k_w \chi) - \epsilon\} \quad (1.1)$$

where (ξ, χ) are the local downslope and spanwise horizontal coordinates, rotated in the direction of SMF motion θ , and $k_b = 2C/b$, $k_w = 2C/w$, with $C = \text{acosh}(1/\epsilon)$. With this geometry and parameters, the SMF volume is given by

$$V_s = bwT \frac{I_2}{C^2} \left(\frac{\frac{I_1}{I_2} - \epsilon}{1 - \epsilon} \right) \quad \text{with } I_{1;2} = \int_0^C f(\mu) d\mu; g(\mu) d\mu \quad (1.2)$$

and

$$f(\mu) = \text{sech}(\mu) \arctan(\sinh g(\mu)) , \quad g(\mu) = \text{acosh} \left(\frac{\text{sech}(\mu)}{\epsilon} \right) \quad (1.3)$$

[Note that Eqs. 2 and 3 have been corrected and are different from those reported in earlier papers (e.g., Enet and Grilli 2007; Grilli et al. 2015), which resulted from a mistake in the volume calculation.] For the specified ϵ , we find, $C = 0.8616$, $I_1 = 0.4804$, $I_2 = 0.5672$, and $V_s = 0.3508 bwT$.

Earlier modeling work (Locat et al. 2009) indicates that, during its tsunami-genic period of motion, the Currituck SMF achieved a relatively small maximum displacement (runout) $S_f < b$ in its main direction of motion down the slope, over an unknown time of motion t_f . The combination of rigid block SMF and small displacement parallel to the slope supports modeling the SMF kinematics as a

Figure 1.23: Geometric parameterization of a SMF initially centered at (x_0, y_0) moving in direction n , with an azimuth angle θ from North and center of mass motion $S(t)$ measured parallel to the mean local slope of angle α ; (x, y) denote the longitudinal and latitudinal horizontal directions, respectively

rigid slump or a deforming slide with moderate deformation achieving a similar runout over the same time. In either cases, one can assume a constant basal friction (i.e., slide-to-substrate friction) and negligible hydrodynamic drag (Grilli and Watts 2005). This type of rigid-body motion kinematics was investigated in earlier work (see above-listed references), leading for the slump to a pendulum-like center of mass motion $S(t)$ parallel to the local mean slope of angle α . Here, this simple law of motion for rigid slumps is used, which reads

$$S(t) = S_0(1 - \cos(\frac{t}{t_0})) \text{ for } 0 \leq t \leq t_f \quad (1.4)$$

with $S_0 = S_f/2$ and $t_0 = t_f/\pi$, and $S = S_f$ for $t > t_f$ (assuming SMF triggering occurs at $t = 0$).

At $t = 0$, the SMF elevation is specified below the current seafloor bathymetry $h_0(x, y)$. Given the SMF initial center of mass location (x_0, y_0) in global axes (x, y) (i.e., coordinates of the center of the elliptical footprint) and azimuth angle of SMF motion θ , the coordinate transformation to the local SMF slope parallel coordinate

system (ξ, χ) (Fig. 1.23) is defined as

$$\begin{aligned}\xi &= (x - x_0) \cos(\theta) - (y - y_0) \sin(\theta) - S(t) \cos(\alpha) \\ \chi &= (x - x_0) \sin(\theta) + (y - y_0) \cos(\theta)\end{aligned}\tag{1.5}$$

with $S(t)$ given by Eq. (4).

Then, assuming $\sin(\alpha) \simeq 0$ for small slopes, the instantaneous seafloor depth above the SMF is given by

$$h(x, y, t) = h_0(x, y) + \zeta\{\xi(x, y, t), \chi(x, y, t)\} - \zeta\{\xi(x, y, 0), \chi(x, y, 0)\}\tag{1.6}$$

with $\Delta h = h - h_0$. The seafloor motion described by Eq. (6) is similar to a translation parallel to the average slope of part of the seabed, over the actual bathymetry. The vertical seafloor velocity (used in NHWAVE as a bottom boundary condition) is computed as

$$\frac{dh}{dt}(x, y, t) = \frac{d\zeta}{dt}\{\xi(x, y, S(t), \chi(x, y, t))\}\tag{1.7}$$

which can easily be derived from Eqs. (1) to (6) as

$$\frac{dh}{dt}(x, y, t) = k_b \cos \alpha \left(\zeta + \frac{\epsilon T}{1 - \epsilon} \right) U \tanh(k_b \zeta)\tag{1.8}$$

with

$$U(t) = \frac{dS}{dt} = U_{max} \sin\left(\frac{t}{t_0}\right)$$

the slump velocity obtained from Eq. (4), with $U_{max} = S_0/t_0$ the maximum velocity. Similarly, the slump acceleration is found as

$$A(t) = \frac{d^2S}{dt^2} = A_0 \cos \frac{t}{t_0} \text{ with } A_0 = \frac{S_0}{t_0^2}\tag{1.9}$$

the initial acceleration.

For rigid slumps, hydrodynamic drag can be neglected due to low velocity and small amplitude of motion, and inertia includes both the SMF mass $M_s = \rho_s V_s$, with ρ_s denoting the sediment bulk density, and the specific density being defined as $\gamma = \rho_s / \rho_w$, with ρ_w the water density, and an added mass $\Delta M_s = C_M \rho_w V_s$, defined by way of an added mass coefficient C_M . Assuming a constant basal friction, a nearly circular rupture surface of radius R , and a small angular displacement $\Delta\Phi$, Grilli and Watts (2005) derived the characteristic distance and time of motion for rigid slumps as

$$S_0 = \frac{R\Delta\Phi}{2} \text{ and } t_0 = \sqrt{\frac{R}{g} \frac{\gamma + C_M}{\gamma - 1}} \text{ with } R \simeq \frac{b^2}{8T} \quad (1.10)$$

with g denoting the gravitational acceleration. Equation (8), proposed by Watts et al. (2005), is a semi-empirical relationship to estimate the radius of slump motion as a function of slump downslope length and maximum thickness.

References

- Abadie, S., Harris, J. C., Grilli, S. T., and Fabre, R. (2012). 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. *Journal of Geophysical Research*, 117, C05030. <https://doi.org/10.1029/2011JC007646>.
- Amante, C, and Eakins, B. W. (2009). ETOPO1 1 arc-minute global relief model: Procedures, data sources and analysis. NOAA Technical Memorandum NESDIS NGDC-24. National Geophysical Data Center, NOAA. <https://doi.org/10.7289/V5C8276M> (March 15, 2017).
- Booth, J. S., O’Leary, D. W., Popenoe, P., and Danforth, W. W. (1993). US Atlantic Continental Slope Landslides: Their distribution, general attributes, and implications. In: Schwab, W. C., Lee, H. J., and Twichell, D. C. (Eds.) *Submarine landslides: Selected studies in the US exclusive economic zone*. US Geological Survey Bulletin, 2002, pp. 14–22.
- ten Brink, U. S., Barkan, R., Andrews, B. D., and Chaytor, J. D. (2009b). Size distributions and failure initiation of submarine and subaerial landslides. *Earth and Planetary Science Letters*, 287, 31–42.
- ten Brink, U. S., Chaytor, J. D., Geist, E. L., Brothers, D. S., and Andrews, B. D. (2014). Assessment of tsunami hazard to the US Atlantic margin. *Marine Geology*, 353, 31–54.
- ten Brink, U. S., Lee, H. J., Geist, E. L., and Twichell, D. (2009a). Assessment of tsunami hazard to the US East Coast using relationships between submarine landslides and earthquakes. *Marine Geology*, 264, 65–73.

- 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. (2008). Evaluation of tsunami sources with the potential to impact the US Atlantic and Gulf coasts. USGS Administrative report to the US Nuclear Regulatory Commission, p. 300.
- Chaytor, J. D., ten Brink, U. S., Solow, A. R., and Andrews, B. D. (2009). Size distribution of submarine landslides along the US Atlantic margin. *Mar Geol*, 264(12), 16–27.
- Chock, G. Y. (2016). Design for tsunami loads and effects in the ASCE 7–16 standard. *Journal of Structural Engineering*, 142(11), 04016093. [https://doi.org/10.1061/\(ASCE\)ST.1943541X.0001565](https://doi.org/10.1061/(ASCE)ST.1943541X.0001565).
- ECMAP. (2017). NTHMP tsunami inundation maps for the US East Coast. <https://www.udel.edu/kirby/nthmp.html>.
- Eggeling, T. (2012). Analysis of earthquake triggered submarine landslides at four locations along the US east coast. Masters Thesis. University of Rhode Island.
- Enet, F., and Grilli, S. T. (2007). Experimental study of tsunami generation by three-dimensional rigid underwater landslides. *Journal of Waterway Port Coastal and Ocean Engineering*, 133(6), 442–454. [https://doi.org/10.1061/\(ASCE\)0733-950X\(2007\)](https://doi.org/10.1061/(ASCE)0733-950X(2007)).
- Fine, I. V., Rabinovich, A. B., Bornhold, B. D., Thomson, R., and Kulikov, E. A. (2005). The Grand Banks landslide-generated tsunami of November 18, 1929: Preliminary analysis and numerical modeling. *Marine Geology*, 215, 45–57.

- Geist, E. L., and Lynett, P. J. (2014). Source processes for the probabilistic assessment of tsunami hazards. *Oceanography*, 27(2), 86–93.
- Geist, E., Lynett, P., and Chaytor, J. (2009). Hydrodynamic modeling of tsunamis from the Currituck landslide. *Marine Geology*, 264, 41–52. <https://doi.org/10.1016/j.margeo.2008.09.005>.
- Glimsdal, S., Pedersen, G. K., Harbitz, C. B., and Løvholt, F. (2013). Dispersion of tsunamis: Does it really matter? *Natural Hazards and Earth System Sciences*, 13, 1507–1526. <https://doi.org/10.5194/nhess-13-1507-2013>.
- González, F. I., Geist, E. L., Jaffe, B., Kânoglu, U., Mofjeld, H., Synolakis, C. E., et al. (2009). Probabilistic tsunami hazard assessment at seaside, Oregon, for near- and far-field seismic sources. *Journal of Geophysical Research Oceans*, 114, C11.
- Grezio, A., Babeyko, A., Baptista, M. A., Behrens, J., Costa, A., Davies, G., et al. (2017). Probabilistic tsunami hazard analysis: Multiple sources and global applications. *Reviews of Geophysics*, 55(4), 1158–1198. <https://doi.org/10.1002/2017RG000579>.
- Grilli, S. T., and Watts, P. (1999). Modeling of waves generated by a moving submerged body. Applications to underwater landslides. *Engineering Analysis with Boundary Elements*, 23, 645–656.
- Grilli, S. T., and Watts, P. (2005). Tsunami generation by submarine mass failure. I: Modeling, experimental validation, and sensitivity analyses. *Journal of Waterway Port Coastal and Ocean Engineering*, 131(6), 283–297. [https://doi.org/10.1061/\(ASCE\)0733-950X\(2005\)](https://doi.org/10.1061/(ASCE)0733-950X(2005)).

- Grilli, S. T., Dubosq, S., Pophet, N., Pérignon, Y., Kirby, J. T., and Shi, F. (2010). 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*, 10, 2109–2125. <https://doi.org/10.5194/nhess-2109-2010>.
- Grilli, S. T., Grilli, A. R., Tehranirad, B., and Kirby, J. T. (2017a). Modeling tsunami sources and their propagation in the Atlantic Ocean for coastal tsunami hazard assessment and inundation mapping along the US East Coast. In *Proceedings of coastal structures and solutions to coastal disasters 2015: Tsunamis* (pp. 1–12). Boston: American Soc. Civil Eng, 2015.
- Grilli, S. T., Harris, J. C., Tajalibakhsh, T., Masterlark, T. L., Kyriakopoulos, C., Kirby, J. T., et al. (2013b). 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*, 170, 1333–1359. <https://doi.org/10.1007/s00024-012-0528-y>.
- Grilli, S. T., Harris, J. C., Tappin, D. R., Masterlark, T., Kirby, J. T., Shi, F., and Ma, G. (2013a). Numerical modeling of coastal tsunami dissipation and impact. In P. Lynett, and J. Mc Kee Smith (Eds.), *Proceedings of 33rd international coastal engineering conference (ICCE12, Santander, Spain, July, 2012)* (p. 12). World Scientific Publishing Co. Pte. Ltd.
- Grilli, S. T., O'Reilly, C., Harris, J. C., Tajalli Bakhsh, T., Tehranirad, B., Banihashemi, S., et al. (2015). Modeling of SMF tsunami hazard along the upper US East Coast: Detailed impact around Ocean City, MD. *Natural Hazards*, 76(2), 705–746. <https://doi.org/10.1007/s11069-014-1522-8>.

- Grilli, S. T., Shelby, M., Kimmoun, O., Dupont, G., Nicolsky, D., Ma, G., et al. (2017b). Modeling coastal tsunami hazard from submarine mass failures: Effect of slide rheology, experimental validation, and case studies off the US East Coast. *Natural Hazards*, 86(1), 353–391. <https://doi.org/10.1007/s11069-0162692-3>.
- Grilli, S. T., Taylor, O.-D. S., Baxter, C. D. P., and Marezki, S. (2009). Probabilistic approach for determining submarine landslide tsunami hazard along the upper East Coast of the United States. *Marine Geology*, 264(1–2), 74–97.
- Grilli, S. T., Vogelmann, S., and Watts, P. (2002). Development of a 3D Numerical Wave Tank for modeling tsunami generation by underwater landslides. *Engineering Analysis with Boundary Elements*, 26(4), 301–313.
- Grothe, P. R., Taylor, L. A., Eakins, B. W., Warnken, R. R., Carignan, K. S., Lim, E., Caldwell, R.J., and Friday, D. Z. (2010). Digital elevation model of Ocean City, Maryland: Procedures, data sources and analysis. NOAA Technical Memorandum NESDIS NGDC-37. National Geophysical Data Center, NOAA.
- Heidarzadeh, M., and Kijko, Andrzej. (2011). A probabilistic tsunami hazard assessment for the Makran subduction zone at the northwestern Indian Ocean. *Natural Hazard*, 56(3), 577–593.
- Hill, J. C., Brothers, D. S., Craig, B. K., ten Brink, Uri S., Chaytor, J. D., and Flores, C. H. (2017). Geologic controls on submarine slope failure along the central US Atlantic margin: Insights from the Currituck Slide Complex. *Marine Geology*, 385, 114–130. <https://doi.org/10.1016/j.margeo.2016.10.007>.
- Horspool, N., Pranantyo, I., Griffin, J., Latief, H., Natawidjaja, D. H., Kongko,

- W., et al. (2014). A probabilistic tsunami hazard assessment for Indonesia. *Natural Hazards and Earth System Sciences*, 14(11), 3105–3122.
- Jakeman, J. D., Nielsen, O. M., van Putten, K., Mleczko, R., Burbidge, D., & Horspool, N. (2010). Towards spatially distributed quantitative assessment of tsunami inundation models. *Ocean Dynamics*, 60(5), 1115–1138. <https://doi.org/10.1007/s10236-010-0312-4>.
- Kaiser, G., Scheele, L., Kortenhaus, A., Løvholt, F., Römer, H., & Leschka, S. (2011). The influence of land cover roughness on the results of high resolution tsunami inundation modeling. *Natural Hazards and Earth System Sciences*, 11, 2521–2540. <https://doi.org/10.5194/nhess-11-2521-2011>.
- Kirby, J. T., Shi, F., Nicolsky, D., & Misra, S. (2016). The 27 April 1975 Kitimat, British Columbia submarine landslide tsunami: A comparison of modeling approaches. *Landslides*, 13(6), 1421–1434. <https://doi.org/10.1007/s10346-016-0682-x>.
- Kirby, J. T., Shi, F., Tehranirad, B., Harris, J. C., & Grilli, S. T. (2013). Dispersive tsunami waves in the ocean: Model equations and sensitivity to dispersion and Coriolis effects. *Ocean Modelling*, 62, 39–55. <https://doi.org/10.1016/j.ocemod.2012.11.009>.
- Krauss, T. (2011). Probabilistic tsunami hazard assessment for the United States East Coast. Masters Thesis, University of Rhode Island. <http://chinacat.coastal.udel.edu/nthmp/krause-ms-uri11.pdf>.
- Løvholt, F., Bondevik, S., Laberg, J. S., Kim, J., & Boylan, N. (2017). Some giant submarine landslides do not produce large tsunamis. *Geophysical Research Letters*, 44(16), 8463–8472. <https://doi.org/10.1002/2017GL074062>.

- Løvholt, F., Pedersen, G., & Gisler, G. (2008). Oceanic propagation of a potential tsunami from the La Palma Island. *Journal of Geophysical Research*, 113, C09026.
- Løvholt, F., Pedersen, G., Harbitz, C. B., Glimsdal, S., & Kim, J. (2015). On the characteristics of landslide tsunamis. *Philosophical Transactions of the Royal Society A*, 373, 20140376. <https://doi.org/10.1098/rsta.2014.0376>.
- Løvholt, F., Schulten, I., Mosher, D., Harbitz, C., & Kraste, S. (2018). Modelling the 1929 Grand Banks slump and landslide tsunami. In D. G. Lintern, D. C. Mosher, L. G. Moscardelli, P. T. Bobrowsky, C. Campbell, J. D. Chaytor, J. J. Clague, A. Georgiopoulou, P. Lajeunesse, A. Normandeau, D. J. W. Piper, M. Scherwath, C. Stacey, & D. Turmel (Eds.), *Subaqueous mass movements* (p. 477). London: Geological Society London Special Publications. <https://doi.org/10.1144/SP477.28>.
- Locat, J., Lee, H., ten Brink, U. S., Twitchell, D., Geist, E., & Sansoucy, M. (2009). Geomorphology, stability and mobility of the Currituck slide. *Marine Geology*, 264, 28–40.
- Lynett, P., et al. (2017). Inter-model analysis of tsunami-induced coastal currents. *Ocean Modeling*, 114, 14–32. <https://doi.org/10.1016/j.ocemod.2017.04.003>.
- Ma, G., Shi, F., & Kirby, J. T. (2012). Shock-capturing non-hydrostatic model for fully dispersive surface wave processes. *Ocean Modelling*, 43–44, 22–35.
- Madsen, P. A., Fuhrman, D. R., & Schaffer, H. A. (2008). On the solitary wave paradigm for tsunamis. *Journal of Geophysical Research*, 113(C12012), 22.
- Matsuyama, M., Ikeno, M., Sakakiyama, T., & Takeda, T. (2007). A study of

- tsunami wave fission in an undistorted experiment. *Tsunami and Its Hazards in the Indian and Pacific Oceans. Pure and Applied Geophysics*, 164, 617–631.
- McMaster, R. L. (1960). Sediments of Narragansett Bay system and Rhode Island Sound, Rhode Island. *Journal of Sedimentary Research*, 30(2), 249–274.
- National Geophysical Data Center. (1999a). US coastal relief model—Northeast Atlantic. National Geophysical Data Center, NOAA. <https://doi.org/10.7289/V5MS3QNZ>(March 15, 2017)
- National Geophysical Data Center. (1999b). US coastal relief model—Southeast Atlantic. National Geophysical Data Center, NOAA. <https://doi.org/10.7289/V53R0QR5>(March 15, 2017).
- National Centers for Environmental Information. (2014). Tiled coastal elevation models. NOAA National Centers for Environmental Information. Dataset. <https://doi.org/10.7289/V5P55KQC>.
- Piper, D. J. W., Cochonat, P., & Morrison, M. L. (1999). The sequence of events around the epicentre of the 1929 Grand Banks earthquake: Initiation of the debris flows and turbidity current inferred from side scan sonar. *Sedimentology*, 46, 79–97.
- Shelby, M., Grilli, S. T., & Grilli, A. R. (2016). Tsunami hazard assessment in the Hudson River Estuary based on dynamic tsunami tide simulations. *Pure and Applied Geophysics*, 173(12), 3999–4037. <https://doi.org/10.1007/s00024-016-1315-y>.
- Shi, F., Kirby, J. T., Harris, J. C., Geiman, J. D., & Grilli, S. T. (2012). A high-order adaptive time-stepping TVD solver for Boussinesq modeling

- of breaking waves and coastal inundation. *Ocean Modelling*, 43–44, 36–51. <https://doi.org/10.1016/j.ocemod.2011.12.004>.
- SLIDE. (2017). Slope Stability Analysis Model. RockScience Inc. <https://www.rocscience.com/rocscience/products/slide>.
- Tappin, D. R., Grilli, S. T., Harris, J. C., Geller, R. J., Masterlark, T., Kirby, J. T., et al. (2014). Did a submarine landslide contribute to the 2011 Tohoku tsunami? *Marine Geology*, 357, 344–361. <https://doi.org/10.1016/j.margeo.2014.09.043>.
- Tehrani-rad, B., Harris, J. C., Grilli, A. R., Grilli, S. T., Abadie, S., Kirby, J. T., et al. (2015). Far-field tsunami hazard in the north Atlantic basin from large scale flank collapses of the Cumbre Vieja volcano, La Palma. *Pure and Applied Geophysics*, 172(12), 3589–3616. <https://doi.org/10.1007/s00024-015-1135-5>.
- Tehrani-rad, B., Kirby, J. T., Grilli, S. T., Grilli, A. R., & SHi, F. (2017). Continental shelf bathymetry controls the spatial distribution of tsunami hazard for the US East Coast. *Geophysical Research Letters* (in preparation).
- Towns, J., Cockerill, T., Dahan, M., Foster, I., Gaither, K., Grimshaw, A., et al. (2014). XSEDE: Accelerating scientific discovery. *Computing in Science and Engineering*, 16(5), 62–74. <https://doi.org/10.1109/MCSE.2014.80>.
- Twichell, D. C., McClennen, C. E., & Butman, B. (1981). Morphology and processes associated with the accumulation of the fine-grained sediment deposit on the southern New England shelf. *Journal of Sedimentary Research*, 51(1), 269–280. USGS. (2002). US Geological Survey “2002 interactive deaggregations”. <https://geohazards.usgs.gov/deaggint/2002/>. Accessed 2012.
- Watts, P. & Grilli, S. T. (2003). Underwater landslide shape, motion, deformation, and tsunami generation. In: *Proceedings of 13th ISOPE, Honolulu* (pp.

364–371). <http://personal.egr.uri.edu/grilli/3055p364.pdf>.

Watts, P., Grilli, S. T., Tappin, D. R., & Fryer, G. (2005). Tsunami generation by submarine mass failure, II: Predictive equations and case studies. *Journal of Waterway Port Coastal and Ocean Engineering*, 131(6), 298–310.

Yavari-Ramshe, S., & Ataie-Ashtiani, B. (2016). Numerical simulation of subaerial and submarine landslide generated tsunami waves—recent advances and future challenges. *Landslides*, 13(6), 1325–1368. <https://doi.org/10.1007/s10346-016-0734-2>.

MANUSCRIPT 2

**New Simulations and Understanding of the 1908 Messina Tsunami for
a Dual Seismic and Deep Submarine Mass Failure Source**

by

Lauren Schambach, Stéphan T. Grilli, David R. Tappin, Maria Donatella

Gangemi, and Guiseppe Barbaro

*Schambach, L., Grilli, S.T., Tappin, D.R., Gangemi, M.D., and G. Barbaro
(2020). New simulations and understanding of the 1908 Messina tsunami for a
dual seismic and deep submarine mass failure source. Marine Geology, vol. 421,
20 pp.; <https://doi.org/10.1016/j.margeo.2019.106093>*

Abstract. Over 100 years after the event, the mechanism of the 1908 Messina tsunami remains unresolved. The up to 12 m runups observed along the coasts of Sicily and Calabria cannot be explained by the coseismic tsunami, so recent studies have proposed a dual earthquake/submarine mass failure (SMF) mechanism. Here we propose a new dual source and use it to simulate tsunami generation with a three-dimensional non-hydrostatic model, coupled to a two-dimensional fully nonlinear and dispersive model, to simulate tsunami propagation to shore. We first reanalyze observations of tsunami arrival times from eyewitnesses acquired shortly after the 1908 event, and a tsunami record at a tide gauge in Malta. Similar to earlier work, this data is used to locate the likeliest tsunami source area by inverse wave ray tracing, but accounting for frequency dispersion effects on wave celerity, uncertainty in reported arrival times, and a time delay between the EQ and SMF triggering. Analyzing the seafloor morphology in this area, we identify a new SMF at the foot of the Fiumefreddo Valley, northeast of Mount Etna. The general location is consistent with earlier studies, however our SMF is much smaller ($\sim 2 \text{ km}^3$) than, e.g., that of Billi et al. (2008) and is a fairly rigid-block-slump, rather than a translational SMF. We model the block motion and simulate tsunami generation from a dual EQ/SMF source, and its propagation to shore, in higher resolution grids and based on more accurate bathymetry and topography than in earlier work. Runups and travel times agree well with observations, except for runups on either side of the Messina Straits north of the SMF, which are still underpredicted. In the far field, simulations reproduce well the arrival time and initial wave amplitudes at the Malta tide gauge. Our newly parameterized SMF and modeling improve tsunami runups simulated near the SMF location and south of it. However, as with all previous modeling of this event, additional sources are required to explain runups in the northern Messina Straits, which we suggest might

be smaller and shallower SMFs located in this area. These will be considered in future work.

2.1 Introduction

On December 28, 1908, at 5:20 am local time, a strong earthquake (EQ) struck the Messina Straits (Fig. 2.1), causing extensive destruction and reducing large areas of Messina and Reggio Calabria to rubble. The EQ was the largest and deadliest recorded in Europe in the 20th and 21st centuries and one of the largest in European history (Guidoboni et al., 2007). As the shaking subsided, eyewitnesses reported the water receding along the Sicilian coast. Soon after, at least three large tsunami waves swept the Sicilian and Calabrian coasts, causing additional damage (Platania, 1909; Baratta, 1910). The combined fatalities from the EQ and the tsunami totaled nearly 80,000 (Guidoboni et al., 2007), making the Messina 1908 event the worst such disaster in modern history in the Mediterranean region.

Field surveys of tsunami runup and eyewitness interviews were conducted soon after the event, based on which the sequence of tsunami waves and their arrival time were estimated (Table 2.1; Mercalli, 1909; Omori, 1909; Platania, 1909; Baratta, 1910). Loperfido (1909) resurveyed elevations in areas that had been measured a few years prior to the EQ and reported changes at 114 benchmarks – 82 on the Calabrian and 32 on the Sicilian side of the Straits; these served to constrain the EQ source (e.g., Meschis et al., 2019 and references therein). Rizzo (1911) published a compilation of seismological instrumental information, including geographic coordinates of recording stations, arrival times, and amplitudes of the seismic wave phases, as well as several seismograms.

From the early reports, it became quickly apparent that the Messina 1908 tsunami was anomalous because, whereas the areas most devastated by the EQ

were located in the northern, narrower section, of the Straits, the largest tsunami runups had occurred over a 40-50 km stretch of coast, south of Messina in Sicily, with the highest (12 m) around Mount Etna (Platania, 1909; Baratta, 1910; Table 2.1, Figs. 2.1-2.5). Omori (1909) noted, “the tsunami was strongest at those places where the EQ shock was not the most violent, indicating the probable non-coincidence in position of the origin of the EQ with that of the tsunami.” To explain this discrepancy, he suggested that there should have been a secondary tsunami mechanism south of the main rupture, causing seabed motion in the center of the Straits. Ryan and Heezen (1965) reported that one cable break occurred between Gazzi and Gallico (site 51 in Figs. 2.1, 2.2; Table 2.1) almost simultaneously with the EQ, consistent with its location close to the epicenter, but another one between Malta and Zante 10 hours later, which they attributed to a turbidity current generated by the EQ (MZ-CB in Fig. 2.2).

Since 1908, many studies have analyzed the Messina EQ mechanism (e.g., Schick, 1977; Caputo et al., 1981; Capuano et al., 1988; Boschi et al., 1989; De Natale and Pingue, 1991; Pino et al., 2000; Amoruso et al., 2002; Amoruso et al., 2006; Valensise et al., 2008; Aloisi et al., 2013; see also review by Pino et al., 2009 and references therein; and finally the recent work by Meschis et al., 2019). Most investigators agree that it was a Mw 7.1 magnitude EQ. Geodetic and seismological analyses indicate that the fault was normal and confined mainly within the Messina Straits (Fig. 2.1) (De Natale and Pino, 2014; Meschis et al., 2019), although some researchers still contend that there could have been a more complex rupture and set of faults involved outside of the Straits. There are still uncertainties in the dip of the fault plane, whether it was to the east or west, its strike (5 epicenters for the EQ proposed in earlier work are marked in Fig. 2.1, corresponding fault planes are shown in Fig. 2.8, and detailed parameters for the first 4 sources, which involve a

Figure 2.1: Main study area with marked footprint of grid G0 (red box) and locations of observed runup/travel times for Messina 1908 tsunami (1–58; red dots; Table 2.1); white stars mark 5 coseismic source epicenters (Table 2.2). (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

single fault plane are given in Table 2.2), and whether the fault was blind or came to seabed.

All the tsunami models based on the EQ sources proposed to date, however, significantly underpredict the recorded runups (Piatanesi et al., 1999; Tinti et al., 1999; Tinti and Armigliato, 2001; Piatanesi et al., 2008; Favalli et al., 2009). For example, Tinti and Armigliato (2003), using the Capuano et al. (1988) rupture, extended the fault southward to generate the tsunami along the coast of Sicily in

Table 2.1: Locations where tsunami runup (R) (58) and/ or arrival time (t_t) (25) were observed and reported by Platania (1909) and Baratta (1910). See Figs. 2.4, 2.5.

N	Location names	N Lat. (deg.)	E Lon. (deg.)	R (m)	t_t (min.)	t_t (used) (min.)
1	Torre Faro	38.268	15.651	0.75, 0.8	5.0-20.0	Not used
2	Fortino	38.243	15.579	2.8	-	-
3	Pace	38.238	15.574	4.7	-	-
4	Paradiso	38.223	15.564	2.4, 3.7	-	-
5	Messina 1	38.194	15.574	2.4	8.5-10.0	9.25
6	Messina 2	38.193	15.572	2.6, 3.0	-	-
7	Messina 3	38.187	15.559	2.7	-	-
8	Galati Marina	38.106	15.507	8.0	5.0-7.0	6.0
9	Briga Marina	38.070	15.487	6.3, 7.6, 8.5	-	-
10	Giampileri Marina	38.064	15.480	7.2	-	-
11	Scaletta	38.046	15.463	8.0	-	-
12	Guidomandri Marina	38.041	15.456	6.4	-	-
13	Itala Marina	38.039	15.453	7.9	-	-
14	Ali Marina	38.005	15.423	6.9, 7.3, 8.4	-	-
15	Nizza di Sicilia	37.993	15.412	5.4, 9.2	5.0	5.0
16	Roccalumera	37.970	15.390	7.0, 8.0	6.0	6.0
17	Furci	37.962	15.382	5.8	<10.0	Not used
18	Bucalo	37.948	15.369	6.1	6.0	6.0
19	Barracca	37.940	15.362	6.0	-	-
20	S. Alessio 1	37.922	15.346	7.1	<15.0	Not used
21	S. Alessio 2	37.915	15.344	11.7	-	-
22	Letojanni 1	37.910	15.343	5.0	-	-
23	Letojanni 2	37.882	15.306	3.9	4.0	4.0
24	Letojanni 3	37.877	15.303	5.9	-	-
25	Isolabella	37.853	15.299	4.0	-	-
26	Giardini 1	37.846	15.282	9.5	-	-
27	Giardini 2	37.837	15.272	4.2, 5.0, 5.3	5.0	5.0
28	Giardini (Capo Schiso)	37.823	15.275	8.4	-	-
29	Gurna	37.783	15.229	5.6	-	-
30	Fondachello	37.762	15.213	5.6	-	-
31	Riposto	37.738	15.200	5.8	10.0	10.0
32	Torre di Archirafi	37.709	15.216	5.7	-	-
33	Pozzillo	37.660	15.196	4.8	-	-
34	Stazzo	37.648	15.190	4.3	-	-
35	S. Tecla	37.641	15.183	5.7	10.0	10.0
36	Acireale	37.614	15.170	3.7	8.0	8.0
37	Capo Molini	37.577	15.171	3.5, 4.9	5.0-6.0	5.5
38	Aci Trezza	37.565	15.160	4.6, 5.0, 5.9, 7.1	8.0-10.0	9.0
39	Aci Castello	37.555	15.148	3.5	5.0	5.0
40	Ognina	37.528	15.115	3.3, 5.0	10.0	10.0
41	Catania	37.503	15.092	2.7	10.0	10.0
42	Brucoli 1	37.285	15.187	4.3	-	-
43	Brucoli 2	37.283	15.188	1.75, 3.75	8.0	8.0
44	Augusta	37.245	15.232	0.6, 1.9, 2.0	-	-
45	S Panagia	37.102	15.271	1.0	-	-
46	Siracusa	37.072	15.289	0.7, 1.6	10.0	10.0
47	Scilla	38.253	15.712	1.0	5.0-6.0	5.5
48	Cannitello	38.236	15.662	<1.0	10.0	10.0
49	Villa S Giovanni	38.210	15.637	2.5, 3.0, 3.9	10.0-12.0	11.0
50	Catona	38.184	15.640	4.0	-	-
51	Gallico	38.164	15.648	5.4	10.0-12.0	11.0
52	Reggio Calabria	38.109	15.645	6.0, 7.0	10.0	10.0
53	Pellaro	38.017	15.637	6.0, 7.0	A few min	-
54	Lazzaro	37.972	15.664	10.0	-	-
55	Capo dell'Armi	37.955	15.678	4.0	-	-
56	Melito	37.918	15.782	3.0	A few min	-
57	Bova	37.930	15.925	A few m	10.0-15.0	12.5
58	Palizzi	37.920	15.979	3.0	20.0	20.0
59	Malta	35.883	14.517	-	65.0	65.0

Table 2.2: Parameters of single fault (C-B-A-V) or multiple (M) subfault plane(s) (length L , width W and depth D at top of fault/subfault) proposed for Messina 1908 earthquake by: (C) Capuano et al. (1988); (B) Boschi et al. (1989); (A) Amoruso et al. (2002); (V) Valensise et al. (2008); (M) Meschis et al. (2019) (epicenters are marked in Fig. 2.1); all sources have a Mw 7.1 magnitude. Corresponding co-seismic tsunami source elevations computed with Okada’s (1985) method are shown in Fig. 2.8 (fault plane footprint, oriented in strike direction, is marked in each subfigure; note for (M) this is based on the superposition of 36 subfaults with varying slip covering the marked area, detail in the reference).

Fault Model	Epicenter (Lat., Lon.)	D (km)	L (km)	W (km)	Strike (deg.)	Dip (deg.)	Rake (deg.)	S (m.)	μ (N/m ²)	M_0 (N.m)
(C)	38.1907°, 15.6348°	1.1	56.7	18.5	355.8	38.6	-132.5	1.75	$3.1 \cdot 10^{10}$	$5.7 \cdot 10^{19}$
(B)	38.0460°, 15.6314°	4.0	45.0	18.0	11.5	29.0	-90.0	1.42	$3.2 \cdot 10^{10}$	$3.7 \cdot 10^{19}$
(A)	38.0820°, 15.6334°	1.5	29.8	19.8	354.5	42.4	-118.3	1.31	$3.1 \cdot 10^{10}$	$2.4 \cdot 10^{19}$
(V)	38.0645°, 15.6078°	3.0	40.0	20.0	20.0	29.0	-90.0	1.42	$3.0 \cdot 10^{10}$	$3.5 \cdot 10^{19}$
(M)	36 subfaults	0 - 5 - 10	5.0	5.32	15.8-54.6	70.0	-90.0	0.14-4.6	$3.2 \cdot 10^{10}$	$4.9 \cdot 10^{19}$

this direction. In doing so, however, they reduced the correlation with the geodetic data.

Significant progress was made, with an improved explanation for the tsunami observations, when Billi et al. (2008) proposed a submarine landslide (or submarine mass failure – SMF) as a secondary tsunami source, which they sited northeast of Mount Etna, at the foot of the Fiumefreddo Valley (Fig. 2.3), off of Giardini (sites 26, 27 in Fig. 2.1; Table 2.1). This SMF was located by performing an inverse wave ray tracing, based on 23 tsunami arrival times reported by eyewitnesses (Baratta, 1910; Table 2.1; Fig. 2.5) and using a 2D version of bathymetric data (Scrocca et al., 2004). Once located, the SMF was identified in seismic data and estimated to have a 20 km^3 volume. In their analysis, however, these authors used a constant wave celerity of 83 m/s, obtained from averaging long wave celerities, $c = (gh)^{0.5}$ for shallow and deep water depths of $h = 80$ and 2,000 m, respectively (with g the gravitational acceleration). This approximation led to overestimating both celerities and distances traveled, resulting in the SMF being located too far offshore of the largest runup area. The deeply incised seabed in the area of Billi et al.’s (2008) SMF also suggested that it was much older than 1908 (Argnani et

al., 2009a).

Piatanesi et al. (2008) modeled the tsunami from dual sources combining Billi et al.'s (2008) 20 km³ SMF and two different coseismic sources, from Boschi et al. (1989) and Amoruso et al., (2002). They used a nonlinear shallow water (i.e., non-dispersive) model in a fairly coarse 200 m resolution grid, which they initialized with the SMF surface elevation calculated with Okal and Synolakis' (2004) empirical method, combined with the two EQ sources, with a zero initial velocity assumed in both cases. Quantitatively, at all sites, the tsunami runups simulated for both EQ mechanisms alone were 3 to 4 times smaller than those recorded (Table 2.1; < 2.5 and 2 m, respectively). Qualitatively, however, for both EQ sources the shape of the normalized computed runup distributions matched well with that observed along the coasts of Sicily and Calabria. In contrast, maximum runups simulated with the dual EQ/SMF mechanism (see Fig. 2.12) were locally dominated by those from the SMF, with elevated runups (> 2 m) generated along a 60 km stretch of coast near Giardini, directly landward of the SMF, and a maximum (13 m) about 50% larger than observed in this area (8.4 – 9.5 m), suggesting that the estimated SMF volume was too large. As distance from the SMF increased, there were large discrepancies between the model and the observations. Piatanesi et al. (2008) concluded that the 1908 tsunami mechanism “remained an open question” and that further, more detailed, modeling (both in terms of model physics/resolution and bathymetric/topographic data) was required to establish it. In particular, they suggested that, to improve results, the tide gauge data available in Malta and other northern locations in the Tyrrhenian Sea should be quantitatively analyzed and modeled.

Tappin et al. (2008a), using the same bathymetric and seismic data as Billi et al. (2008), only in a 3D version, confirmed the location of their SMF as a potential

tsunami mechanism. However, their dual source tsunami simulations based on a 20 km^3 SMF and the EQ mechanisms of Tinti and Armigliato (2003), similar to Piatanesi et al. (2008), found the SMF volume to be too large to explain runup observations, which were overpredicted. They then investigated another smaller volume and shallower depth SMF (on the order of a few cubic kilometers), located southwest of Billi et al.'s (2008) SMF, at the foot of Mount Etna (Mt. Etna amphitheater; Fig. 2.3), which they interpreted as a slump, analogous to that of Papua New Guinea 1998 (e.g., Tappin et al., 2008b). Modeling of a dual EQ/SMF mechanism from this slump, using a simplified semi-empirical slump model (Watts et al., 2003, 2005; Grilli and Watts, 2005), suggested that this was as a more likely candidate for the 1908 SMF tsunami. More recent research (Chiocci et al., 2011), however, found no evidence for a slump at Tappin et al.'s (2008a) proposed location.

Argnani et al. (2009a), from newly acquired high-resolution bathymetry and seismic data, also concluded that Billi et al.'s (2008) SMF volume was overestimated and proposed a maximum of $\sim 2 \text{ km}^3$ for it. They also argued that Billi et al.'s (2008) SMF was much older than 1908.

Favalli et al. (2009) similarly questioned the SMF mechanism proposed by Billi et al. (2008) and, performing more accurate inverse ray-tracing based on a depth- and frequency-dependent celerity (Hatori, 1966), identified 5 possible SMF locations offshore Sicily. In their analysis, they assumed a short wave period of $T = 160 \text{ s}$, consistent with the short waves generated by a SMF, and used the same 23 travel times and locations as Billi et al.'s, but added a Gaussian-distributed uncertainty to each of these with a standard deviation of 60 s. They modeled tsunami generation from each of these SMFs combined with Capuano et al.'s (1988) coseismic source. The SMFs were assumed to be semi-ellipsoids of

volume 1-5 km³, with sliding motion prescribed by the semi-empirical equations of Fryer et al. (2004) (which are based on the laws of motion of Grilli and Watts 1999, 2005; and Watts et al., 2005). Initial landslide tsunami waves were generated using an analytic landslide model based on linear superposition. Simulations of tsunami propagation were then performed in a fairly coarse 200 m resolution grid, using the fully nonlinear Boussinesq model FUNWAVE (version 1.0; Kirby et al. 1998), which includes frequency dispersion effects. The best agreement with runup observations along the Sicilian coast were from the SMF with a 1.57 km³ volume located off of Nizza (site 15 in Fig. 2.1; Table 2.1). Runups in the Straits at most sites to the north of the SMF and along the Calabrian coast, however, were significantly overpredicted (Fig. 2.12).

In their recent high-resolution, morpho-bathymetric, Multibeam Echosounder (MBES) surveys, however, Ridente et al. (2014) did not find evidence of a SMF at the location proposed by Favalli et al. (2009).

2.2 Rationale

Despite some recent progress in modeling dual EQ/SMF sources, the mechanism of the 1908 Messina tsunami remains unresolved. Previous work clearly indicates that the tsunami cannot be explained by the EQ alone when considering faults confined within the Messina Straits. So, unless some complex rupture mechanism with localized high/and or slow shallow slip is invoked (such as e.g., proposed for Tohoku 2011 by Romano et al., 2014), for which there is no evidence here, a secondary mechanism is required. Multibeam Echosounder (MBES) data shows that SMFs are widespread on the steep underwater slopes of Mount Etna and Calabria (Fig. 2.3; Marani et al., 2004) and, besides analyses based on ray tracing and the MBES data as will be detailed in this work, there is various evidence from cable breaks and turbidites (detailed later) that at least 3 SMFs

occurred during the 1908 event. Hence, we propose that the secondary source mechanism for Messina 1908 was likely from at least one SMF located offshore of the eastern Sicilian coast, between Giardini and Nizza.

Earlier results of dual EQ/SMF source modeling, however, have not been able to explain all of the runup observations and the proposed SMF locations have not been confirmed in seafloor MBES surveys (e.g., Billi et al., 2008; Favalli et al., 2009). Based on these, it appears that both the SMF location and size must be reconsidered and that more than one SMF may be necessary to explain runups. The support for the likelihood of multiple seafloor failures offshore of the eastern Sicilian coast, triggered by the 1908 EQ, is provided by Comerci et al. (2015), who catalogued numerous on-land rock falls and landslides triggered by the event, especially north of Mount Etna, where the likeliest SMF mechanism to generate the 1908 tsunami in the north Messina Straits should be located. Offshore evidence for seabed failures in the northern Straits also comes from sediment cores sampled in the southern Ionian Sea (Polonia et al., 2017; Fig. 2.2), in which turbidites were found that resulted from the 1908 event and originated from both a deep water environment offshore of Mount Etna and a shallower environment south of Calabria and off northeast Sicily.

Based on earlier studies, the goals of this paper are to: (i) more accurately identify and confirm the likeliest location and parameters for the main SMF triggered by the 1908 EQ; (ii) more accurately model the tsunami coastal impact from a dual tsunami source EQ/SMF based on this; and (iii) based on this modeling, clearly identify where unexplained tsunami runups remain that may be from additional SMF sources, whose identification and modeling will be the objects of future work.

After presenting our data and methodology, the likeliest SMF location will be

identified by performing an inverse wave ray tracing analysis with a method similar to Favalli et al.'s (2009), but using a higher resolution bathymetry and including a time delay between the EQ and the SMF triggering. In the ray tracing, we use observations of tsunami arrival time from 24 locations along the coasts of Sicily and Calabria, reported by Platania (1909) and Baratta (1910) (Table 2.1; 23 of these were used in earlier work by Billi et al., 2008); these observations were all re-verified by going back to the original sources and confirming their magnitude and location. In addition, as recommended by Piatanesi et al. (2008), the arrival time from a time series of surface elevation recorded at a tide gauge in Malta (Platania, 1909) is also used in the ray tracing. Finally, the 3D multi-beam echosounder (MBES) data used by Tappin et al. (2008a), from Marani et al. (2004), is used to identify and parameterize an actual SMF, on the seafloor within the likeliest source location found from ray tracing. We then simulate the coastal impact of coseismic tsunamis generated from five published EQ sources, confirming that an additional mechanism is required for the 1908 event, which we model based on the identified SMF. Thus, simulations are performed next for a dual tsunami mechanism based on the latest of the EQ sources and our new SMF source.

In the modeling, we use higher resolution bathymetric/topographic data and computational grids (Figs. 2.1, 2.2) than in previous studies, and two state-of-the-art dispersive wave models, the: (i) 3D non-hydrostatic wave model NHWAVE (Ma et al., 2012) for tsunami generation by the SMF (combined with the initial coseismic source), and (ii) 2D fully nonlinear Boussinesq wave model FUNWAVE-TVD (Shi et al., 2012) for tsunami propagation and coastal impact in higher-resolution nested grids. Results are compared to runups surveyed at 58 locations (Fig. 2.4), the arrival times reported at 24 locations (Fig. 2.5; Platania, 1909; Baratta, 1910), and the tide gauge record in Malta (Platania, 1909) (Fig. 2.6).

2.3 Methods

2.3.1 Study area, computational grids, and bathymetric/topographic data

The study area encompasses part of southern Italy and the Mediterranean Sea surrounding the Messina Straits (Fig. 2.1), which separates eastern Sicily from western Calabria, connecting the Tyrrhenian Sea in the north to the Ionian Sea in the south. The seabed morphology along the Sicilian and Calabrian margins of the Ionian Sea is deeply gullied as a result of continuing uplift of the adjacent land areas along the Messina Straits (Antonioli et al., 2009). The seabed slopes immediately offshore are steep ($\sim 1:4$ to $1:5$) and deeply incised, with numerous channels and gullies (Fig. 2.3). These incisions lead into several major channels that may be traced southeastward towards the deeper basinal region of the southern Ionian Sea. Erosion along the Messina Straits is particularly intense, with a central erosion channel crosscutting those leading from Sicily in the west and Calabria to the east. The water depth is about 250 m in the narrow section of the Straits, and reaches more than 3,000 m in the Ionian Sea (Figs. 2.2, 2.3).

Tsunami simulations are performed in model grids defined based on new high-resolution nearshore/coastal topographic and bathymetric data, combined with the 230 m resolution European Marine Observation and Data Network (EMODnet; Marine Information Service, 2016) digital terrain model, where the former are not available. Coastal topography data was obtained from the 10 m resolution TINITALY digital elevation model (Tarquini et al., 2012) and nearshore bathymetric data from the Italian Navy, at high-resolution (< 50 m) nearshore in 3 to 20 m depth, and at 50-200 m resolution in deeper water. To compensate for the lower resolution in deep water, the bathymetric data of Marani et al. (2004) was also used. High-resolution bathymetric data in the Malta Grand Harbor was derived from Malta Environmental Planning Authority bathymetric maps (MEPA, 2013).

Figure 2.2: Footprints of computational grids (red boxes; Table 2.3) of resolution: (G4) 240 m, (G0, G5) 120 m, (G1-G3, G6) 30 m; colour scale and black contours are bathymetry/topography (m); yellow ellipses show footprints of proposed SMF before/after failure (solid/chained), as modeled in NHWAVE; yellow diamonds indicate locations of cable breaks (Ryan and Heezen, 1965; MZ-CB, G-CB: Malta and Zante and Gallico cable breaks, respectively), white triangles indicate core locations (Polonia et al., 2017), and white star marks Meschis et al.’s (2019) co-seismic source epicenter. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

All datasets were combined and interpolated onto eight horizontal Cartesian grids of resolution (Fig. 2.2; Table 2.3): (G0) 120 m, covering our main area of interest (Figs. 2.1, 2.7); (G1, G2, G3) 30 m, higher-resolution nested grids covering the coastline of the areas most impacted by the tsunami, where observations are available; (G4) 240 m, extended grid that includes the location of the Malta tide gauge

Figure 2.3: 3D view of most of the area covered by grid G0 (Figs. 2.1, 2.2), from Multibeam Echosounder (MBES; Marani et al., 2004) bathymetry; red outline marks location of Billi et al.'s (2008) SMF; black and white outlines mark proposed SMF location within it (before/after failure); box marks zoom-in area detailed in Fig. 2.7b. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

(within the modern harbor of La Valletta); (G5) 120 m, nested grid surrounding Malta; (G6) 30 m, higher-resolution nested grid surrounding the estimated location of the Malta tide gauge; (G7) 120 m, grid used for inverse wave ray tracing that extends to Malta. The grids are centered on the projection origin listed in Table 2.3, except for the oblique grids G1 and G6. Cartesian grids are used, considering the small geographic extent of the study area; for interpolation and plotting purposes, the grid (x, y) coordinates are projected using the MATLAB mapping toolbox onto their corresponding latitude-longitude coordinates.

2.3.2 Observations

Results of tsunami simulations will be compared to and validated against field observations made during or soon after the event. These are the tsunami runups

Table 2.3: Parameters of computational grids used in simulations with NHWAVE (G0), FUNWAVE (G1-G6), and in ray tracing (G7) (Figs. 2.2, 2.7a). The projection origin is at the center of each grid, or (*) at the southernmost corner for grids G1 and G6.

Grid	Mesh size (M, N)	Resolution (m)	Projection origin (Lat., Lon.)	Angle (deg. E)
G0	768, 1440	120	37.750°, 15.570°	0
G1	240, 2760	30	37.600°, 15.190°*	-62
G2	1320, 1320	30	38.070°, 15.835°	0
G3	480, 1200	30	37.430°, 15.140°	0
G4	1140, 1416	240	37.025°, 15.300°	0
G5	600, 600	120	36.000°, 14.500°	0
G6	288, 288	30	35.852°, 14.520°*	-50
G7	1392, 2376	120	37.033°, 15.132°	0

and arrival times reported by Platania (1909) and Baratta (1910), which span the coastlines of eastern Sicily and southwestern Calabria (Table 2.1, Figs. 2.4, 2.5). Observation locations are numbered from 1-46 going from north to south along the Sicilian coast, and 47-58 going from northwest to southeast along the Calabrian coast (Fig. 2.1). Unlike Billi et al.’s (2008) and Favalli et al.’s (2009) studies, which only used the Baratta (1910) report, data from both references was considered in the present analyses and directly obtained from the original publications. Additionally, a surface elevation time series recorded at the Malta tide gauge on 12/28/1908, reported by Platania (1909), was digitized and compared to simulated surface elevations (Fig. 2.6).

Platania’s (1909) and Baratta’s (1910) reports were organized by names of towns or cities in which data had been collected. Each section included a description of the coastal damage, information about where and how runup measurements were made and, for some locations, the names and/or occupation of the person who made the observation. In many of the towns and cities, the tsunami data collected by Platania was simply re-reported by Baratta, with the main exceptions being located on the Calabrian coast where Platania did not visit. Table 2.1 summarizes

our compilation of the data from both original sources (details on which source the data came from and its specific location, as noted by each author, are provided in the supplementary file, Table S1).

Figure 2.4: Tsunami runup at 58 observation sites (labeled N_s) along the Sicilian and Calabrian coasts (Table 2.1): (blue and green bullets) measured by Platania (1909) and Baratta (1910); fine grid (G1-G3) simulation results for coseismic sources: (black) Meschis et al. (2019) (yellow) Capuano et al. (1988), (green) Boschi et al. (1989), (blue) Amoruso et al. (2002), and (purple) Valensise et al. (2008); fine grid (G1-G3) simulation results for dual EQ/SMF sources: (grey x) G200, (turquoise x) E200, (orange x) G250, (red x) E220 (Table 2.5). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

Figure 2.5: Tsunami travel times at 24 observation sites (labeled N_s) along the Sicilian and Calabrian coasts (Table 2.1), measured from the start of the Messina 1908 earthquake (Dec. 28, 1908, at 5:20 am local time): (blue and green bullets) reported by Platania (1909) and Baratta (1910) based on eyewitness interviews (Table 2.1); (red/black crosses) fine grid (G1-G3) simulation results for leading elevation/depression waves, respectively (based on results of Fig. 2.11), for dual EQ/ SMF source E220 (Table 2.5; Fig. 2.10). (Travel times simulated for other dual sources were found to be similar.) (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

The British Admiralty stationed in the Grand Harbour of Malta took tide

gauge measurements at various locations from 1871 to 1926. The original charts collected by the Royal Navy were stored by the United Kingdom Hydrographic Office, and later photographed and made available through the British Oceanographic Data Center (BODC, 2018). Most tide gauge charts during this period were made by float type gauges. Platania (1909) wrote in his report that the Malta data he received and presented was from a tide gauge located at the French Creek. Admiral Astley Cooper Key described the location of the French Creek tide gauge, which was reported by Airy (1878): “The place of the tide-gauge was about 40-50 yards from the Somerset Dock (which is in the French Creek, so called by us), on the western side or left hand when entering the dock. A channel, about a yard wide and 8 or 10 yards long, led to a deep recess in which the gauge was placed. No ripple was felt from the effect either of wind or of ships moving in the neighborhood.” The gauge signal here was graphed on paper and fairly distorted (Platania, 1909; see Supplementary material, Fig. S1). After manually digitizing it and removing the low frequency tidal signal, the resulting tsunami signal ranged between $[-0.45, 0.45]$ m (Fig. 2.6). However, during the 1908 tsunami event, there were a few interruptions in the record, during which the gauge “stylus” went off charts. Note that the thickness of the drawn line in the record amounted to 3-5 minutes and, in some locations, the line slightly wrapped back upon itself (Fig. S1). This was manually corrected using our best judgment, but both of these sources of error affected the tsunami timing accuracy.

The BODC (2018) Maltese tide gauge dataset also included a chart recorded at the Ricasoli Breakwater during the 1908 tsunami event. The tsunami signal here ranges between $[-0.007, 0.007]$ m. There was a handwritten note on the chart indicating, “The tide gauge is fixed inside the root of the breakwater and is throttled.” The tsunami arrival time in this chart was in agreement with that

Figure 2.6: Time series of surface elevation at the Malta Grand Harbor French Creek tide gauge (time is on 12/28/1908; earthquake occurred at 5:20 am): (black) recorded tsunami signal (digitized from paper chart; see Supplementary material); (red/blue) computed with FUNWAVE-TVD in grid G6, for (blue) Meschis et al.'s (2019) coseismic source (Table 2.2), (red) dual EQ/SMF-E220 source (Table 2.5) (simulations are initialized in grid G4 based on NHWAVE results in grid G0 at $t = 585$ s). Note, the 70 min (or 6.5 h) tsunami arrival time corresponds to the arrival of the first (0.15-0.20 m) elevation wave in both the tide gauge record and the EQ/SMF source simulation, whereas 65 min corresponds to that of a small (-0.05 m) depression wave in the record. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

of the French Creek tide gauge, however the elevation data was not used in our comparisons, as the note indicated that it was not reliable (see Supplementary material, Fig. S2).

2.3.3 Inverse wave ray tracing

Inverse ray tracing was performed, based on geometric optic theory, from the 25 locations where tsunami arrival times were recorded (Table 2.1; Fig. 2.7a), to identify the likeliest seafloor area for the tsunami source. Our method follows that of Tappin et al. (2014), who similarly sited a secondary SMF source for the Tohoku 2011 tsunami, and takes into account both depth and frequency dispersion effects on wave celerity, assuming a dominant period T for the tsunami (i.e., a narrow banded wave train). Dispersion, which for instance was neglected in Billi et al.'s (2008) study, may be an important factor when propagating shorter period SMF tsunami wave trains. Effects of both a Gaussian distributed uncertainty in

reported arrival times and a time delay between the EQ and the SMF triggering were included in the analysis.

In forward wave ray tracing, waves such as tsunamis propagate away from their source and refract as a function of the local bathymetry $h(x, y)$ and their depth and frequency dependent phase velocity given by, $c = \omega/k = ((g/k) \tanh(kh))^{0.5}$ (with $k = 2\pi/L$ the wavenumber, $\omega = 2\pi/T$ the wave angular frequency, T the period, and L the wavelength), based on linear wave theory. Wave rays, which mark the waves' instantaneous direction of propagation, are calculated by solving the geometric optic eikonal equation. Here, this is done using a fast-marching algorithm (Tappin et al., 2014), in which rays are defined as lines orthogonal to wave fronts representing contours of wave travel time. In inverse ray tracing, the tsunami is propagated backwards along a series of rays originating at each location where travel time was recorded, which yields a locus connecting points of identical travel time for each site. When many observation points are available (such as 25 here) and well distributed along the coast as in the present case (Table 2.1, Figs. 2.5, 2.7a), the likeliest area for the tsunami source can be identified as that where the majority of travel time loci intersect.

In ray tracing, travel times can be calculated using either the wave phase c or group $c_g = 0.5c(1 + 2kh/\sinh(2kh))$ celerity (which are close to each other for moderately dispersive waves). While the latter tracks the speed of a wave train front (i.e., energy), the former should better track the propagation of an individual large leading wave (i.e., a narrow-banded wave train). Since the arrival of such a large wave was the one reported by eyewitnesses during the 1908 event, phase celerity has been preferred here. Tappin et al. (2014) had similarly used the phase celerity in their inverse ray tracing, to find the source of steep and narrow waves measured at nearshore GPS buoys associated with a SMF. As a confirmation of

Figure 2.7: (a) Probability (colour scale) of tsunami source from inverse ray tracing in grid G7 (Table 2.3), based on 25 locations with reported tsunami travel time (Table 2.1; white dots and labels; travel time in min is marked for each site and mean travel time contours are marked by red lines); a 120 s delay was assumed between EQ and SMF; the highest probability area (> 95%; contoured by black ellipse) is located off of Giardini. (b) Zoom-in of Fig. 2.3 (see marked box in figure), showing MBES bathymetry and identified SMF features off of Giardini; black contour is modeled SMF before failure (center of mass location: 37.772° N Lat., 15.410° E Lon, marked by white star in (a); dashed line across slide block marks seismic line in (c)). (c) Multichannel seismic line across slump block in (b) – see text for explanation. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

the relevance of this methodology in the present case, we shall see that, after identifying a SMF location in part based on ray tracing, the direct simulation of landslide tsunami waves all the way to the Malta tide gauge, using the complete tsunami model (FUNWAVE-TVD), will nearly recover the assumed 65 min arrival time (Fig. 2.7a). When considering the tide gauge record (site 59 in Table 2.1), both raw and filtered (Figs. S1 and 2.6), and that the ray tracing would be performed only in a coarse resolution grid (G7) this arrival time was on purpose

slightly reduced from the 70 min. arrival time of the first elevation wave, to the 65 min. time corresponding to that of a very small (-0.05 m) leading depression wave (see Fig. 2.6), to account for the poor representation of the Malta harbor in the coarse grid. (Finally, note that our ray tracing method is similar in principle to Favalli et al.'s (2009), except that we include a triggering delay for the SMF and use the phase velocity instead of the group velocity.)

Fig. 2.7a shows results of inverse ray tracing performed in the 120 m resolution grid G7, from 24 coastal observation points and the Malta tide gauge (Table 2.1; Figs. 2.5, 2.6). In the figure, the density of travel time loci has been transformed into a probability of finding the source at a given location (color scale [0, 1]), computed based on the number of points found in each grid cell from all the computed loci contours, normalized by the largest number found in one cell. A dominant period $T = 240$ s was selected based on estimated wave periods reported by various eyewitnesses (Platania, 1909; Baratta, 1910). This is a fairly long period and, based on the criterion $h < L/20$, such waves are only dispersive for $h > 1,413$ m; accordingly one should not expect a large sensitivity of the computed travel time loci to moderate changes in the selected dominant period. Additionally, as proposed by Favalli et al. (2009), uncertainty in the reported arrival times was factored in by assuming a Gaussian probability distribution with zero mean and 60 s standard deviation when computing the propagation time loci. Besides uncertainty in eyewitness reports, this standard deviation accounts for errors in travel time calculated in the ray tracing method based on the linear dispersion relationship, whereas nonlinear amplitude dispersion effects may increase tsunami celerity nearshore. However, as already mentioned above, after identifying a SMF location in part based on ray tracing, the direct simulation of landslide tsunami waves using a nonlinear and dispersive model (FUNWAVE-TVD), and resulting

travel times, will be fully consistent with observations reported in Table 2.1 used in the ray tracing. This validates the relevance of the linear ray tracing approach, within the measure of uncertainty that has been accounted for. (Note, as results of the FUNWAVE-TVD simulations shown later will yield a 6-7 min. periodicity of leading tsunami waves at most coastal sites, for verification, the inverse ray tracing will be recomputed in Supplementary Fig. S3, using a $T = 390$ s equivalent period and either c or c_g . No significant difference will be found for the highest probability area.)

The entire ray tracing analysis was repeated considering several SMF triggering delay times, ranging from 0 to 160 s after the EQ. Among all those tested, a SMF trigger delay time of 120 s resulted in the smallest region of maximum probability. This is longer than the approximately 10-25 s it would have taken P-seismic waves to travel from the EQ epicenter (Fig. 2.1) to this region and could result from the time required for pore pressure to build up in the SMF sediment, as a result of the EQ cyclic loading, to eventually cause failure. This type of mechanism and an even longer time delay were reported by Tappin et al. (2008b) for the Papua New Guinea 1998 SMF tsunami.

2.3.4 Siting of submarine mass failure

Based on the backward ray tracing, the location with a maximum probability of finding the 1908 SMF tsunami source ($> 95\%$) is identified and marked by an ellipse in Fig. 2.7a. This area is located off of Giardini (Sicily), at the foot of the Fiumefreddo Valley, northeast of the underwater slope of Mount Etna (Fig. 2.7b). This general area was also considered to be the most likely location for a SMF by both Billi et al. (2008) and Tappin et al. (2008a). This interpretation was supported by the shortest tsunami travel times to the coast reported from northeast of Mt. Etna, at Letojanni and Giardini (4 and 5 min., respectively;

Table 2.1 and Fig. 2.7a). Further support was from turbidites sampled in the southern Ionian Sea (see core locations in Fig. 2.2), analyzed by Polonia et al. (2017), which were sourced from this general location.

Upon our re-examination of the bathymetry in the area to the northeast of the Etna amphitheater, at the foot of the Fiumefreddo Valley, we identified a positive seabed feature, which we interpret as an allochthonous slide block (Figs. 2.3 and 2.7b,c). The block is 7.5 km long (downslope) and widest towards the east-southeast at 3.6 km (cross-slope), with at least a 180 m thickness. As seen in Fig. 2.7b, the block is arcuate, and banana shaped. Striking north-northeast to south-southwest within the slide block is a $\sim 5^\circ$ west-northwest dipping plane (Fig. 2.7b). Fig. 2.7c shows a north-northeast to south-southwest oriented multi-channel seismic line crossing the central region of the block. The block surface is planar as are internal seismic reflections, with the lower reflections slightly faulted. The base is irregular and marked by high amplitude reflections. The reflections underlying the block are low amplitude and undulating, slightly dipping to the south-southwest. We interpret the arcuate block plan and internal structure as the result of deformation during eastward sliding along a basal failure plan and coming to a rest after colliding with sediments on the ocean bottom (Fig. 2.7b). Comparing the morphology of the block to the shape of the Fiumefreddo Valley immediately to the west, we consider it likely that it originated here. The floor of the Fiumefreddo Valley is smooth and planar, with slopes at a 3° angle, suggesting that this is swept clean of sediment by the active uplift processes along the east coast of Sicily. There is evidence of an older gravity slide on the southern flank of the Fiumefreddo Valley (Fig. 2.7b) and active landsliding in this general region (Argnani et al., 2009b, 2014; Gross et al., 2014), which further supports our interpretation that this region is prone to landslides, with the 1908 SMF being the

most recent. The block is significantly smaller in volume than that proposed by Billi et al. (2008), with $V_s \sim 2 \text{ km}^3$.

2.3.5 Tsunami generation and propagation simulations

Numerical tsunami models

Tsunami simulations are performed in nested grids following the one-way coupling methodology validated, e.g., by Tappin et al. (2014) for the Tohoku 2011 tsunami. In this method, the tsunami sources (EQ and SMF) are generated in the coarser level parent grids (here G0 and G4; Fig. 2.2) and, during these simulations, time series of surface elevation and horizontal velocity are calculated for many coupling stations (save points) along the boundary of the next finer nested child grid level (here G1-G3 and G5-G6; Fig. 2.2). Computations are then restarted from $t = 0$ in the finer grid levels, using these time series as boundary conditions, and so forth. To prevent reflection from the open boundary of the parent grids (G0 and G4), sponge layers are specified along all the boundaries.

Two open source models are used in simulations, which are briefly described below, with more details on the first one provided in Appendix:

(1) NHWAVE (Ma et al., 2012), a three-dimensional (3D) non-hydrostatic model, is used to simulate tsunami generation by both SMF and coseismic (EQ)/SMF dual sources, in grid G0. In the horizontal direction, the model uses a Cartesian coordinate system of resolution $(\Delta x, \Delta y)$ and in the vertical direction it uses a so-called σ -layers boundary fitted grid. Only a few σ -layers are sufficient in most cases as tsunami waves are only moderately 3D; 5 layers are used in the present simulations. (Note, with 3 layers, NHWAVE provides the same order of dispersion as a Boussinesq model such as FUNWAVE-TVD, detailed hereafter, and higher-order dispersion for more layers.) As will be detailed later, for EQ sources, the initial tsunami elevation is specified on the free surface and, for rigid (i.e., non-

deforming SMFs), both the geometry and kinematics are specified as a function of time as a boundary condition on the model seafloor (e.g., Grilli et al., 2015, 2017; Schambach et al., 2019). NHWAVE has been extensively validated based on laboratory and field benchmarks (Ma et al., 2012;) The model was extended by Ma et al. (2015) and Kirby et al. (2016), to simulate tsunami generation by deforming SMFs, assumed to behave as a granular medium or a dense fluid. To do so, the model has a second bottom layer of granular material or dense fluid in which equations are depth-integrated and coupled to the overlying fluid motion; applications to SMF case studies can be found in, e.g., Grilli et al. (2017) and Schambach et al. (2019) and, for the recent Anak Krakatau subaerial collapse, in Grilli et al. (2019). Tappin et al. (2014) used NHWAVE to simulate a dual source EQ/solid SMF for the Tohoku 2011 tsunami.

For rigid SMFs, slide or slump geometry is assumed to be quasi-Gaussian with an elliptical footprint, and the law of motion is specified based on the semi-empirical equations of Grilli and Watts (1999, 2005), Grilli et al. (2002), and Watts et al. (2005) (see Appendix). This case was validated by Ma et al. (2012) based on the laboratory experiments of Enet and Grilli (2007). Simulating both rigid and deforming SMFs off of the US East Coast, Schambach et al. (2019) showed, consistent with Grilli and Watts' (2005) conclusions in their 2D simulations, that rigid SMFs cause larger coastal impact than deforming SMFs of identical initial geometry and volume. Here, in the absence of detailed field data and consistent with the slump-like features seen on the seafloor in the maximum probability area of Fig. 2.7a (Figs. 2.7b, c), a solid SMF with a slump law of motion was assumed in the simulations, in combination with various EQ sources (see details below and in Appendix).

(2) FUNWAVE-TVD (Shi et al., 2012; version 3.0 is used), a two-dimensional

(2D) fully nonlinear and dispersive Boussinesq wave model, is used to simulate the propagation of coseismic (EQ) tsunamis in grid G0 and nested grids G1-G3, or the propagation and coastal impact of dual EQ/SMF tsunamis in the nested grids G1-G3. Additionally, the model is used to propagate tsunamis to Malta in grid G4 and nested grids G5-G6 (Fig. 2.2; Table 2.3). In view of the small geographic area considered here, a Cartesian coordinate system of resolution $(\delta x, \delta y)$ is used. To improve dispersive properties, the horizontal velocity used in FUNWAVE-TVD is that corresponding to a depth $z = -0.531 h$.

Both NHWAVE and FUNWAVE-TVD are open source (available on github) and parallelized with MPI to efficiently run on large computer clusters. Both models have been extensively validated through benchmarking against laboratory and field data, as well as in the context of operational tsunami hazard assessment, for hypothetical (Ma et al., 2012; Grilli et al., 2015, 2017; Kirby et al., 2016; Guérin et al., 2018; Schambach et al., 2019), or actual (Grilli et al., 2007, 2013, 2019; Ioualalen et al., 2007; Tappin et al., 2014) landslide tsunamis and other tsunami events.

Modeling methodology

When simulating coseismic (EQ) tsunami sources only, computations are first performed with FUNWAVE-TVD in the 120 m resolution grids G0, initialized with the surface elevation computed for each source using Okada’s (1985) method (see details below), with no initial velocity. During these computations, time series of surface elevations and horizontal currents are saved at many points (“numerical wave gauges”) along the boundaries of the nested 30 m resolution grids G1-G3, in which computations will be restarted. For the dual EQ/SMF source simulations, NHWAVE is similarly initialized in grid G0 with the coseismic source elevation and, at $t = 120$ s, the slump is triggered on the seafloor; thus for times $t > 120$

s, the model is forced with the space- and time-dependent bottom geometry and kinematics of the rigid SMF (see details below). During this simulation, time series of surface elevation and horizontal currents (interpolated at $0.531h$) are similarly saved at “numerical wave gauges” along the grid G1-G3 boundaries. In both the EQ and EQ/SMF source cases, simulations are then rerun with FUNWAVE-TVD in each of the 30 m resolution nested grids G1-G3, using the time series of surface elevations and velocities previously computed in grid G0 as boundary conditions. (Note that for the oblique grid G1, velocities in grid G0 are first rotated to the grid direction.) In the G1-G3 fine grid simulations, time series of surface elevation are interpolated at many numerical wave gauges, located along the 10 m isobath, at the shortest distance from each site N_s (Fig. 2.1, Table 2.1). In both models, these simulations are performed up to $t = 3,600$ s.

FUNWAVE-TVD is also used to propagate tsunami waves to the location of the Malta tide gauge, in the larger 240 m resolution grid G4, for both EQ and dual EQ/SMF sources. To do so, the model is initialized with surface elevations at time $t = 0$ s for the EQ source, or surface elevations and horizontal currents (at $z = 0.531 h$) computed in grid G0 with NHWAVE at $t = 585$ s (9.75 min), i.e., after the SMF has stopped moving, for the dual EQ/SMF source. Simulations in grid G4 are performed for a longer time, up to $t = 13,200$ s (3h 40 min), during which results are saved at many “numerical wave gauges” along the boundary of nested grid G5. Simulations in the 120 m grid G5 are rerun, forced by results saved during simulations in grid G4; this procedure is repeated for the 30 m resolution nested grid G6 (after rotating velocities from grid G5 to its oblique direction). A time series of surface elevation is finally interpolated near the inferred location of the Malta tide gauge.

On open boundaries of grid G0, 5 km wide sponge layers are specified in

NHWAVE, for the EQ/SMF simulations, and 15 km wide in FUNWAVE-TVD, for the EQ simulations. In grid G4, sponge layers are set to 60 km on the eastern boundary and 20 km on all other boundaries, for FUNWAVE-TVD simulations. These various sponge layer widths were found to be adequate for eliminating reflection at open boundaries of the models, in the simulations. (Note, for the FUNWAVE-TVD simulations in nested grids G1-G3, no sponge layers are needed since reflection from the coast is already included in the time series used as boundary conditions; hence this is akin to an open boundary condition.)

Coseismic tsunami source generation.

For each EQ source simulated here, the coseismic seafloor deformation is first computed with a model based on Okada's (1985) method, using the fault plane parameters of the source (Table 2.2). Okada's method assumes a semi-infinite, homogeneous, elastic medium of shear modulus μ with a single fault plane, or multiple subfault planes. As is standard in coseismic tsunami simulations, this deformation is then used as a free surface elevation to initialize tsunami models, with a zero velocity. In simulations of tsunami coastal impact we considered 5 coseismic (EQ) sources proposed earlier for the Messina 1908 earthquake by (Table 2.2): (C) Capuano et al. (1988), (B) Boschi et al. (1989), (A) Amoruso et al. (2002), (V) Valensise et al. (2008), and (M) Meschis et al. (2019). The first 4 sources are defined by a single fault plane while the latter is based on 36 subfault planes.

Assuming that slip S is homogeneous on each rectangular fault (or sub-fault for (M)) plane, of length L (strike direction) by width W (dip direction), the EQ moment magnitude is, $M_o = \mu LWS$ for each fault or subfault plane (in the latter case, the total moment magnitude is the sum of that for each subfault), and the magnitude is defined as, $M_w = \log(M_o)/1.51 - 6$ (Hanks and Kanamori, 1979).

Table 2.2 shows that, despite having nearly the same magnitude, each EQ source has quite different fault parameters, shear modulus, and moment magnitudes. The initial surface elevations calculated with Okada’s method for each source are plotted in Fig. 2.8 (after interpolation onto grid G0), together with the corresponding fault plane footprint. Consistent with the different fault parameters and moment magnitudes, the elevation pattern and its magnitude are also quite different for each source. Additionally, while initial elevations for the single fault plane sources ((C), (B), (A), and (V)) are all within a $[-0.7, +0.25]$ m range, those of the multi-subfault source (with varying slip) (M) are much larger, within $[-1.28, +0.66]$ m. Despite these differences, all sources predict an initial surface depression in the narrow part of the Messina Straits.

In dual source simulations combining EQ sources with a SMF, we will use the more recent source by Meschis et al.’s (2019), as it was both more carefully and accurately parameterized using 36 subfault planes and it provides the largest coastal tsunami impact in the area onshore of the estimated location of the SMF. The coastal impact of other EQ sources will be simulated for comparison with Meschis et al.’s.

SMF tsunami source geometry and kinematics

NHWAVE is used to simulate tsunami generation by a rigid SMF failing as a rotational slump, sited within the maximum probability area (ellipse) of Fig. 2.7a, identified by inverse ray tracing. From the seafloor bathymetric features seen in the MBES and seismic data (Figs. 2.7b,c), the slump runout is estimated at $Sf = 5$ km in direction $\theta = 120^\circ$ (clockwise from N), based on which the initial SMF center of mass location is found as 37.772°N Lat., 15.410°E Lon, in depth $h_0 = 1,430$ m (white star in Fig. 2.7a and $x = x_0$ in Fig. 2.9); from the same field data, the maximum thickness for the SMF is estimated to be within the range,

Figure 2.8: Initial surface elevations of co-seismic (EQ) tsunami sources used to initialize tsunami simulations with FUNWAVE-TVD and/or NHWAVE, proposed for the Messina 1908 earthquake by: (a) Capuano et al. (1988), (b) Boschi et al. (1989), (c) Amoruso et al. (2002), (d) Valensise et al. (2008), and (e) Meschis et al. (2019). Each source was calculated with Okada's (1985) method using the fault plane parameters listed in Table 2.2. Solid black lines mark the horizontal footprint of each fault plane.

$Th = [0.2, 0.25]$ km (Fig. 2.7c). Based on this data, four different rigid SMFs are parameterized with different thickness, geometry, and kinematics, according to the methodology of Schambach et al.'s (2019) (summarized in Appendix).

Figure 2.9: Bathymetric transects in direction x of rigid slump (SMF) motion ($\theta = 120^\circ$ clockwise from N), through unfailed (red) and failed (blue) SMF deposits, assuming a: (a) quasi-Gaussian, (b) or semi-ellipsoidal geometry, for 4 cases: (solid and black) G200, E200, or (dashed and grey) G250, E220, with parameters listed in Tables 2.4, 2.5. Black and grey lines show current bathymetry with subtracted failed SMF; before simulations, all SMFs have first been moved upslope by the estimated runout distance $S_f = 5$ km, from their center of mass failed location, to the initial location: 37.772° N Lat., 15.410° E Lon (Fig. 2.7b); distances $x - x_0$ is measured from the unfailed SMF center of mass. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

Specifically, each SMF has an elliptical footprint on the seafloor, centered at its estimated initial location and, based on seafloor features (Fig. 2.7b), a downslope length $b = 7$ km and width $w = 3.5$ km (as marked in Fig. 2.2; Table

2.4). Similar to Enet and Grilli's (2007) experiments and model simulations of Ma et al. (2012), Grilli et al. (2015) and Schambach et al.'s (2019), the SMF geometry is first assumed to be quasi-Gaussian, with a shape parameter $\epsilon = 0.717$ (based on the latter value, the SMF volume is found as, $V_s = 0.351bwT_h$; see Appendix). Two different thicknesses are modeled for this geometry, $T_h = 0.2$ or 0.25 km, yielding $V_s = 1.72$ or 2.15 km³ (cases referred to as G200 and G250 in Table 2.5). However, considering the fairly plain cross-section of the sliding block shown in Fig. 2.7c, two SMFs with a semi-ellipsoidal geometry are also considered, which given the same footprint and thickness have a volume about 50% larger than for the Gaussian geometry ($V_s = 0.524bwT_h$; see Appendix); two different thicknesses, $T_h = 0.2$ or 0.22 km, are also modeled for this geometry, yielding $V_s = 2.57$ or 2.82 km³ (cases referred to as E200 and E220 in Table 2.5).

Table 2.4: Common parameters of 4 different rigid slumps modeled with NHWAVE, with an elliptical footprint (b by w) oriented in azimuth θ (of average bottom slope angle α_s), thickness Th , and runout S_f . Assumed water and SMF densities are ρ_w and ρ_s respectively. (*) Shape parameter for quasi-Gaussian (QG) geometry only (see Appendix).

SMF param.	Value	SMF param.	Value	SMF param.	Value
b	7000 m	$\epsilon*$	0.717	α_s	3°
w	3500 m	ρ_s	2140 kg/m ³	θ	120°
Th	200-250 m	ρ_w	1025 kg/m ³	S_f	5 km

Table 2.5: Parameters used to define the rigid slump kinematics of SMFs of various geometry simulated in NHWAVE (see Table 2.4 for other parameters). Note actual times the SMFs stop moving are $t_f + 120$ s. The SMFs' initial center location is: 37.772° N Lat., 15.410° E Lon. (Figs. 2.2 and 2.7).

Case	SMF Shape	Th (km)	V_s (km ³)	R (km)	t_f (s)	U_{max} (m/s)	A_0 (m/s ²)
G200	Gaussian	0.20	1.75	30.6	296	26.5	0.28
G250	Gaussian	0.25	2.15	24.5	265	29.6	0.35
E200	Semi-ellipse	0.20	2.57	34.1	312	25.2	0.25
E220	Semi-ellipse	0.22	2.82	31.3	299	26.2	0.28

Based on the MBES data (Figs. 2.7b,c), the 4 considered SMFs are assumed

to fail as rigid slumps, with their kinematics described by that of their center of mass motion $S_c(t)$ in direction θ , parallel to the seafloor (assumed to have a locally constant slope $m = \tan \alpha_s$). For the seafloor bathymetry shown in Fig. 2.7b, the average seafloor slope angle in direction θ is $\alpha_s = 3^\circ$. Grilli et al. (1999, 2005) and Watts et al. (2005) derived the slump law of motion given by Eqs. 6 and 7 (Appendix), as a function of S_f , and the SMF time of motion t_f (for which the corresponding time after the EQ is, $t = t_f + 120$ s), itself function of, $\gamma = \rho_s/\rho_w$ the specific density, $\rho_s = 2,140$ kg/m³ the SMF bulk sediment density for this volcanic slope, $\rho_w = 1,025$ kg/m³ the average sea water density (with this data, $\gamma = \rho_s/\rho_w = 2.08$), and $CM = 1$ a hydrodynamic added mass coefficient. Further, the slump maximum velocity U_{max} and initial acceleration A_0 are given in Eq. 8 (see details in Appendix). Applying Eqs. 6 to 8 (Appendix) to the 4 SMF cases defined above, whose parameters are summarized in Table 2.4, we find $R = 24.5$ to 34.1 km, $t_f = 265$ to 299 s, $U_{max} = 25.2$ to 29.6 m/s, and $A_0 = 0.25$ to 0.35 m/s² (Table 2.5), which appear to be realistic values based on previous work (e.g., Watts et al., 2005; Greene et al., 2006).

Before performing simulations with NHWAVE, each SMF, of geometry and kinematics detailed above, is first translated in the direction of motion θ by a distance S_f upslope from its failed to its initial unfailed location on the seafloor, at: 37.772°N Lat., 15.410°E Lon. (Fig. 2.7b). This is done by subtracting the SMF geometry from its current resting location on the seafloor and adding it at the estimated initial location. Fig. 2.9 shows vertical transects in direction through each SMF, both at their current failed (blue) and their estimated initial (red) locations. Parameters computed for each SMF (Table 2.5) are used in the slump law of motion, together with the SMF geometry, to specify the bottom boundary conditions in NHWAVE (see details in Grilli et al., 2015; Schambach, 2018; and in

Appendix).

2.4 Tsunami simulation results

2.4.1 Coseismic sources

Tsunami generation was first simulated with FUNWAVE-TVD in grid G0 (Fig. 2.2; Table 2.3), for each of the 5 EQ sources shown in Fig. 2.8 (Table 2.2) and then by one-way coupling in grids G1-G3 (Fig. 2.2; Table 2.3). These are not detailed here, but an animation of simulated surface elevations, EQ_Messina.mp4, is provided as supplementary material for the Meschis et al. (2019) source. Fig. 2.4 compares computed and measured runups, for each EQ source, at the observation sites located within grids G1-G3. Consistent with earlier work, the present high-resolution simulations confirm that the coseismic tsunamis generated by any of the proposed sources cannot explain observations at almost all locations. Except for 2 locations where they reach 4-4.5 m, the simulated runups are typically in the 1-2 m range, compared to 1-12 m for the observed runups, with many of those over 5 m. The impact at the Malta tide gauge of the tsunami generated by Meschis et al.'s (2019) EQ source (Table 2.2) is simulated next with FUNWAVE-TVD. Computations are first performed in grid G4, and then by one-way coupling in nested grids G5 and G6 (Fig. 2.2; Table 2.3). Fig. 2.6 shows that, at the location of the Malta tide gauge, the simulated coseismic tsunami arrives 15 minutes after the beginning of the tsunami record and does not reproduce the initial tsunami wave train. For $t > 7.25$ h, however (after 7:15 am in Fig. 2.6), both the amplitude and period (phase) of simulated tsunami waves agree well with observations. The addition of a SMF to these simulations is detailed next.

2.4.2 Dual coseismic-SMF sources

Simulations of dual coseismic-SMF source were performed using NHWAVE and FUNWAVE-TVD, as detailed above. Meschis et al.'s (2019) EQ source was

used in combination with one of 4 different rigid SMF (slump) scenarios, of volume $V_s = 1.72\text{-}2.82 \text{ km}^3$, sited and parameterized as detailed before (Fig. 2.2; Tables 2.4, 2.5). Fig. 2.8e shows Meschis et al.'s (2019) coseismic surface elevation used to initialize NHWAVE at $t = 0$. As inferred from the ray tracing, the SMFs are triggered at $t = 120 \text{ s}$.

Fig. 2.10 shows instantaneous surface elevations computed in grids G0-G3, for the dual EQ/SMF source E220 (Table 2.5), up to $t = 1,500 \text{ s}$ (a 3D animation of the SMF motion and corresponding tsunami generation, EQ_SMF_3D_Messina.mp4, as well as an animation of surface elevations EQ_SMF_Messina.mp4, are provided as supplementary material). At $t = 150 \text{ s}$, Fig. 2.10a shows that the SMF has already generated dipole waves of elevations much larger than those of the coseismic tsunami. These have started propagating both towards Sicily, as a leading depression wave ($< -3 \text{ m}$), and offshore towards Calabria as a leading elevation wave ($> 3 \text{ m}$). At the same time, coseismic waves are causing a small coastal impact in the northern areas of the domain. At $t = 300 \text{ s}$ (5 min), Fig. 2.10b shows that, while coseismic waves continue to mostly and mildly impact the northern coastal areas of the domain, the large depression waves from the SMF have already reached the Sicilian coast, between Roccalumera and Giardini (sites 16-28); this is consistent with the 4-6 min tsunami arrival times reported in this area, as an ocean withdrawal (Table 2.1; Figs. 2.5, 2.7a). At $t = 600 \text{ s}$ (10 min), after the SMF stopped moving at $t = 429 \text{ s}$ (Table 2.5), Fig. 2.10c shows that the onshore moving waves it generated are causing significant coastal inundation along a large section of the Sicilian coast, while offshore moving waves have continued propagating away from the SMF source area and started impacting the southern coast of Calabria, with positive or negative elevation waves. On the eastern Sicilian coast, the first elevation wave from the SMF is strongly impacting Riposto (site 31; Fig.

2.1) and its surrounding sites, while the SMF leading elevation wave is moving in the south towards Catania (site 41) and Brucoli (site 42); to the north of the Sicilian coast, a large coastal impact can also be seen in Giardini (sites 26-28). As the simulation progresses, Fig. 2.10d shows, at $t = 900$ s, that Capo Molini (site 37) is strongly impacted on the Sicilian coast and, at the same time, Melito (site 56) on the Calabrian coast is feeling the brunt of the tsunami impact. At $t = 1,200$ s, Fig. 2.10e shows that as waves continue propagating away from the SMF source area and refracting/reflecting in coastal areas, causing increasingly complex wave patterns, on the Sicilian coast, large impacts occur at sites 1-2 (Torre Faro, Fortino), 41 (Catania), 33-35 (Pozzillo, Stazzo, S. Tecla), 42-43 (Brucoli), and on the Calabrian coast at sites 52-53 (Reggio Calabria), 56 (Melito), and 58 (Palizzi). Finally, at $t = 1,500$ s, while most of the large waves have left the computational domain (i.e., been absorbed in sponge layers), Fig. 2.10f shows that, although elevations are smaller, due to multiple reflections between the Sicilian and Calabrian coasts, some sites continue being significantly impacted by tsunami waves.

Both the snapshots of Fig. 2.10 and the corresponding animation show that, as the EQ and SMF tsunami wave trains propagate (and more strongly for the latter), as a result of dispersion, they develop into an increasing large number of leading waves and a more pronounced tail of oscillations. This is confirmed in Fig. 2.11, which for the same case shows time series of surface elevations computed in the high resolution grids G1-G3 at 10 numerical gauges. These were sited along the 10 m isobath, at the shortest distance from observation sites distributed along the SE coasts of Sicily (a-f) and Calabria (g-j), i.e.: Messina 1 (5), Nizza di Sicilia (15), Giardini 2 (27), S. Tecla (35), Ognina (40), Brucoli 2 (43), Cannitello (48), Gallico (51), Lazzaro (54), and Bova (57) (Fig. 2.1, Table 2.1). At most gauges, as reported by eyewitnesses, the tsunami wave train is composed of at least 3 large

Figure 2.10: Instantaneous surface elevation (colour scale in meter; at different scales) computed with NHWAVE in grid G0, and with FUNWAVE-TVD in grids G1-G3 (Fig. 2.2), for the combined EQ (Meschis et al., 2019)/SMF (E220, $V_s=2.82$ km³, triggered at 120 s; Table 2.5) source, at $t=(a)$ 150, (b) 300, (c) 600, (d) 900, (e) 1200, and (f) 1500 s from the start of the event (Dec. 28, 1908, at 5:20 am local time). White bullets mark locations of observation sites 1–58 (Table 2.1; Fig. 2.1).

waves of dominant period 6 to 7 mins, over which some oscillations of both smaller elevation and period are superimposed. As would be expected, gauges located directly onshore of the SMF in Sicily experience some of the largest waves (sites 27, 35, 40), but so do sites located south of the SMF (sites 40, 43), due to shelf refraction (see animation). In the far-field, the SMF tsunami also causes a large

impact along the coast of Calabria (for instance at site 57).

The time progression of tsunami coastal impact seen in Fig. 2.10 was found to be quite similar for each of the 4 simulated dual coseismic/SMF sources (Table 2.5) (details are not shown here). Thus, for the case of Fig. 2.10 (Meschis et al.'s EQ/E220 SMF), Fig. 2.5 compares the tsunami arrival times computed in grids G1-G3 for the first depression and elevation wave with travel times t_t estimated from eyewitness reports, at 24 observation sites (Table 2.1). Computed travel times were obtained from time series of surface elevation similar to those shown in Fig. 2.11, located near all the observation sites within grids G1-G3. Overall, there is a good agreement between simulations and observations, particularly in view of the approximate nature of many of the eyewitness reports (Table 2.1). In simulations, sites 1-36 and 50-55, which correspond to the northern part of the study area (Fig. 2.1), are first impacted (within 2-9 mins) by a leading depression wave, mostly small and associated with the EQ source (e.g., Fig. 2.10a). Later on (within 2.5-13.5 mins), the same sites are impacted by the first elevation wave, which is usually much larger (e.g., see Fig. 2.11). Overall, observations agree best with the simulated arrival times of the first elevation wave, except at sites 23-28, which are located directly onshore, south, or slightly north of the SMF source, where they best agree with the arrival of a (larger in this case) depression wave associated with the SMF (see, e.g., Fig. 2.11c). In the figure, the largest discrepancy in arrival times occur at the southern site 43 where the tsunami was reported to have arrived much earlier (8 and 10 min) than predicted in simulations for the first elevation wave (15 min).

Fig. 2.12a shows the envelope of maximum surface elevations computed in grids G0-G3 (whichever is finest), for the EQ/E220 SMF dual source case already shown in Figs. 2.10 and 2.11 (which was found to cause the largest coastal runups;

Figure 2.11: Same case as in Fig. 2.10. Time series of surface elevations computed in fine grids (G1-G3) at numerical gauges located along the 10 m isobath, at the shortest distance from sites (Fig. 2.1, Table 2.1) $N_s =$ (a) 5 (Messina 1); (b) 15 (Nizza di Sicilia); (c) 27 (Giardini 2); (d) 35 (S. Tecla); (e) 40 (Ognina); (f) 43 (Brucoli 2); (g) 48 (Cannitello); (h) 51 (Gallico); (i) 54 (Lazzaro); (j) 57 (Bova). The periodicity between the main leading waves is about 6–7 min.

see below), and Fig. 2.12b shows a zoom-in of this envelope along the Sicilian coast, in the most impacted area centered directly onshore of the SMF location. In these figures the maximum coastal inundation is clearly irregularly distributed, as a results of both the SMF tsunami directionality (Tappin et al., 2008b, 2014; Schambach et al., 2019) and wave focusing/defocusing effects caused by nearshore bathymetric features (Figs. 2.1-2.3). Based on the envelope in Fig. 2.12 and similar results obtained for the other 3 other dual source cases (Table 2.5), tsunami runup was computed at each of the 58 observation sites and plotted in Fig. 2.4, together with observations (Table 2.1). In the figure, both the largest runup at most sites and best agreement with observations are obtained for the E220 dual EQ/SMF

source, except in the northern parts of the Sicilian and Calabrian coasts (sites 1-20 and 50-54; Fig. 2.1), where tsunami impact is underpredicted.

Figure 2.12: Same case as in Figs. 2.10 and 2.11. Envelope of maximum surface elevations computed in grids G0-G3 (colour scale in meter) up to $t = 3600$ s: (a) over G0's footprint; (b) along the Sicilian coast (chained line box in (a)), in the area of maximum runup onshore of the SMF. Solid black boxes mark footprints of 30 m resolution grids G1-G3 (Fig. 2.2); SMF initial and final locations are marked by solid/dashed ellipses, respectively, and arrow shows failure's direction θ ; white bullets (numbered in (b)) are observation sites; Table 2.1). Note the different scale in each figure.

More specifically, in Fig. 2.4, simulations reproduce observations reasonably well for sites 21-43 along the Sicilian coast, south of S. Alessio (site 21). In particular, the largest runups observed directly onshore of our proposed SMF location, at sites 22-33, are well captured in simulations (e.g., in good agreement with the observed 9.7 m runup in Giardini/Capo Schiso (sites 26-28), and the 5.8/5.7 m runups in Riposto/Torre di Archirafi (sites 31-32)). As indicated before, tsunamis generated by the coseismic sources only very significantly underpredict runups,

particularly in this area. North of S. Alessio (site 21), however, runups are systematically underpredicted in simulations up to Galati Marina (site 8), which is directly across the Straits from Reggio Calabria (Fig. 2.1); this includes the largest observed runup of 11.7 m at site 21 near the train tunnel in S. Alessio, for which simulations only predict a 6 m runup. Along the Calabrian coast, the far-field impact of the SMF well explains runup observations at the easternmost sites 55-58 (Capo dell’Armi-Palizzi), however, the maximum runups on the western side of it (sites 50-55) are significantly underestimated, especially near Reggio Calabria, Pellarò, and Lazzaro (sites 52-54), where maximum runups are predicted to be approximately 2 m, compared to reported values of 6-10 m (Table 2.1).

Finally, Fig. 2.6 compares the surface elevation time series simulated in grid G6 (Fig. 2.2), for the E220 dual Meschis et al.’s EQ/SMF source and for the EQ source only, to that measured during the event, at the estimated location of the Malta Grand Harbour French Creek tide gauge (an animation of surface elevations simulated in grids G4-G6, from $t = 585$ - $5,400$ s (10-90 min), EQ_SMF_Malta.mp4, is provided as supplementary material). In the dual EQ/SMF source simulations shown in the figure, a small initial elevation wave arrives approximately 70 min ($t = 6.5$ h in the figure) after the earthquake, in good agreement with the first elevation wave recorded at the tide gauge. A sequence of larger lower period (6.5-7 min.) waves then follows during the first 40 min. of simulated tsunami impact (6.5 to 7.17 h), which is consistent with the SMF; at least one similar large and lower period wave (6.5 min) was recorded at the gauge, although its “stylus” apparently went off charts around 6.88 h. Both the initial elevation wave and these larger oscillations are not present in the EQ source only simulation results. In the dual source simulations maximum amplitudes at the tide gauge are $[-.85, 0.88]$ m, which is larger than in the recorded signal; however, as indicated before, there were breaks

in the latter at the time of largest tsunami impact, so the maximum recorded amplitude is unknown (above +0.4 m). Later on (for time greater than 7.17 h in the figure), waves simulated for the dual source have similar amplitude and period (phase) as both the recorded signal and those of the EQ source simulations.

2.5 Discussion

The mechanism of the 1908 Messina tsunami has been a source of controversy for over 100 years, with the major challenge being the identification of a secondary tsunami source mechanism, besides the earthquake, to explain the recorded impacts. Since Omori (1909) first proposed an additional tsunami source from seabed movement in the Messina Straits (considered in recent work and in this study to be a submarine landslide or SMF), its location has never been positively identified. A combination of hydroacoustic data (bathymetric and seismic) and numerical modeling confirms that the most likely location of the seabed movement is offshore of Mount Etna, at the foot of the Fiumefreddo Valley. This area is particularly prone to landsliding, and SMFs previously attributed to 1908 here have been identified as landslides (Billi et al., 2008) or slumps (Tappin et al., 2008). Subsequent research using hydroacoustic data, however, has disproved these (Argnani et al., 2009; Chiocci et al., 2011). Omori (1909) originally suggested a location off Giampilieri, and numerical modeling of a SMF located to the south, offshore of Nizza (Fig. 2.3; Favalli et al., 2009), satisfied some of the observations, but no evidence for a SMF has been found here.

The simulations of a dual coseismic/SMF tsunami source for the 1908 event, presented here, agree well with the tsunami runups observed in the region south of S. Alessio (observation sites 21-43), especially directly onshore and on either sides of the SMF location (sites 22-33), both in terms of arrival time and runup magnitude (Figs. 2.4, 2.5). This we attribute to our newly identified SMF at

the foot of the Fiumefreddo Valley, parameterized as the tsunami source, and modeled as a slump motion on higher-resolution grids of bathymetric/topographic data than used in previous studies, together with a state-of-the-art 3D landslide tsunami generation model. The tsunami recorded at the Malta tide gauge (Fig. 2.6) provides a strong constraint on the location of the highest probability source area in Fig. 2.7a, as well as on tsunami magnitude, i.e., the size of the first few waves approximately matches observations, within the large uncertainty of the recorded signal amplitude and exact tide gauge location. Therefore, the modeled tsunami elevation at this gauge supports both our SMF location, its volume and block failure mechanism.

Fig. 2.13 compares the measured runups (from Fig. 2.4) to our results for the EQ/E220 SMF dual source, and the earlier dual EQ/SMF source simulations of Piatanesi et al. (2008) and Favalli et al. (2009). For comparison, we also show results of our simulations of Meshchi et al.'s (2019) single EQ source, which underpredicts observations by a factor of 3-4. In the area of maximum runup along the Sicilian coast (Sites 22-33; Figs. 2.1, 2.12), our results clearly improve on all of the dual mechanism models. Besides, as noted above, these models are not based on realistic mechanisms: the landslide of Billi et al. (2008) simulated by Piatanesi et al. (2008), was too old to have been associated with the event as well as being too large a volume (Tappin et al., 2008); and the SMF modeled by Favalli et al. (2009) was not validated by marine surveys.

In the north of the Ionian Sea, along both Sicilian and Calabrian coasts (sites < 22 and > 49 ; Fig. 2.1; Table 2.1), our simulations still underpredict tsunami runups, confirming the problems encountered by other authors in earlier simulations. To explain the runups observed here, at least one additional, earthquake-triggered, smaller-volume SMF located at shallower depth between the latitudes

Figure 2.13: Tsunami runup at 58 observation sites along the Sicilian and Calabrian coasts (Table 2.1): (blue and green bullets) measured by Platania (1909) and Baratta (1910); simulations for: (black) Meschis et al.’s (2019) EQ source in high resolution grids G1-G3, (red x) E220 dual EQ/SMF source (Table 2.5), (orange +) Favalli et al. (2009), and (green +) Piatanesi et al. (2008). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

of 37.95° and 38.15°N is required (Figs. 2.1-2.4; Table 2.1); or another mechanism such as slow slip or more complex rupture mechanism, for which there is no evidence as of yet. No landslides were found yet along the Calabrian margin (Ridente et al., 2014), located close to the cable break off Gallico identified by Ryan and Heezen (1965; G-CB in Fig. 2.2). The evidence from the seismoturbidites in the southern Ionian Sea, confirms this region to be their landslide source as well as off northeast Sicily, although these have still to be identified. The landslides off Calabria, however, are relatively small volume with scarps in water deeper than 500 m. They have low tsunamigenic potential, hazardous only if they failed simultaneously (Ridente et al., 2014). In addition, tsunamis from landslides on the Calabrian margin would generate a positive wave striking the Sicilian coast, and a negative wave against the Calabrian coast, which contradicts the observations. Another explanation here is that the small depression wave due to the coseismic source that arrived immediately in this area, could have been what the observers first noticed. Local coastal failures south of Messina, identified in the geodetic survey of Loperfido (1909), suggest a potential tsunami mechanism here, however,

travel times reported in Messina were 8-10 minutes so, unless there was a long delay between the EQ and slide failure, this seems unlikely.

A fundamental problem with using inverse wave ray tracing to locate the primary tsunami mechanism, is the assumption that the arrival times correspond to tsunami waves generated by a single dominant mechanism. When considering multiple mechanisms such as a coseismic and multiple SMFs, it is difficult (if not impossible) to distinguish which travel times correspond to which source. An additional complication is the possibility/probability that there were several SMFs triggered at different times, with different delays to the EQ. A more rigorous analysis of the tsunami travel times combined with knowledge of where SMF deposits are located may help determine the appropriate timing of events.

Another area of interest for a potential SMF suggested by our results is off Capo Molini (site 37), where the travel time is 5 minutes (Fig. 2.5). In our simulations, the tsunami reached this area and those south of it well after the reported arrival times, and runups were underestimated there as well (Figs. 2.4, 2.13). A SMF in this area would not necessarily have to be very large, as the reported runups were only around 3-4 m.

Despite the uncertainty on its exact location, results at the Malta tide gauge for the first 40 min. of tsunami impact (about 6.5 to 7.15 h; Fig. 2.6) similarly confirm that a dual tsunami source mechanism is necessary to reproduce the tsunami arrival time and recorded surface elevation. The proposed coseismic sources are all located too far north for their waves to arrive at the recorded time (for the Meschis et al.'s (2019) EQ source, they arrive 12 minutes late, at about 6.7 h; Fig. 2.6), and the generated waves are initially too small. At later time, both the EQ only and dual EQ/SMF simulation results are in good agreement with the tide gauge signal, which shows oscillations of about 18-24 min. period, with initially a

[-0.3 to +0.3 m] amplitude, that gradually decreases with time. Tinti and Giuliani (1983) pointed out that these are consistent with seiches of period ranging from 23-25 min. in the La Valletta (Malta) harbor area, that would have been excited by the leading waves of the tsunami. Small amplitude seiches with this periodicity were present in the tide gauge record prior to tsunami arrival (Fig. S1), and larger seiches would have been easily excited by the tsunami. In time domain simulations with FUNWAVE-TVD, seiching is part of the solution if such resonances are able to occur based on the specified bathymetry. Finally, based on the tide gauge results, if the EQ had triggered several SMFs, some located north of our SMF, the waves they generated would have arrived later in the record, at a time when the EQ waves, and related seiches, can explain the observations quite well (Fig. 2.6). This implies that such additional SMFs would have caused small far-field waves and, hence, been both of small volume and have occurred in fairly shallow water.

2.6 Conclusions

Our new modeling of the tsunami associated with the 1908 Messina EQ, consistent with previous studies, shows that an EQ source alone cannot cause runups of large enough magnitude to match observations along the Sicilian and Calabrian coasts. The EQ tsunami runups there are too small by a factor of 3-4 and, at Malta Grand Harbour (La Valletta), the tsunami arrives significantly later than the tide gauge record indicates. From our inverse wave ray tracing analysis we confirmed that the most likely location of the main tsunamigenic source was northeast of Mount Etna and, in this area, at the foot of the Fiumefreddo Valley, we identified a $\sim 2 \text{ km}^3$ volume SMF (a slump) in the MBES data that appears recent enough to be associated with the 1908 event. From the ray tracing, this SMF was estimated to be triggered 120 s after the EQ. New dual source tsunami generation and propagation simulations were performed that combined this SMF and an EQ

source from the most reasonable and recently proposed rupture mechanisms. Unlike earlier studies, which used empirical or analytical landslide models and coarse grids, our tsunami generation was simulated with a non-hydrostatic 3D model (NHWAVE), in a 120 m resolution grid, and tsunami propagation and coastal impact were simulated with a 2D nonlinear and dispersive model (FUNWAVE-TVD), in 30 m resolution grids, both using higher-resolution bathymetric and topographic data than in earlier studies. Runups from this dual mechanism agree better with observations on the Sicilian coast, onshore and on either sides of the SMF, than previously found. Along the northeastern Sicily and western Calabrian coasts, however, where elevated runups were also reported, these are still underestimated in our modeling. At the tide gauge in the Malta Grand Harbor, the modeled dual-source tsunami agrees well with observations of the tsunami signal (both phase and amplitude).

Our new work provides strong evidence for the contribution of more than one SMF to the 1908 tsunami. One or a few additional landslides are necessary in the northern Messina Straits, with available MBES data suggesting these to be offshore of the highest unexplained runups, off Calabria, as identified by Ridente et al. (2014); additional landslides here would explain the observed high local impacts. This multiple SMF interpretation is further supported by the evidence from cable breaks at this location (Ryan and Heezen, 1965) and from the seismoturbidites identified by Polonia et al. (2017) in the southern Ionian Sea, sourced from offshore Mount Etna and the northern Messina Straits. Although our inverse ray tracing analysis provides the strongest evidence for our primary (SMF) tsunami source location off of Giardini, we recognize its limitations in identifying multiple tsunami sources, especially those that are secondary in tsunami generation. Despite of this, Fig. 2.5a shows an area with a 50-60% probability, north of our main SMF area,

that could be associated with other SMF sources, for which further modeling is required to test their tsunamigenic potential.

Finally, note, considering the large uncertainty in the EQ rupture mechanisms proposed to date, that we cannot entirely rule out that the large tsunami impact observed in the Straits would have been caused by a more complex rupture mechanism (perhaps involving slow slip) than has currently been proposed, or even additional EQ sources outside of the Straits, as hypothesized by Polonia et al. (2012). As we have discussed in detail in this paper, however, SMFs are widespread on the steep underwater slopes of the Straits and further south (see Fig. 2.3), and appear to be our most likely candidates for additional sources. Recently, in the Palu, Indonesia, 2018 event, it was reported that at least 14 coastal slides (subaerial and underwater), most of them tsunamigenic, had been triggered by an earthquake of similar magnitude as here (Mw 7.5), on similarly steep underwater slopes (e.g., Carvajal et al., 2019).

In conclusion, we believe that the primary mechanism of the 1908 Messina tsunami is at least one SMF at our proposed location, the 1908 EQ, and likely other SMFs further north, for which there is evidence but, as yet, no positive identification. Accordingly, in future work, assuming additional sub-bottom and geologic data are available, we will attempt to site, parameterize, and investigate additional SMFs for their contribution to the tsunami recorded in the northern Messina Straits.

Supplementary data to this article can be found online at <https://doi.org/10.1016/j.margeo.2019.106093>.

Acknowledgements LS and SG gratefully acknowledge support from grant CMMI-15-35568 of the United States (US) National Science Foundation (NSF), Engineering for Natural Hazard program. LS and SG performed numerical simu-

lations on “Extreme Science and Engineering Discovery Environment” (XSEDE) HPC resources, as part of NSF grant ACI-1548562. FUNWAVE-TVD is open source software available at <http://github.com/fengyanshi/FUNWAVE-TVD/>. NHWAVE is open source software available at <http://github.com/jimkirby/nhwave/>. Thanks to Sebastian Krastel for provision of the seismic section in Fig. 2.7c. Dr. Meschis is gratefully acknowledged for providing his recent EQ source parameters (Meschis et al., 2019) and Dr. Lorito for suggesting that we use the Malta tide gauge record. DRT publishes with the permission of the CEO, British Geological Survey.

Appendix: NHWAVE model set-up for simulating tsunami generation by rigid SMFs

Figure 2.14: (a) Planview and (b) vertical cross-section (in direction θ), for rigid SMF of QG or SE geometry, with an elliptical footprint (b, w) on the seafloor (Fig. 2.2), maximum thickness T_h and elevation below seafloor $\zeta(\xi, \chi, t)$ (Eqs. (1) and (2)), initially centered at (x_0, y_0) , moving in downslope direction ξ , in azimuthal direction θ (deg.) from North, with a center of mass motion $S_c(t)$ measured parallel to the mean local seafloor slope of angle α ; (x, y) denote the longitudinal and latitudinal horizontal directions, respectively, and $h_0(x, y)$ and $h(x, y, t)$ the initial and instantaneous bathymetry, respectively.

As discussed in the main text, the 3D non-hydrostatic model NHWAVE (Ma et al., 2012) is used to simulate generation and initial propagation for tsunamis triggered by rigid SMFs, whose motion and geometry are specified on the seafloor

as bottom boundary conditions (e.g., Grilli et al., 2015, 2017; Schambach et al., 2019). For a rigid body motion, the SMF kinematics is identical to that of its center of mass motion, $S_c(t)$ parallel to the seafloor of slope $m = \tan \alpha_s$. The SMF is assumed to have an elliptical footprint on the seafloor (Figs. 2.2, 2.14), of major axis b (downslope) and minor axis w (cross-slope) geometry, and a maximum thickness at its center, T_h . The SMF geometry is assumed to have either quasi-Gaussian (QG) or semielliptical (SE) cross-sections; for the QG geometry, steepness is controlled by a shape parameter ϵ (here $\epsilon = 0.717$) (Table 2.4; Fig. 2.14).

Specifically, the SMF elevation is parameterized below seafloor, over its elliptical footprint, as,

$$\zeta(\xi, \chi) = \frac{T_h}{1 - \epsilon} \max\{0, \text{sech}(k_b \xi) \text{sech}(k_w \chi) - \epsilon\} \quad (2.1)$$

for the QG geometry, with $k_b = 2C/b$, $k_w = 2C/w$, and $C = \text{acosh}(1/\epsilon)$, and as,

$$\zeta(\xi, \chi) = T_h \max\{0, \left(1 - \frac{4\xi^2}{b^2} - \frac{4\chi^2}{w^2}\right)^{1/2}\} \quad (2.2)$$

for the SE geometry, where (ξ, χ) are the local down-slope and cross-slope horizontal coordinates, rotated in the direction of SMF motion θ (Fig. 2.14), i.e.,

$$\begin{aligned} \xi &= (x - x_0) \cos \theta - (y - y_0) \sin \theta - S_c(t) \cos \alpha_s \\ \chi &= (x - x_0) \sin \theta + (y - y_0) \cos \theta \end{aligned} \quad (2.3)$$

With this parameterization, the SMF volume is found for the QG geometry as,

$$V_s = bwT_h \frac{I_2}{C^2} \left(\frac{\frac{I_1}{I_2} - \epsilon}{1 - \epsilon} \right) \quad \text{with} \quad (2.4)$$

$$I_{1,2} = \int_0^C f(\mu) d\mu; g(\mu) d\mu$$

and

$$\begin{aligned} f(\mu) &= \text{sech}(\mu \text{atan}(\sinh(g(\mu)))) \\ g(\mu) &= \text{acosh} \left(\frac{\text{sech} \mu}{\epsilon} \right) \end{aligned} \quad (2.5)$$

For the specified value $\epsilon = 0.717$, we find, $C = 0.8616$, $I_1 = 0.4804$, $I_2 = 0.5672$, and $V_s = 0.351bwT_h$. For the SE geometry, the SMF volume is simply given by, $V_s = (\pi/6)bwT_h = 0.524bwT_h$.

For rigid slumps failing in a circular motion of large radius R and small angular displacement $\Delta\Phi$ Grilli and Watts (1999, 2005) and Watts et al. (2005) derived an analytical law of motion for the center of mass, by expressing a dynamic balance between weight, buoyancy, inertia, and basal friction forces, as,

$$S_c(t) = \frac{S_f}{2} \left(1 - \cos \pi \frac{t}{t_f} \right) \text{ for } 0 \leq t \leq t_f \quad (2.6)$$

where $S_f \approx R\Delta\Phi$ is the slump runout, and t_f its time of motion given by,

$$t_f = \pi \sqrt{\frac{R}{g} \frac{\gamma + C_M}{\gamma - 1}} \text{ with } R \simeq \frac{b^2}{8T_h} \quad (2.7)$$

where $\gamma = \rho_s/\rho_w$ is the SMF specific density, ρ_s the bulk sediment density, ρ_w the average seawater density, C_M a hydrodynamic added mass coefficient, and g the gravitational acceleration. Note that, Grilli et al. (1999, 2005) assumed that hydrodynamic drag is negligible as compared to basal friction, in particular, considering that the slump only achieves a moderate maximum velocity, as shown next.

From Eq. (2.6), the slump velocity and acceleration are found as,

$$U_c(t) = \frac{\partial S_c}{\partial t} = U_{max} \sin \pi \frac{t}{t_f}$$

and

$$A_c(t) = \frac{\partial^2 S_c}{\partial t^2} = A_0 \cos \pi \frac{t}{t_f} \quad (2.8)$$

respectively, for which maximum velocity and initial acceleration are given by, $U_{max} = (\pi/2)S_f/t_f$ and $A_0 = \pi U_{max}/t_f$, respectively.

Given an existing or reconstructed (unfailed) seafloor bathymetry $h_0(x, y)$, with a mild slope (such that, $\sin \alpha_s \approx 0$), the instantaneous seafloor depth above the moving SMF is given by,

$$h(x, y, t) = h_0(x, y) + \zeta\{\xi(x, y, t), \chi(x, y, t)\} - \zeta\{\xi(x, y, 0), \chi(x, y, 0)\} \quad (2.9)$$

with $\zeta(\xi, \chi)$ given by Eqs. (2.1) or (2.2), for the QG or SE geometry, respectively, and $\xi(x, y)$ and $\chi(x, y)$ given by Eq. (2.3), as a function of the slump law of motion $S_c(t)$ of Eqs. (2.6) and (2.7). Based on Eq. (2.9), the vertical seafloor velocity specified as bottom boundary condition in NHWAVE, together with $h(x, y, t)$, is given by,

$$\frac{\partial h}{\partial t}(x, y, t) = k_b \cos \alpha_s \left(\zeta + \frac{\epsilon T_h}{1 - \epsilon} \right) U_c \tanh(k_b \xi) \quad (2.10)$$

for instance for the QG geometry, where ζ is given by Eq. (2.1) and U_c by Eq. (2.8); a similar equation is derived for the SE geometry.

Note that the reader can find many typical plots of the center of mass motion, velocity and acceleration for rigid slumps, as compared to those of deforming SMFs, in our earlier published work (e.g., Schambach et al., 2019).

References

- Airy, G.B. (1878). On the tides at Malta. *Philosophical Transactions of the Royal Society of London*, vol. 169, pp. 123-138; doi: 10.1098/rstl.1878.0006
- Aloisi, M., Bruno, V., Cannavò, F., Ferranti, L., Mattia, M., Monaco, C., and M. Palano (2013). Are the source models of the M 7.1 1908 Messina Straits earthquake reliable? Insights from a novel inversion and a sensitivity analysis of levelling data. *Geophysical Journal International*, vol. 192, no. 3, pp. 1025-1041; doi:10.1093/gji/ggs062.
- Amoruso, A., Crescentini, L., and R. Scarpa (2002). Source parameters of the 1908 Messina Straits, Italy, earthquake from geodetic and seismic data. *Journal of Geophysical Research*, vol. 107, no. B42080; doi:10.1029/2001JB000434.
- Amoruso, A., Crescentini, L., Neri, G., Orecchio, B., and R. Scarpa (2006). Spatial relation between the 1908 Messina Straits earthquake slip and recent earthquake distribution. *Geophysical Research Letters*, vol. 33, L17309; doi:10.1029/2006GL027227.
- Antonioli, F., Ferranti, L., Fontana, A., Amorosi, A., Bondesan, A., Braitenberg, C., Dutton, A., Fontolan, G., Furlani, S., Lambeck, K., Mastronuzzi, G., Monaco, C., Spada, G., and P. Stocchi (2009). Holocene relative sea-level changes and vertical movements along the Italian and Istrian coastlines. *Quaternary International*, vol. 206, pp. 102-133; doi: 10.1016/j.quaint.2008.11.008.
- Argnani, A., Chiocci, F.L., Tinti, S., Bosman, A., Lodi, M.V., Pagnoni, G., and F. Zaniboni (2009a). Comment on “On the cause of the 1908 Messina tsunami,

- southern Italy”, by Andrea Billi et al. *Geophysical Research Letters*, vol. 36, L13307; doi:10.1029/2009GL037332.
- Argnani, A., Brancolini, G., Bonazzi, C., Rovere, M., Accaino, F., Zgur, F. and E. Lodolo (2009b). The results of the Taormina 2006 seismic survey: Possible implications for active tectonics in the Messina Straits. *Tectonophysics*, vol. 476, no. 1-2, pp. 159-169; doi: 10.1016/j.tecto.2008.10.029.
- Argnani, A., Mazzarini, F., Bonazzi, C., Bisson, M., and I. Isola (2013). The deformation offshore of Mount Etna as imaged by multichannel seismic reflection profiles. *Journal of Volcanology and Geothermal Research*, vol. 251, pp. 50-64; doi: 10.1016/j.jvolgeores.2012.04.016.
- Baratta, M. (1910). La catastrofe sismica calabro-messinese (28 dicembre 1908). Relazione alla Società Geografica Italiana. Roma: Presso la Società Geografica Italiana.
- Billi, A., Funiciello, R., Minelli, L., Faccenna, C., Neri, G., Orecchio, B., and D. Presti (2008). On the cause of the 1908 Messina tsunamis, southern Italy. *Geophysical Research Letters*, vol. 35, L06301; doi:10.1029/2008GL033251.
- Boschi, E., Pantosti, D., and G. Valensise (1989). Modello di sorgente per il terremoto di Messina del 1908 ed evoluzione recente dell’area dello Stretto. *Atti VIII Convegno G.N.G.T.S.*, Roma, 245-258.
- BODC (2018). Scanned images of tide gauge charts from Grand Harbour, Malta, 1871-1926. <https://www.bodc.ac.uk/resources/inventories/edmed/report/6282/> British Oceanographic Data Centre.
- Capuano, P., De Natale, G., Gasparini, P., Pingue, F. and R. Scarpa (1988). A model for the 1908 Messina Straits (Italy) earthquake by inversion of leveling

- data. Bulletin of the Seismological Society of America, vol. 78, no. 6, pp. 1930-1947.
- Caputo, M., Pieri, L., Barbarella, M., Gubellini, A., Russo, P., and R. Console (1981). Geophysical and geodetic observations in the Messina Straits. Tectonophysics, vol. 74, no. 1-2, pp. 147-154; doi:10.1016/0040-1951(81)90132-3.
- Chiocci, F.L., Coltelli, M., Bosman, A., and D. Cavallaro (2011). Continental margin large-scale instability controlling the flank sliding of Etna volcano. Earth and Planetary Science Letters, vol. 305, no. 1-2, pp. 57-64; doi: 10.1016/j.epsl.2011.02.040.
- Comerci, V., Vittori, E., Blumetti, A. M., Brustia, E., Di Manna, P., Guerrieri, L., Lucarini, M., and L. Serva (2015). Environmental effects of the December 28, 1908, Southern Calabria-Messina (Southern Italy) earthquake. Natural Hazards, vol. 76, no. 3, pp. 1849-1891; doi: 10.1007/s1169-014-1573-x.
- De Natale, G., and F. Pingue (1991). A variable slip fault model for the 1908 Messina Straits (Italy) earthquake, by inversion of levelling data. Geophysical Journal International, vol. 104, no. 1, pp. 73-84; doi:10.1111/j.1365-246X.1991.tb02494.x.
- De Natale, G. and N. A. Pino (2014). Comment on 'Are the source models of the M 7.1 1908 Messina Straits earthquake reliable? Insights from a novel inversion and sensitivity analysis of levelling data' by M. Aloisi, V. Bruno, F. Cannavò, L. Ferranti, M. Mattia, C. Monaco and M. Palano. Geophysical Journal International, vol. 197, no. 3, pp. 1399-1402; doi:10.1093/gji/ggu063.
- Enet, F., and S. T. Grilli (2007). Experimental study of tsunami generation by three-dimensional rigid underwater landslides. Journal of Waterway,

- Port, Coastal, and Ocean Engineering, vol. 133, no. 6, pp. 442-454; doi: 10.1061/(ASCE)0733-950X(2007)133:6(442).
- Favalli, M., Boschi, E., Mazzarini, F., and M. T. Pareschi (2009). Seismic and landslide source of the 1908 Straits of Messina tsunami (Sicily, Italy). *Geophysical Research Letters*, vol. 36, L16304; doi:10.1029/2009/GL039135.
- Fryer, G.J., Watts, P., and L. F. Pratson (2004). Source of the great tsunami of 1 April 1946: A landslide in the upper Aleutian forearc. *Marine Geology*, vol. 203, no. 3-4, pp. 201-218; doi:10.1016/S0025-3227(03)00305-0.
- Greene, H.G., L.Y. Murai, P. Watts, N.A. Maher, M.A. Fisher, C.E. Paull, and P. Eichhubl (2006). Submarine landslides in the Santa Barbara Channel as potential tsunami sources. *Natural Hazards and Earth System Science*, vol. 6., no. 1, pp. 63-88; doi: 10.5194/nhess-6-63-2006.
- Grilli, S.T., and P. Watts (1999). Modeling of waves generated by a moving submerged body. Applications to underwater landslides. *Engineering Analysis with Boundary Elements*, vol. 23, no. 8, pp. 645-656; doi:10.1016/S0955-7997(99)00021-1.
- Grilli, S.T., Vogelmann, S., and P. Watts (2002). 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; doi: 10.1016/S0955-7997(01)00113-8.
- Grilli, S.T., and P. Watts (2005). Tsunami generation by submarine mass failure, I: modeling, experimental validation, and sensitivity analysis. *Journal of Waterway, Port, Coastal, and Ocean Engineering*, vol. 131, no. 6, pp. 283-297; doi: 10.1061/(ASCE)0733-950X(2005)131:6(283).

- Grilli, S.T., O'Reilly, C., Harris, J.C., Bakhsh, T. T., Tehranirad, B., Banihashemi, S., Kirby, J.T., Baxter, C.D.P., Eggeling, T., Ma, G., and F. Shi (2015). Modeling of SMF tsunami hazard along the upper US East Coast: detailed impact around Ocean City, MD. *Natural Hazards*, vol. 76, pp. 705-746; doi: 10.1007/s11069-014-1522-8.
- Grilli, S.T., Shelby, M., Kimmoun, O., Dupont, G., Nicolsky, D., Ma, G., Kirby, J.T. and F. Shi (2017). Modeling coastal tsunami hazard from submarine mass failures: effect of slide rheology, experimental validation, and case studies off the US East coast. *Natural Hazards*, vol. 86(1), pp. 353-391; doi:10.1007/s11069-016-2692-3.
- Gross, F., Krastel, S., Chiocci, F.L., Ridente, D., Bialas, J., Schwab, J., Beier, J., Cukur, D., and D. Winkelmann (2014). Evidence for Submarine Landslides Offshore Mt. Etna, Italy, in: Krastel, S. (Ed.), *Submarine Mass Movements and Their Consequences*. Springer, pp. 307-316.
- Guidoboni, E., Ferrari, G., Mariotti, D., Comastri, A., Tarabusi, G., and G. Valensise (2007). CFT4Med, Catalogue of Strong Earthquakes in Italy (461 B.C. – 1997) and Mediterranean Area (760 B.C. – 1500). INGV-SGA. Available from <http://storing.ingv.it/cfti4med/>.
- Hanks, T.C., and H. Kanamori (1979). A moment magnitude scale. *Journal of Geophysical Research*, vol. 85, no. 5, pp. 2348-2350; doi: 10.1029/JB084iB05p02348.
- Hatori, T. (1966). Vertical Displacement in a Tsunami Source Area and the Topography of the Sea Bottom. *Bulletin of the Earthquake Research Institute*, vol. 44, pp. 1449-1464.

- Kirby, J. T., G. Wei, Q. Chen, A. B. Kennedy, and R. A. Dalrymple (1998). FUNWAVE 1.0: Fully nonlinear Boussinesq wave model documentation and user's manual. Research Report CACR-98-06, Cent. for Appl. Coastal Res., Univ. of Del., Newark.
- Kirby, J.T., Shi, F., Nicolsky, D., and S. Misra (2016). The 27 April 1975 Kiti-mat, British Colombia, submarine landslide tsunamis: a comparison of modeling approaches. *Landslides*, vol. 13, no. 6, pp. 1421-1434; doi: 10.1007/s10346-016-0682-x.
- Loperfido, A. (1909). Livellazione geometrica di precisione eseguita dall'I.G.M. sulla costa orientale della Sicilia, da Messina a Catania, a Gesso ed a Faro Peloro e sulla costa occidentale della Calabria da Gioia Tauro a Melito di Porto Salvo. In *Relazione della Commissione Reale incaricata di designare le zone più adatte per la ricostruzione degli abitati colpiti dal terremoto del 28 dicembre 1908 o da altri precedenti*. Roma: Accademia Nazionale dei Lincei, pp. 131-156.
- Ma, G., Shi, F., and Kirby, J.T. (2012). Shock-capturing non-hydrostatic model for fully dispersive surface wave processes. *Ocean Modelling*, vol. 43-44, pp. 22-35; doi:10.1016/j.ocemod.2011.12.002.
- Ma, G., Kirby, J.T., Hsu, T.J., and Shi, F. (2015). A two-layer granular landslide model for tsunami wave generation: theory and computation. *Ocean Modelling*, vol. 93, pp. 40-55.
- Marani, M.P., Gamberi, F., Bortoluzzi, G, Carrara, G., Ligi, M., and D. Penitenti (2004). Seafloor bathymetry of the Ionian Sea. From Seafloor to Deep Mantle: Architecture of the Tyrrhenian Backarc Basin. *Mem. Descr. Carta Geol. It.*, vol. XLIV, 216 pp., APAT, Rome.

- Marine Information Service (2016). EMODnet Digital Bathymetry (DTM). Marine Information Service. <http://doi.org/10.12770/c7b53704-999d-4721-b1a3-04ec60c87238>. MEPA (2013). ERDF 156 data: Developing National Environmental Monitoring Infrastructure and Capacity, Malta Environment and Planning Authority, Floriana, Malta.
- Mercalli, G. (1909). Contributo allo studio del terremoto calabro-messinese del 28 dicembre 1908. Atti del Reale Istituto d’Incoraggiamento di Napoli, Ser. VI-7.
- Meschis, M., Roberts, G.P., Mildon, Z.K., Robertson, J., Michetti, A.M., and J.F. Walker (2019). Slip on a mapped normal fault for the 28th December 1908 Messina earthquake (Mw 7.1) in Italy. Scientific reports, vol. 9, no. 1, pp. 6481.
- Okada, Y. (1985). Surface deformation due to shear and tensile faults in a half space. Bulletin of the Seismological Society of America, vol. 75, no. 4, pp. 1135-1154.
- Okal, E.A., and C.E. Synolakis (2004). Source discriminants for near-field tsunamis. Geophysical Journal International, vol. 158, no. 3, pp. 899-912; doi:10.1111/j.1365-246X.2004.02347.x.
- Omori, E. (1909). Preliminary report on the Messina-Reggio earthquake of Dec. 28, 1908. Bulletin of the Imperial Earthquake Investigation Committee of Japan, vol. 3, no. 2, pp. 37-46.
- Piatanesi, A., Tinti, S., and E. Bortolucci (1999). Finite-elements simulations of the 28 December 1908 Messina Straits (southern Italy) tsunami. Physics and Chemistry of the Earth (A), vol. 24, no. 2, pp. 145-150.

- Piatanesi, A., Lorito, S., and F. Romano (2008). Il grande maremoto del 1908: analisi e modellazione. In: Il terremoto e il maremoto del 28 dicembre 1908: analisi sismologica, impatto, prospettive (G. Bertolaso, E. Boschi, E. Guidoboni, G. Valensise, DPC-INGV), pp. 183-196.
- Pino, N. A., Giardini, D., and E. Boschi (2000). The December 28, 1908, Messina Straits, southern Italy, earthquake: Waveform modeling of regional seismograms. *Journal of Geophysical Research*, vol. 105, no. B11, pp. 25,473-25,492; doi:10.1029/2000JB900259.
- Pino, N. A., Piatanesi, A., Valensise, G., and E. Boschi (2009). The 28 December 1908 Messina Straits Earthquake (MW 7.1): A Great Earthquake throughout a Century of Seismology. *Seismological Research Letters*, vol. 80, no. 2, pp. 243-259; doi:10.1785/gssrl.80.2.243.
- Platania, G. (1909). Il maremoto dello Stretto di Messina del 28 dicembre 1908. *Bollettino della Società Sismologica Italiana*, vol. 13, pp. 369-458.
- Polonia, A., Nelson, C.H., Romano, S., Vaiani, S.C., Colizza, E., Gasparotto, G., and L. Gasperini (2017). A depositional model for seismo-turbidites in confined basins based on Ionian Sea deposits. *Marine Geology*, vol. 384, pp. 177-198; doi:10.1016/j.margeo.2016/05/010.
- Ridente, D., Martorelli, E., Bosman, A., and F. L. Chiocci (2014). High-resolution morpho-bathymetric imaging of the Messina Straits (Southern Italy). New insights on the 1908 earthquake and tsunamis. *Geomorphology*, vol. 208, pp. 149-159; doi:10.1016/j.geomorph.2013.11.021.
- Rizzo, G. B. (1911). Sulla propagazione dei movimenti prodotti dal terremoto di Messina del 28 dicembre 1908. *Memorie della Reale Accademia delle Scienze di*

- Torino, Serie 2, vol. 61, pp. 355-417.
- Ryan, W.B.F., and B.C. Heezen (1965). Ionian Sea Submarine Canyons and the 1908 Messina Turbidity Current. *Geological Society of America Bulletin*, vol. 76, pp. 915-932.
- Schambach L., Grilli S.T., Kirby J.T. and F. Shi 2018. Landslide tsunami hazard along the upper US East Coast: effects of slide rheology, bottom friction, and frequency dispersion. *Pure and Applied Geophysics*, pp. 1-40; doi.org/10.1007/s00024-018-1978-7.
- Schick, R. (1977). Eine seismotektonische Bearbeitung des Erdbebens von Messina im Jahre 1908. *Geologisches Jahrbuch*, E11, pp 3-74.
- Scrocca, D., et al. (Eds.) (2004). *CROP Atlas*, 197 pp., APAT, Rome.
- Shi, F., Kirby, J.T., Harris, J.C., Geiman, J.D. and S.T. Grilli (2012). A high-order adaptive time-stepping TVD solver for Boussinesq modelling of breaking waves and coastal inundation. *Ocean Modelling*, vol. 43-44, pp. 36-51; doi:10.1016/j.ocemod.2011.12.004.
- Tappin, D.R., Watts, P., Grilli, S.T., Dubosq, S., Billi, A., Pophet, N., and M.P. Marani (2008a). The 1908 Messina tsunami. Some comments on the source: earthquake, submarine landslide or a combination of both? *Eos Trans. AGU*, 89(53), Fall Meet. Suppl., Abstract S41D-07.
- Tappin, D.R., Watts, P., and S.T. Grilli (2008b). The Papua New Guinea tsunami of 17 July 1998: anatomy of a catastrophic event. *Natural Hazards and Earth System Sciences*, vol. 8, no. 2, pp. 243-266; doi: 10.5194/nhess-8-243-2008.
- Tappin, D.R., Grilli, S.T., Harris, J.C., Geller, R.J., Masterlark, T., Kirby, J.T., Shi, F., Ma, G., Thingbaijam, K.K.S., and P.M. Mai (2014). Did a submarine

- landslide contribute to the 2011 Tohoku tsunami? *Marine Geology*, vol. 357, pp. 344-361; doi:10.1016/j.margeo.2014.09.043.
- Tarquini, S., Vinci, S., Favalli, M., Doumaz, F., Fornaciai, A., and L. Nannipieri (2012). Release of a 10-m-resolution DEM for the Italian territory: Comparison with global-coverage DEMs and anaglyph-mode exploration via the web. *Computers and Geosciences*, vol. 38, no. 1, pp. 168-170; doi:10.1016/j.cageo.2011.04.018.
- Tinti, S., Armigliato, A., Bortolucci, E., and A. Piatanesi (1999). Identification of the source fault of the 1908 Messina earthquake through tsunami modelling. It is a possible task? *Physics and Chemistry of the Earth (B)*, vol. 24, no. 5, pp. 417-421.
- Tinti, S., and A. Armigliato (2001). Impact of large tsunamis in the Messina Straits, Italy: the case of the 28 December 1908 Tsunami. In: Hebenstreit, G.T. (eds) *Tsunami research at the end of a Critical Decade. Advances in Natural and Technological Hazards Research*, vol. 18, pp. 139-162.
- Tinti, S., and A. Armigliato (2003). The use of scenarios to evaluate the tsunami impact in southern Italy. *Marine Geology*, vol. 199, pp. 221-243; doi:10.1016/S0025-3227(03)00192-0.
- Valensise, G., Basili, R., and P. Burrato (2008). La sorgente del terremoto del 1908 nel quadro sismotettonico dello Stretto di Messina. In: *Il terremoto e il maremoto del 28 dicembre 1908: analisi sismologica, impatto, prospettive* (G. Bertolaso, E. Boschi, E. Guidoboni, G. Valensise, DPC-INGV), pp. 161-182.
- Watts, S. T. Grilli, J. T. Kirby, G. J. Fryer, and Tappin, D. R. (2003). Landslide tsunami case studies using a Boussinesq model and a fully nonlinear tsunami

generation model. *Natural Hazards and Earth System Sciences*, vol. 3, pp. 391-402.

Watts, P., Grilli, S.T., Tappin, D.R., and J.G. Fryer (2005). 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; doi: 10.1061/(ASCE)0733-950X(2005)131:6(298).

MANUSCRIPT 3

**Dual earthquake/landslide source modeling of the 2018 Palu tsunami
generation and hazard**

by

Lauren Schambach, Stéphan T. Grilli, and David R. Tappin

*Schambach, L., Grilli, S.T., and D.R. Tappin (2020). Dual earthquake/landslide
source modeling of the 2018 Palu tsunami generation and hazard. In preparation
for submission to Frontiers.*

Abstract. The Mw 7.5 earthquake that struck Central Sulawesi, Indonesia, on September 28, 2018, was rapidly followed by coastal landslides and destructive tsunami waves within Palu Bay. This earthquake was supershear and predominantly strike-slip, with most published mechanisms predicting limited seabed uplift/subsidence, which make it an unlikely source of the up to 10.5 m runups recorded in the southern portion of the bay. Scientific debate has continued over the tsunami mechanism; earthquake, coastal landslides, or a combination of both. Published research has been inconclusive, with some studies simulating an earthquake generated tsunami as explaining most observations, with others focusing solely on landslide sources. For the latter, most simulations are based on hypothetical landslides not identified in post-tsunami onland field and bathymetric surveys. In this work, we simulate the tsunamis generated by the earthquake models of Jamelot et al. (2019), Socquet et al. (2019), and Ulrich et al. (2019), alone and in combination with seven coastal landslides that were confirmed by the field and bathymetric surveys (Liu et al., 2020; Takagi et al., 2019) which, from video evidence, produced significant waves. To generate and propagate the tsunamis, we use a combination of two numerical models, the 3D non-hydrostatic wave model NHWAVE and the 2D Boussinesq wave model FUNWAVE-TVD. Both models address the physics of wave frequency dispersion, which is important for modeling landslide tsunamis identified in the event. The coastal landslides are modeled in NHWAVE as granular material. Our combined earthquake and coastal landslide cases recreate the observed tsunami runups except for those in the southeast of the bay where they were most elevated (10.5 m). With regard to the timing of tsunami impact on the coast, results for the dual landslide/earthquake sources are in reasonable agreement with reconstructed time series at several locations around the bay, particularly using the models of Ulrich et al. (2019). In agreement with

other studies, our work suggests an additional tsunami mechanism is necessary in the southeast of Palu Bay to explain observations there. Using partial information from bathymetric surveys, we site an additional landslide in the SE of Palu Bay and show that, when simulated together with the other slides and Ulrich et al. (2019)’s earthquake, results can better explain observations in the Southeast. This supports the need for future marine geology work in this area.

3.1 Introduction

On September 28, 2018 at 6:02:45 PM local time (10:02:45 AM UTC), a magnitude Mw 7.5 earthquake struck Central Sulawesi, Indonesia, with the epicenter located approximately 70 km north of the city of Palu (USGS, 2018) (Fig. 3.1). The earthquake ruptured the Palu-Koro fault system, a predominantly strike-slip left-lateral fault (e.g., Socquet et al., 2019), in which large earthquakes have occurred in the past (Watkinson and Hall, 2017), two of which have caused tsunamis in Palu Bay within the last century, in 1927 and 1968 (Prasetya et al., 2001). The rupture was supershear (Bao et al., 2019; Socquet et al., 2019; Fang et al., 2019) (i.e., it propagated faster along the fault than the local shear wave velocity), and the resulting ground motions caused widespread damage throughout the western Central Sulawesi region. The earthquake induced liquefaction and landsliding, both inland (Bradley et al., 2019; Watkinson and Hall, 2019; Miyajima et al., 2019), and on the coast (Sassa and Takagawa, 2019). From eyewitness accounts and video evidence, almost immediately after the earthquake, numerous coastal areas along Palu Bay experienced landslides (see locations of main ones marked LS- in Fig. 3.1b), which were rapidly followed by destructive tsunami waves (Arikawa et al., 2018; Carvajal et al., 2019). The earthquake, ground liquefaction, landslides, and tsunamis resulted in 4,340 fatalities and approximately 68,500 buildings were damaged or destroyed (BNPB, 2019).

Figure 3.1: (a) Study area with *base model grid* (BG) over Palu Bay (white box), epicenter location (USGS, 2018, yellow star), and traces of local faults used in earthquake source models by: (blue) Jamelot et al. (2019), (red) Socquet et al. (2019), and (green) Ulrich et al. (2019); (b) Footprint of BG with locations of: (red dots) measured runups (Mikami et al., 2019; Omira et al., 2019; Pribadi et al., 2018; Putra et al., 2019; Widiyanto et al., 2019), (black dots) surface elevation time series inferred from shore-based videos (Carvajal et al., 2019, GM: Grand Mall, KN: KN Hotel, T: Talise, D: Dupa, P: Pantoloan, W: Wani), (yellow dots) observed landslides, (diamond) location of aircraft at 10:04:33 UTC, that filmed coastal landslides (Fig. 3.2).

Tsunami elevation time series were measured at two tide gauges, in the far-field at Mamuju (-2.66° N, on W Sulawesi) and the near-field in Pantoloan Port (P in Fig. 3.1b) (BIG, 2018). In the months following the event, field surveys were conducted by international research teams who recorder earthquake and tsunami damage, landslides, and tsunami runup and inundation. Red dots in Fig. 3.1b show locations of runups collected by Mikami et al. (2019); Omira et al. (2019); Pribadi et al. (2018); Putra et al. (2019); Widiyanto et al. (2019). Earthquake shaking, tsunami generation (particularly by coastal landslides; e.g., Fig. 3.2), and various tsunami impacts were recorded in many amateur videos posted on social media, that were downloaded and analyzed (e.g., Carvajal et al., 2019), providing critical

information on the timing and sequence of events. From the field and marine surveys, videos, and survivor accounts, it was clear that the earthquake, coastal landslides, and tsunamis closely followed each other, with major tsunami impact often taking place within minutes of the first shaking. Runups were highest in the south of the bay, reaching up to 10 m in the SE (Fig. 3.3). At many locations where the coast was low lying, inundation only penetrated a short distance inland, which was seen as evidence of a landslide rather than earthquake main generation mechanism (e.g. Muhari et al., 2018).

Figure 3.2: Composite picture created from aircraft pilot video (Mafella, video 38 in Carvajal et al., 2019), showing waves generated by coastal landslides LS-B,C,D,E and F* (Fig. 3.1b), at $t \simeq 108$ s into the event (aircraft location in Fig. 3.1b). “Boat” and “NBoat” mark where waves were also recorded on a small boat, as well as active subaerial slides (video 39 in Carvajal et al., 2019).

Although there are now many tsunami simulations (e.g., Heidarzadeh et al., 2019; Takagi et al., 2019; Carvajal et al., 2019; Pakoksung et al., 2019; Gusman et al., 2019; Jamelot et al., 2019; Ulrich et al., 2019; Goda et al., 2019; Nakata et al., 2020; Sepúlveda et al., 2020; Liu et al., 2020, see summary of studies characteristics in Table 3.1), the mechanism, earthquake, coastal landslides, or both in combination, is still uncertain. Most coseismic tsunami sources (e.g., USGS, 2018; Socquet et al., 2019; Jamelot et al., 2019; Yolsal-Çevikbilen and Taymaz, 2019), are based on a primarily horizontal strike-slip earthquake mechanism, with limited vertical seabed motion (1-2 m). This, theoretically, should not be strongly

Figure 3.3: (a,c) Runups R (black dots) measured in Palu Bay by international teams (Mikami et al., 2019; Omira et al., 2019; Pribadi et al., 2018; Putra et al., 2019; Widiyanto et al., 2019) (black dots in (b,d,e)). Lines in (a,c) are runups simulated with FUNWAVE for 3 coseismic sources: (blue) Jamelot et al. (2019), (red) Socquet et al. (2019), and (green) Ulrich et al. (2019), in (—) 30 m resolution BG, and (- - -) 7.5 m resolution EG/SG grids (white footprints in figures (b,d,e)). Maximum surface elevations computed with each source are color scales in: (b,d,e), respectively.

tsunamigenic and should not generate the elevated tsunami runups recorded from the south of the bay. There are many different interpretations of where the rupture is located under the Bay. In some earthquake models, the rupture crosses Palu Bay as a simple, north-south, trending, connection (e.g. Socquet et al., 2019; Ul-

Study/Paper	Numerical Model	Bathymetry Grid	EQ Source	Landslide Source
Heidarzadeh et al. (2019)	COMCOT	5 arc-sec (~ 150 m) resolution in Palu Bay, derived from GEBCO	USGS (2018), using Okada (1985) plus Tanioka and Satake (1996) for seafloor displacement	None, conclude landslides may contribute
Takagi et al. (2019)	Delft 3D Flow, hydrostatic NLSWE mode	20 m resolution, derived from BIG14	None	LS-F*, assume initial surface depression equal to volume lost based on survey
Carvajal et al. (2019)	COMCOT	23 m resolution, derived from BIG14	Socquet et al. (2019) and USGS (2018), using Okada (1985) and Tanioka and Satake (1996) for seafloor displacement	None, conclude landslides may contribute
Pakoksung et al. (2019)	Two-layer model solving NLSWE for fluids of different density	30 m resolution, derived from DEMNAS and BATNAS	None	6 landslides located in areas with reported subsidence, 4 hypothesized, modeled as a dense fluid
Gusman et al. (2019)	COMCOT	0.48 arc-sec (~ 14 m), topography derived from 2011 LiDAR and inSAR, bathymetry derived from BIG14 and BATNAS	Joint inversion method of SAR vertical displacement measurements and Pantoloan tide gauge waveform data	3 coastal landslides located off Palu City modeled as solid blocks based off equations of Enet and Grilli (2007)
Jamelot et al. (2019)	NLSWE (Heinrich et al. (1998) and Hébert et al. (2001))	200 m resolution grid with two nested 10 m grids in Palu City and Pantoloan, derived from DEMNAS and BATNAS	USGS (2018) and their own, using Okada (1985) and Tanioka and Satake (1996) for seafloor displacement	None
Ulrich et al. (2019)	Coupled EQ + tsunami model, Seisol + Storm-Flash2D	Triangular grid with maximum resolution 80 m in Palu Bay, derived from BATNAS	Their own, modeled with Seisol and coupled to the wave model, use Tanioka and Satake (1996) to account for steep slopes	None, conclude that landslides are most likely secondary contributors to the overall tsunami
Goda et al. (2019)	NLSWE (Goto et al. (1997))	Nested to 10 m, derived from DEMNAS and BATNAS	USGS (2018) considering different spatial slip distribution and rake angles	None, conclude that landslides may contribute
Sepúlveda et al. (2020)	COMCOT	~ 45 m, derived from BIG14 and DEMNAS	USGS (2018); Socquet et al. (2019) and their own 12 sources	Use Carvajal et al. (2019) time series estimates to invert for initial elevations at suspected slide locations, not modeled together with EQ sources
Nakata et al. (2020)	JAGURS	10 m, derived from DEMNAS and BATNAS	USGS (2018) and Jamelot et al. (2019)	Modeled 6 hypothetical slides as granular material with Titan2D
Liu et al. (2020)	COMCOT	0.012 arc-min (~ 20 m) derived from BIG14	None	7 coastal landslides evident in their bathymetric survey, modeled using semi-empirical equations

Table 3.1: Overview of main characteristics of earlier studies of the 2018 Palu event and tsunami modeling.

rich et al., 2019) (Fig. 3.1b). In others, there is change in direction under the bay (e.g. Jamelot et al., 2019) (Fig. 3.1b), and some locate the rupture along the west coast (e.g. Song et al., 2019). There is no resolution to these alternatives from the bathymetric data in the deeper waters of the bay (Frederik et al., 2019), because there is no feature identified as a possible rupture.

In this work, we will model three coseismic sources, from Jamelot et al. (2019) and Socquet et al. (2019) who derived parameters for 9 and 294 sub-faults, respectively and Ulrich et al. (2019), who modeled the supershear seabed deformation as a function of space and time. Note that the latter more advanced study predicted

a 1.5 m maximum vertical seabed motion.

Other coseismic source mechanisms derived from geodetic observations, yield larger vertical seabed motions and, therefore, should be more tsunamigenic (e.g., ~ 3 m just south of the Balaesang Peninsula, Fig. 3.1a, Song et al., 2019; Fang et al., 2019; He et al., 2019). These, however, are not within Palu Bay but farther north, hence, cannot explain the tsunami here, particularly the fast arrival of large waves that impacted Palu City (the timing of events and waves will be detailed later). Some argue (Ulrich et al., 2019) that the horizontal fault movement along the steep slope margins of Palu Bay resulted in an increase in vertical water displacement causing elevated runups, in the manner proposed by Tanioka and Satake (1996). The main challenge here, however, with the single earthquake mechanism, is that it cannot explain the timing of the tsunami impacts along the Bay from the coastal landslides reported in the survivor accounts and video evidence, on land and that captured by an aircraft pilot flying over the bay shortly after the earthquake happened (Fig. 3.2).

With regard to the landslide mechanism, Pakoksung et al. (2019); Nakata et al. (2020) and Sepúlveda et al. (2020) identified and modeled these as important if not the principal contributors to the tsunami. Their landslide parameters and corresponding wave generation, however, were hypothetical and selected to match observations. To date, only Liu et al. (2020) modeled the landslide tsunamis from mapped landslide locations, but they did not model the tsunami generation from the landslides directly, using semi-empirical sources, did not simulate the additional earthquake mechanism, and simulated tsunami propagation with a non-dispersive model.

Here, for the first time, we will demonstrate that to explain the tsunami observations in Palu Bay requires a combined coseismic and landslide mechanisms,

and that a model addressing dispersion is required to simulate the shorter wavelength landslide tsunamis propagating into the deeper waters in the center of Palu Bay. Tsunami simulations published so far for 2018 Palu, used non-dispersive models and did not account for the physics of landslide tsunami generation. Takagi et al. (2019) used a simplified numerical model of a dual earthquake/landslide mechanism with a landslide, located in the southwest of Palu Bay, mapped by high-resolution multibeam echosounder (MBES) and identified in the aircraft pilot video. Their model resulted in shorter period waves from the coastal landslide, followed by longer waves from the earthquake. Finally, whereas many published models are able to simulate some, or even most, recorded runups around the Bay, these are based on using “ad hoc” mechanisms and do not reproduce the timing of tsunami waves from eyewitness accounts or the video evidence.

To better understand the 2018 Palu tsunami mechanism, in this work we propose to: (i) use a physics-based numerical model of landslide tsunami generation from mapped (rather than hypothetical) coastal landslides, (ii) use an appropriate tsunami model that addresses the physics of deforming slide motion, corresponding wave generation, and wave dispersion, critical in landslide tsunami propagation; (iii) simultaneously model earthquake and landslide tsunamis, since both are important to explain observations and wave-wave interactions may play a role in shallower waters; (iv) in the latter, investigate a range of published earthquake mechanisms; and (v) apply a delay for triggering landslides in the model, inferred from the aircraft pilot and other videos. Additionally, our simulations use improved bathymetry, coastal landslide evidence, a more comprehensive database of post-tsunami field survey results, including tsunami elevations, inundation limits and timings, together with a novel analysis of the tsunami videos.

Specifically, in order to reproduce the nearshore wave processes to high res-

olution, we use higher-resolution model grids than previously, using earthquake mechanisms alone, and in combination with coastal landslides identified from field studies, bathymetric surveys, and video evidence. We use the highest resolution bathymetric and topographic data available, interpolated onto our high-resolution computational grids in Palu Bay (Fig. 3.1). Our numerical modeling utilizes a combination of two well-benchmarked and validated numerical models, the three-dimensional (3D) non-hydrostatic model NHWAVE, and the two-dimensional (2D) Boussinesq model FUNWAVE-TVD, to generate and/or propagate the tsunami waves. Both models incorporate the physics of wave dispersion as well as full nonlinearity. NHWAVE can simulate landslide tsunami generation by dense mud flows or granular material, the latter of which is used in this study. We validate numerical results from runup measured in post-tsunami field surveys, waveform time-series estimated by Carvajal et al. (2019)’s video analysis, and the Pantoloan tide gauge data.

In the following, in Section 2, we detail and analyze tsunami observations, present the modeling methodology and data used to define tsunami sources and bathymetry/topography in model grids, in Section 3, we present modeling results for earthquake only, landslide only, and combined earthquake/landslide tsunami, and in Section 4 we discuss the latter and offer conclusions and perspectives for future work.

3.2 Materials and Methods

3.2.1 Tsunami Observations

In the following, we define $t = 0$ as the start of the 2018 Palu event (10:02:45 AM UTC), i.e., the time the earthquake rupture begins at the epicenter (yellow star in Fig. 3.1a).

Tide gauges

Two operational tide gauges recorded apparent tsunami signals for the 2018 Palu event: (i) in Mamuju (-2.66° N, 118.89° E), in the Makassar Strait, on western Sulawesi about 250 km SSW from Palu Bay, a maximum trough-to-crest wave height of ~ 0.25 m was recorded, arriving $t = 19$ min; and (ii) in Pantoloan, within Palu Bay (-0.711° N, 119.856° E; site P in Fig. 3.1b), a maximum trough-to-crest wave height of ~ 3.8 m was recorded, arriving at $t = 5 - 6$ min (BIG, 2018, Fig. 3.4b). As noted by many modelers (e.g., Heidarzadeh et al., 2019), a tsunami wave traveling from the approximate location of the epicenter, north of Palu Bay (Fig. 3.1a) would take ~ 45 min to arrive to the Mamuju tide gauge location, indicating either a clock error or that the signal here was caused by some other local source. As we focus here on tsunami waves within Palu Bay, we do not use the Mamuju data nor try to explain this discrepancy. In Pantoloan, as discussed, e.g., by Carvajal et al. (2019), Sepúlveda et al. (2020), and Liu et al. (2020), the gauge is located inside a harbor basin protected by a slotted seawall, which are not represented in the available bathymetric data and hence in our model grids. The gauge sensor recorded surface elevation every second, however, that data was averaged over 30 s and then output every minute, which yields the time series plotted in Fig. 3.4b. Therefore, while the gauge accurately captures long waves, shorter waves such as from landslides are missed. This was confirmed by a closed circuit television (CCTV) video recording (video 11 in Carvajal et al. (2019)) overlooking the docks at Pantoloan Port, that showed shorter period waves arriving approximately 2 min before the gauge registers any tsunami wave activity. This suggests that the Pantoloan tide gauge data, while useful, needs to be considered with this limitation in mind. Analysis of the tide gauge record before and after the tsunami shows that the mean sea level at Pantoloan did not change

during the earthquake, indicating that large permanent vertical elevation changes at this location from earthquake deformation did not occur (Sepúlveda et al., 2020).

Video analysis overview

Carvajal et al. (2019) compiled and analyzed 41 amateur (taken with mobile phones) and CCTV videos that were taken around the bay during the earthquake and tsunami impact. Based on shore-based videos, they estimated surface elevation time series at six locations (Fig. 3.4a-f): Wani, Pantoloan, Dupa, Talise, the KN Hotel, and the Palu Grand Mall, all marked in Fig. 3.1b. This large video archive also showed ample evidence of tsunami generation by coastal landslides (see https://agsweb.ucsd.edu/tsunami/2018-09-28_palu/carvajal_2019_videos_palu/). Notably, a video of the western side of the bay was made by Batik Airways pilot Ricoseta Mafella, at approximately 10:04:33 UTC, i.e., $t = 108$ s (video 38 in the archive, aircraft location marked by a magenta diamond in Fig. 3.1b, at -0.829°N , 119.869°E ; Mafella, personal communication), based on which a composite picture was created in Fig. 3.2. The figure shows evidence of large tsunami generation, as sets of concentric waves, off of the locations of coastal landslides LS-B,C,D,E and F* (Fig. 3.1b). Two of the smaller sets of waves generated at locations marked “Boat” and “NBoat” are consistent with a video made from a small boat at location “Boat” (video 39 in the archive); this video also shows a failing coastal landslide, which was subaerial. Furthermore, a video taken from a ship dock at Taipa (marked in Fig. 3.1b; video 31 in the archive) showed at least one other landslide failure across the bay, to the north (potentially at location marked by LS-L or M in Fig. 3.1b).

Sunny et al. (2019) analyzed the waves generated on the western side of the bay at locations LS-D,E, and F* (Fig. 3.1b), seen in the pilot video 38,

Figure 3.4: Time series of surface elevation for 2018 Palu tsunami: (—) inferred from shore-based videos (Carvajal et al., 2019) in (a) Wani, (b) Pantoloan, (c) Dupa, (d) Talise, (e) KN Hotel, and (f) Grand Mall; and (b) (- - -) measured at the Pantoloan tide gauge (see Fig. 3.1b for locations), compared to our tsunami simulations of 3 coseismic sources by: (blue) Jamelot et al. (2019), (red) Socquet et al. (2019), and (green) Ulrich et al. (2019). Time t is measured from the start of the Palu earthquake event, on September 28, 2018 at 6:02:45 PM local time (10:02:45 AM UTC). Solid/dashed colored lines are FUNWAVE results (NHAVE for first 60 s with Ulrich et al. (2019)’s source) with dispersion turned on/off.

using Google Earth to match camera viewing angles and compare observed wave measurements to dimensions of known objects. They estimated the widths of the

sharp crescent waves to be 343 m and 461 m, and the heights (trough-to-crest) to be 24.1 m and 28.9 m at locations LS-D and -E, respectively (Fig. 3.1b). Based on the boat video 39, they estimated that at location F*, the splash of the outgoing wave was 28.4 m high, and the height of unbroken outgoing wave traveling towards the Boat location was 8.2 m. Note that these estimated wave heights are all much larger than the initial values predicted by Liu et al. (2020) for the landslide tsunami waves generated at these locations, using semi-empirical sources, which are on the order of 2-10 m.

Timing and wave sequence analyses based on videos

CCTV footage at a house in Wani, in the northern section of the bay (videos 7 and 8 in Carvajal et al. (2019); -0.6935 °N, 119.8417 °E; W in Fig. 3.1b) shows a timestamp of 10:02:54 UTC, or $t = 9$ s, when the ground begins shaking. Combining the supershear rupture speed of 4.81 km/s (Bao et al., 2019) and the distance of ~ 50 km from the epicenter, shaking should have started at $t \simeq 10.4$ s, which is consistent with the camera time stamp. In Ulrich et al. (2019)’s simulations of the earthquake, horizontal and vertical deformations begin at Wani at $t = 11$ s, which is also in agreement with the previous observation and travel time estimate based on a supershear rupture. These results indicate that supershear travel time can be used to estimate the time after rupture initiation that ground begun shaking at various locations around Palu bay.

According to Pilot Ricoseta Mafella’s flight log (personal communication), his aircraft, a large size passenger jet, began taking off at 10:02:40 UTC on the 2,500 m long runway 33 of Mutiara SIS Al-Jufrie Airport (PLW), located southeast of the Bay. In a social media post, the pilot wrote: “I felt something wrong on the runway during takeoff roll.” The airport is ~ 73.5 km from the epicenter, yielding

an estimated start time for ground motion of $t = 15.3$ s based on travel time; in Ulrich et al. (2019)’s simulations horizontal and vertical deformation at the airport start at $t = 15.5$ s. The aircraft reached a 1,000 ft altitude at 10:02:59 UTC or $t = 14$ s and was located at -0.904°N , 119.903°E , just beyond the runway. Seismic travel time estimates are thus consistent with the aircraft flight log, within a few seconds.

Based on the consistent travel time and modeling estimates for the start of ground motion, we conclude that all locations within the bay most likely started shaking at $t = 9 - 15.5$ s, which is important information for constraining tsunami wave arrival times based on videos that also show earthquake shaking. At Wani and Pantoloan, we thus included a 9 s delay, to allow for seismic waves to reach this area, and in the southern sites of Dupa, Talise, KN Hotel, and Grand Mall, we included a 14 s delay. As mentioned earlier, at 10:04:33 UTC, or $t = 108$ s, Pilot Mafella and his aircraft were located at -0.829°N , 119.869°E (near the eastern side of the Bay; Fig. 3.1b), the approximate location where the pilot started recording video 38, which showed widespread evidence of landslide tsunami generation on the west side of the Bay (Fig. 3.2).

Observations in some of Carvajal et al. (2019)’s archived videos are analyzed in the following and time series of surface elevation estimated by Carvajal et al. (2019) at various locations are plotted in Fig. 3.4:

- Videos 7 and 8 show a positive elevation wave impacting the house in Wani at $t = 223$ s; Carvajal et al. (2019) estimated a 5 m/s on-land inundation speed for this wave, which with the house located 150 m from the water puts the arrival time at the shoreline at $t \simeq 193$ s. Crew members on the Sabuk Nusantara vessel, which was docked at Wani (-0.6954°N , 119.8408°E), reported an immediate ~ 7 m withdrawal of the ocean at the shoreline,

followed 3 - 5 min (180-300 s) later by a ~ 15 m wave, whose crest elevation could be estimated at 8m (VOA-News, 2018). Fig. 3.4a shows the short time series of surface elevation estimated here.

- Video 11, taken in nearby Pantoloan Port by a CCTV camera looking towards a crane on the dock (-0.7106°N , 119.8552°E), captured the initial tsunami waves at this location. Assuming that the video footage begins at the time of shaking, the trough of the initial shoreline withdrawal occurs at $t = 189$ s, followed by a large positive wave at $t = 215$ s. Fig. 3.4b shows the time series of surface elevation estimated at this location (solid black line), which is consistent with waves inferred from the video recorded in Wani; but at the Pantoloan tide gauge, those shorter and higher waves were clearly not measured (dashed black line) or, more accurately, filtered out by the gauge.
- In videos 29-31, taken in Taipa (-0.7794°N , 119.8580°E ; Fig. 3.1b), the timing of the videos is unknown and the time series of surface elevation were not estimated. However, in video 29, Carvajal et al. (2019) note, there is a wave to the north across the bay that appears similar to other landslide generated waves located in other parts of the bay, which could potentially be attributed to sites LS-L or M (Fig. 3.1b).
- Video 14, at Dupa (-0.8204°N , 119.8811°E ; D in Fig. 3.1b) begins some time after the earthquake shaking. Carvajal et al. (2019) estimated that the withdrawal observed in the video started at $t = 105$ s and Fig. 3.4c shows the short time series of surface elevation estimated here.
- In video 13, at Talise (-0.8589°N , 119.8789°E ; T in Fig. 3.1b), water withdrawal starts at $t = 39$ s, followed by a large wave impacting the shore at $t \simeq 111$ s, as confirmed by the people transitioning from walking to run-

ning away from the coast. Fig. 3.4d shows the short time series of surface elevation estimated here.

- Six CCTV cameras were operated at the KN Hotel (-0.8650°N , 119.8775°E ; KN in Fig. 3.1b), ~ 750 m south of Talise. The camera timestamps were adjusted to the time shaking started, at $t \simeq 14$ s based on Ulrich et al. (2019). A withdrawal is not apparent, but tsunami inundation begins at $t = 106$ s from a northerly direction. In Carvajal et al. (2019)’s video 3, the camera angle from the KN Hotel points across the bay in the direction of the LS-F* landslide. In Fig. 3.5, one of the video frames, we see, in the upper right corner behind a tree, a disturbance becoming visible at $t = 52$ s (video time 38 s), which could potentially be the wave generated by the LS-F* landslide. Fig. 3.4e shows the short time series of surface elevation estimated here.
- Finally, many people took videos from various floors of a parking structure in Palu Grand Mall (video 43; -0.8836°N , 119.8437°E ; GM in Fig. 3.1b), which were combined into a 11’20’’ video referenced to the time of video 43 by Carvajal et al. (2019). While the exact start time is unclear, they estimated that the major impacts occurred 4-6 min after the main shock ($t = 240\text{-}360$ s). However, Takagi et al. (2019) analyzed other time-stamped videos from eyewitnesses here (see their Fig. 5) and indicate that the second positive wave estimated by Carvajal et al. (2019) hit Palu Grand Mall at 10:10:49 UTC, or $t = 484$ s. To reconcile this discrepancy, we shifted Carvajal et al. (2019)’s time series forward by 150 s. Fig. 3.4f shows the estimated shifted time series of surface elevation.

Post-tsunami surveys

Bathymetry: Results of bathymetric surveys carried out after the event have been published by Takagi et al. (2019), Frederik et al. (2019), and Liu et al. (2020), who compared them with various sources of pre-event bathymetric data (see details in references). Takagi et al. (2019) surveyed a few square kilometers offshore of the Buluri landslide site LS-F* (Figs. 3.1b and 3.2). This data is available at http://www.ide.titech.ac.jp/~takagi/file/2014_bathymetry_Fig3a_in_paper_Landslides.dat and http://www.ide.titech.ac.jp/~takagi/file/2018_bathymetry_Fig3b_in_paper_Landslides.dat (the urls given in their paper wrongly included a period or space after the word “Fig”). They estimated an approximate landslide volume of $V_S = 3.2 \cdot 10^6 \text{ m}^3$. Frederik et al. (2019) surveyed areas deeper than 200 m within Palu Bay, as well as the area southwest of the Balaesang Peninsula. Their analysis of the bathymetric data within Palu Bay shows no clear signs of a fault trace nor any evidence that large deep water submarine mass failures occurred. Liu et al. (2020) also surveyed Palu Bay, including nearshore coastal waters; they provide evidence for 14 locations where coastal landslides occurred, and estimated slide parameters based on differences between their new and the Badan Informasi Geospasial’s (BIG; Geospatial Information Agency, Indonesia) pre-earthquake bathymetric contours. In this work, we specifically studied and modeled the landslides located where video evidence confirmed that wave generation occurred, i.e., at sites LS-B,-C,-D,-E,-F*,-L, and -M (Fig. 3.1b; Table 2). Note, as we used the landslide dimensions adapted from and volumes estimated by Liu et al. (2020) in our modeling, with the exception of landslide F* for which we used data provided by Takagi et al. (2019), for consistency, we used the same labeling scheme as in Liu et al. (2020).

Tsunami Coastal Impact: Post-tsunami surveys were conducted by various

Figure 3.5: Locations and video frames taken during tsunami impact by CCTV cameras at the KN Hotel (KN in (Fig. 3.1b)). (a) Trace of three camera view angles corresponding to Carvajal et al. (2019)’s videos: (blue) 1, (yellow) 2, and (red) 3. Cyan line marks seawall location, other green lines mark location of knee to chest high wall visible in videos. Yellow line pointing away from camera 3 corresponds to view shown in (b). (b) Images from video 3 at $t = 52$ and 59 s. Yellow line corresponds to that marked in (a). Red ellipses encircle a growing free surface disturbance across the bay in landslide LS-F* area (Fig. 3.1b).

international teams, in which flow depth and runup (Fig. 3.1b) were measured around Palu Bay. Fig. 3.3 shows runup values by: (i) Pribadi et al. (2018), 26 runups corrected to MSL, measured September 29 - October 6, 2018 and October 10-17, 2018; (ii) Putra et al. (2019), 6 runups referenced to MSL, measured October 8-18, 2018; (iii) Widiyanto et al. (2019), 28 runups corrected to MSL,

measured October 11-19, 2018; (iv) Mikami et al. (2019), 6 runups corrected to tide level during earthquake (+1 m MSL), measured October 27-31, 2018; and (v) Omira et al. (2019), 55 runups corrected to tide level during earthquake, measured November 7-11, 2018. Here, we compare model results to measured runups, however, flow depth measurements were reported by Arikawa et al. (2018); Cipta et al. (2018); Paulik et al. (2019); Syamsidik et al. (2019). Runups referenced to MSL were transformed to MSL +1 m, to reflect the tide elevation at the time of the tsunami.

Label	Lon (°E)	Lat (°N)	θ (°)	b (m)	w (m)	T (m)	V_S (10^6m^3)
LS-B	119.7890	-0.7554	0	340	380	32	1.44
LS-C	119.7927	-0.7647	345	405	440	36	2.26
LS-D	119.8082	-0.8002	0	410	1220	18	3.07
LS-E	119.8121	-0.8090	0	175	335	18	0.37
LS-F*	119.8240	-0.8411	TA	TA	TA	40	3.2
LS-L	119.8430	-0.7038	225	830	515	44	6.66
LS-M	119.8204	-0.6887	200	800	350	35	3.44

Table 3.2: Parameters used to model coastal landslides in NHWAVE at their estimated unfailed location (Fig. 3.1b): azimuth angle θ (from N), down-slope length b , cross-slope width w , and maximum thickness T (assuming an elliptical footprint b by w and a quasi-Gaussian shape for the slide), volume $V_S = 0.3508 bwT$. Lat./Lon. define slide center of mass initial location. (See Schambach et al. (2019)’s appendix for parameter definition and sketch.) Data was adapted from Liu et al. (2020), except LS-F* for which actual landslide geometry was used in model based on Takagi et al. (2019)’s survey (TA)

3.2.2 Materials and methods

We simulate the 2018 Palu tsunami generation and propagation by earthquake or landslide sources, and the two in combination, with a methodology similar to that used in recent work by the authors and collaborators for other dual earthquake/landslide sources (e.g., Tappin et al., 2014; Grilli et al., 2015, 2017, 2019; Schambach et al., 2019, 2020). Using the best available bathymetric/topographic

data, higher resolution computational grids than used in previous studies, we apply two state-of-the-art dispersive wave models: (i) the 3D non-hydrostatic wave model NHWAVE (Ma et al., 2012), with an underlying slide layer (Ma et al., 2015; Kirby et al., 2016), to simulate both the landslide motion and related tsunami generation, and (ii) the 2D fully nonlinear Boussinesq wave model FUNWAVE-TVD (Shi et al., 2012) to simulate the propagation to the far-field and the coast of the superposition of landslide and coseismic tsunamis, in nested grids of increasing resolution towards the shore. The bathymetric/topographic data and computational grids used as well as the models set-up and modeling methodology are detailed next.

Study area, computational grids, and bathymetric/topographic data

The study area encompasses Palu Bay (Fig. 3.1), which is approximately 30 km long by 7.25 km wide with depths reaching up to ~ 830 m (Fig. 3.6). Analyses of pre- and post-earthquake satellite images show either N-S or NNE-SSW strike-slip ground motion north of Wani, and NNW-SSE motion south of the Palu Grand Mall (e.g., Socquet et al., 2019; Valkaniotis et al., 2018), indicating that the rupture trace changed direction somewhere in the bay. However, as mentioned, the high-resolution bathymetric data of Frederik et al. (2019), and the separate bathymetric survey by Liu et al. (2020) do not show any evidence of a clear rupture trace in the bay. As noted by Liu et al. (2020), since the Palu-Koro fault has not been mapped underwater, earthquake models have adopted various assumptions regarding where the fault trace is located. Fig. 3.1 shows fault traces from Jamelot et al. (2019); Socquet et al. (2019); Ulrich et al. (2019), corresponding to their earthquake sources that are simulated in this work.

Three Cartesian computational grids are used in tsunami simulations (Fig.

Figure 3.6: (a) Base computational grid (BG; 30 m resolution Cartesian, center at $(-0.720^{\circ}\text{N}, 119.810^{\circ}\text{E})$; Fig. 3.1) with two higher-resolution nested grids (red boxes, EG and SG; 7.5 m resolution Cartesian, center at $(-0.705^{\circ}\text{N}, 119.838^{\circ}\text{E})$ and $(-0.850^{\circ}\text{N}, 119.845^{\circ}\text{E})$, respectively) used in tsunami simulations. (b) Footprints of grids EG and SG. Color scale and contours indicates topography (> 0)/bathymetry (< 0) in meter. Various labels are defined in Fig. 3.1b. Yellowed areas indicate failed slide areas estimated from field surveys (Liu et al., 2020; Takagi et al., 2019)

3.6; listed grid center coordinates are used as Mercator transverse geographic projection origin). The 30 m resolution base grid, BG covers the study area in Palu Bay (Fig. 3.1) and is used in FUNWAVE and NHWAVE simulations; the NHWAVE BG grid also includes 5 boundary fitted, equally-spaced, layers in the vertical direction (from ocean surface to seabed). Two 7.5 m resolution grids are nested within BG, and used in FUNWAVE to more accurately model tsunami coastal impact in two areas of particular interest: (i) in the south of the bay, south grid SG includes the observation points of Grand Mall, KN Hotel, Taipa and Dupa; and (ii) near

Figure 3.7: Indonesian Geospatial Information Agency (BIG) 6 arc-sec bathymetric BATNAS dataset coastline (white line) plotted in the area of grid SG, with overlaid georeferenced satellite image from Google EarthTM, showing the discrepancy between the BATNAS coastline and the actual coastline. Red line shows the coastline inferred from the combination of the DEMNAS and BIG14 datasets, in good agreement with the actual coastline.

and around Pantoloan, east grid EG includes the observation points of Wani and Pantoloan (Fig. 3.1b).

A variety of bathymetric and topographic data sets have been used in earlier modeling studies of this event. Here we similarly combine and interpolate onto our grids the best available bathymetric and topographic data to date for our study area. The resulting bathymetry and topography are shown in Fig. 3.6. As an overall coarser data set, the earlier study of Heidarzadeh et al. (2019) used the General Bathymetric Chart of the Oceans 2014 (GEBCO14; Weatherall et al., 2015), which has a horizontal resolution of 30 arc-sec (~ 900 m), referenced to the local MSL. GEBCO data, however, gives a maximum depth of ~ 300 m in the center of Palu Bay, which is largely in error as this depth is greater than 800 m in more accurate data sets (Fig. 3.6); hence using GEBCO data may have caused errors in this earlier modeling study. The Indonesian Geospatial Information Agency (BIG)

provides a national bathymetric dataset, referred to as BATNAS, which has a horizontal resolution of 6 arc-sec (~ 180 m) and is referenced to MSL. When comparing BATNAS to geo-referenced satellite images from Google EarthTM, however, it was observed that the coastline was not accurately located (Fig. 3.7). BIG also provides bathymetric contours measured in Palu Bay during 2014, 2015 and 2017 surveys (referred to as BIG14, BIG15, and BIG17), all referenced to the lowest astronomical tide. These data sets are shown and discussed in detail by Liu et al. (2020), who point out that BIG17 is the most detailed, but only covers a small portion of the northern section of the bay. They also note that BIG15 is lower resolution and has a few anomalies compared to BIG14, concluding that BIG14 should be used to cover the areas of the bay not covered by BIG17. We proceeded similarly in our work. Regarding topographic data, BIG’s national topographic digital elevation model, referred to as DEMNAS, has a horizontal resolution of 0.27 arc-sec (~ 8.3 m) and a vertical datum EGM2008. In contrast to the BATNAS data set, the DEMNAS topography data set is fully consistent with the BIG14 bathymetric contours and satellite images, with both agreeing well at the coast (Fig. 3.7). Finally, as discussed earlier, three post-tsunami bathymetric surveys were reported by Takagi et al. (2019); Frederik et al. (2019) and Liu et al. (2020). In this work, we had access to and used the surveys of Takagi et al. (2019) and Frederik et al. (2019), whose reference vertical datum was MSL.

Thus, to generate our computational grid data, we interpolated the deepwater bathymetry of Frederik et al. (2019) with the shallow water bathymetry from the BIG14 contours, and the topography from DEMNAS, after referencing them all to the same MSL + 1 m vertical datum. To avoid numerical instabilities caused by slight discontinuities in bathymetry from combining different datasets, we applied a 2D Gaussian smoothing filter with a standard deviation of 1. The resulting

bathymetry and topography are shown in Fig. 3.6.

Numerical models and tsunami modeling methodology

Landslide tsunamis Landslide tsunami generation and initial propagation are modeled with the 3D non-hydrostatic model NHWAVE. This model was initially developed to simulate wave generation (including from the motion of rigid slides) and propagation (Ma et al., 2012) and later extended to simulate the motion and wave generation caused by deforming submarine or subaerial slides represented by a bottom layer made of a dense fluid (Kirby et al., 2016) or granular material (Ma et al., 2015). In its basic form, NHWAVE solves Euler equations in the water, in a horizontal Cartesian grid (x, y) with boundary fitted σ -layers specified from the seafloor to the free surface. With one σ -layers, NHWAVE approximately achieves the same level of dispersive properties as a Boussinesq model such as FUNWAVE (Shi et al., 2012). A larger number of layers allows accurately modeling wave dispersion in larger depth to wavelength ratios (see details in Ma et al. (2012)).

In the bottom slide layer, equations of mass and momentum conservation are depth-integrated. While initially vertical accelerations were neglected (Kirby et al., 2016; Ma et al., 2015), in the more recent implementation of NHWAVE used here, these equations also include vertical acceleration effects (Zhang et al., 2020a,b), which were found to be important on steep slopes (Grilli et al., 2019). Both volumetric (i.e., viscous/granular friction) and bottom friction (i.e., basal shear) dissipation terms are included in the slide layer equations. Equations of motion of the slide and water layers are coupled along the slide-water interface through kinematic and dynamic conditions. The model is parallelized using MPI to efficiently run on massively parallel computer clusters.

NHWAVE was extensively validated for a variety of tsunami benchmarks

(Zhang et al., 2017), including laboratory experiments performed by some of the authors (Grilli et al., 2017). The model was also used to simulate historical case studies, for which tsunami coastal impact had been measured (Tappin et al., 2014; Grilli et al., 2019; Schambach et al., 2020). In the latter cases, the initial unfailed landslide geometry is first recreated by moving the failed landslide material upslope. The model then simulates both the down-slope motion of the failing slide, coupled to that of the overlying water. For all benchmarks or actual events, NHWAVE was found to perform well and to adequately reproduce the reference data, provided the discretization was sufficient.

In the present simulations, except for slide LS-F* for which we use the actual mapped slide geometry, as in earlier work (e.g., Enet and Grilli, 2007; Grilli et al., 2015; Schambach et al., 2019), the initial slide geometry is modeled at the unfailed location as a sediment mound of quasi-Gaussian cross-sections, with maximum thickness T , and an elliptical footprint of down-slope length b and cross-slope width w ; with these definitions, the slide volume is calculated as, $V_S = 0.3508 bwT$ (see Appendix in Schambach et al. (2019), for details). Table 2 gives the geometric parameters and initial location estimated for each modeled landslide, based on Liu et al. (2020) (LS-B,C,D,E,L,M) or Takagi et al. (2019) (LS-F*) (Fig. 3.1b). In simulations, the initial geometry of each landslide is carved out of the pre-failed bathymetry BIG14, gridded in NHWAVE’s 30 m grid BG, at each slide estimated location (Table 3; Fig 3.6). Since the post-failure coastal bathymetry did not show clear slide deposits, no material was removed from the downslope bathymetry prior to simulations. On the basis of various field observations made in Palu (e.g. Liu et al., 2020), we assumed that once they failed the considered landslides all behaved as a granular material (i.e., a debris flow). This was represented in NHWAVE by a deforming bottom layer with a $\rho_S = 2,050 \text{ kg/m}^3$ density for the granular material,

a 60% solid volume fraction, a $\rho_w = 1,025 \text{ kg/m}^3$ water density, an internal friction angle $\phi_i = 30 \text{ deg}$, and a basal (bed) friction angle $\phi_b = 2 \text{ deg}$. The same granular density and internal friction values were used by Nakata et al. (2020) and the basal friction value is at the lower end of their tested range (2-6 deg); it is also identical to that used by Grilli et al. (2019) to model the 2018 Anak Krakatau flank collapse.

Based on the shear wave travel time from the earthquake epicenter to Palu Bay discussed above (on the order of 10-15 s), all the landslides should have been affected by ground shaking within seconds of each other; hence, they are assumed to fail at the same time in the model. However, there was a delay between the first instance of ground shaking due to seismic waves and the actual landslide initiation of motion, likely because complete failure required a sufficient built-up of excess pore pressure (and perhaps some liquefaction) in the submerged toe of the slide material (e.g., Tappin et al., 2008). This delay was estimated to $t_S \simeq 75 \text{ s}$ based on the aircraft pilot video and using NHWAVE simulations to estimate how much time was required to achieve the observed wave generation. Fig. 3.2 shows the state of wave generation at $t = 108 \text{ s}$ and, modeling the largest slides (in particular LS-F*), we found that it took 30-35 s of wave generation to qualitatively achieve the same stage as observed in the video, hence on average $t_S \simeq 108 - 33 = 75 \text{ s}$.

On this basis, landslide tsunami generation and their initial propagation up to $t = t_f = 150 \text{ s}$ was simulated in grid BG with NHWAVE, simultaneously for all the considered slides (Table 2; an animation of this simulation *slides.mp4* is provided in supplementary material). Results showed that, at this time, slides are no longer tsunamigenic and maximum landslide tsunami runups have occurred onshore of each slide location and outgoing.

For $t > t_f$, simulations are continued in FUNWAVE for landslide tsunamis alone or in combination with coseismic tsunamis, based on NHWAVE results for

surface elevation and horizontal velocity (interpolated at 0.531 times the local depth for consistency with FUNWAVE). This is detailed later.

Coseismic or dual landslide/coseismic tsunamis Three different coseismic tsunami sources are simulated in grid BG for the 2018 Palu event. Two are modeled for $t \geq 0$ (Jamelot et al., 2019; Socquet et al., 2019), with the 2D fully nonlinear and dispersive Boussinesq model FUNWAVE (Shi et al., 2012), based on specifying an initial static surface elevation, and one (Ulrich et al., 2019) using NHWAVE for $t \leq 60$ s based on directly specifying the bottom deformation in time and space, and then with FUNWAVE for $t > 60$ s (see details of coseismic sources later). FUNWAVE has been extensively validated in various wave propagation and coastal inundation studies (e.g. Shi et al., 2012), including for tsunami inundation/runup and coastal velocity benchmarks (??), and applied to tsunami case studies, both historical and hypothetical (e.g. Tappin et al., 2014; Grilli et al., 2015; Schambach et al., 2019; Grilli et al., 2019; Schambach et al., 2020).

For the dual earthquake/landslide sources, NHWAVE results are linearly superimposed at $t = t_f$ with those of FUNWAVE for the simulation of each coseismic source (i.e., surface elevation and horizontal velocity). Simulations of the combined tsunamis are then continued in FUNWAVE for $t > t_f$. For all cases simulated here, landslide or coseismic tsunamis alone, or dual sources, FUNWAVE simulations are performed by one-way coupling in the 2-level nested Cartesian grids (Table 3; Fig 3.6) BG (30 m resolution) and EG/SG (7.5 m resolution; see the earlier studies for details). To prevent reflection at open boundaries, 1800/4200 m wide sponge layers are specified along the western/northern boundaries of grid BG. Inundation and runup are modeled along coastal boundaries by way of a moving shoreline algorithm, with dissipation by wave breaking and bottom friction being simulated in FUNWAVE (Shi et al., 2012); here, a bottom friction coefficient $C_d = 0.0025$ is

used, which corresponds to coarse sand (also used in NHWAVE).

As discussed in the introduction, all published studies of 2018 Palu used non-dispersive tsunami propagation models. However, earlier work has shown the importance of frequency dispersion when modeling landslide tsunami generation and propagation, particularly when the initial slide footprint, and hence initial wavelength, are small compared to depth (e.g., Ma et al., 2012; Schambach et al., 2019). Here, we use dispersive models (NHWAVE and FUNWAVE) and, to assess the importance of dispersive effects, some simulations are run by turning off dispersion terms in the models (results are detailed later). In Palu Bay, while landslide tsunami waves generated in very shallow water may not initially be significantly dispersive, dispersion would become larger once waves propagated into the much deeper water of the center of the bay. Dispersion would then affect both phase speed and wave-wave interactions during propagation and, ultimately, tsunami coastal impact. Additionally, close to shore, dispersion can create undular bores (a.k.a. dispersive shock waves) near the crest and trough of longer shoaling tsunami waves, enhancing coastal impact (e.g., Madsen et al., 2008). This was demonstrated by Schambach et al. (2019) who also showed that higher resolution nearshore grids must be used to capture undular bores in FUNWAVE, which is one of the reasons here for using the finer grids EG and SG, even though the bathymetric data was not available at that level of detail; but, wave physics may call for it.

Earthquake source models

As discussed in the introduction, there is still no consensus on the 2018 Palu earthquake parameters, which fault(s) was(were) responsible, and how the rupture proceeded. For this reason, we selected and modeled 3 earthquake sources,

Figure 3.8: Maximum seabed uplift/subsidence computed for 2018 Palu Mw 7.5 coseismic sources, (a,b) with Okada (1985)’s method or (c) from instantaneous deformation computed through space-time modeling, all corrected to include horizontal motion effects for steep bottom slopes (Tanioka and Satake, 1996): (a) Jamelot et al. (2019), (b) Socquet et al. (2019), and (c) Ulrich et al. (2019). For sources (a,b), seabed motions are specified in FUNWAVE as static surface elevations at $t = 15$ s, while for source (c), simulations are performed with NHWAVE up to $t = 60$ s, before continuing in FUNWAVE; for comparison with other sources, (c) shows the solution at $t = 15$ s. See Fig. 3.1b for definition of location labels.

which are representative of studies made and will allow assessing the epistemic uncertainty in tsunami simulations related to the selection of an earthquake source model. The source by Jamelot et al. (2019) consists of 9 subfaults, with length and strike set to match onland displacements derived from correlating pre- and post-earthquake Sentinel-2 satellite images. A sensitivity study was performed with Okada (1985)’s half-plane integral equation model for various fault slip, width, dip, rake and depth values, to find the best fit with the observed displacements along each subfault, with the solution constrained by the centroid moment tensor. Socquet et al. (2019)’s earthquake source is more complex, consisting of 294 subfaults, each discretized by 42 elements in the strike direction and 7 elements in the dip direction, with parameters inferred from a joint inversion of Sentinel-2 and

Landsat-8 satellite images, as well as SAR interferograms from ALOS-2 satellite data. For these 2 sources, we computed the maximum seafloor deformation with Okada (1985)’s model and used it as initial surface elevation in FUNWAVE (with a zero initial velocity). As some studies indicated that effects of steep shores might have been important, we computed the additional vertical displacements due to horizontal displacements using Tanioka and Satake (1996)’s method. Assuming a supershear rupture, we specified the initial time for these surface elevations in FUNWAVE to $t = 15$ s into the event. Figs. 3.8a and b show the initial elevations computed for these 2 sources, which clearly are aligned along different fault traces (Fig. 3.1), but both predict a positive initial elevation (seafloor uplift) on the east side of the Bay and a negative initial elevation (seabed subsidence) on the west side of the Bay. Jamelot et al. (2019)’s source was designed by the authors to cause larger elevations both in the area of the Pantoloan tide gauge and Wani to the North, and in the area of Grand Mall and the KN Hotel to the south of the Bay in Palu City, where large tsunami impact was observed. Accordingly, in Figs. 3.8a, we see that the largest initial elevations occur in those areas.

The third earthquake source by Ulrich et al. (2019) was developed by modeling the earthquake failure in space and time with Seisol, an open source model solving the elastodynamic wave equations for spontaneous dynamic ruptures and seismic wave propagation (Dumbser and Käser, 2006; Pelties et al., 2014; Uphoff et al., 2017). Model inputs included geometry and frictional strength of the fault, tectonic stress state, and regional lithological structure, which were determined from datasets specific to the Palu region. Based on the authors’ results for the horizontal and vertical ground motions (provided every 0.5 for 50 s over a dense grid), we created time-series of seabed motion which were used as bottom boundary conditions over grid BG in NHWAVE. As with the other sources, contributions

to vertical displacement due to the horizontal movement of steep slopes were included. NHWAVE was used to simulate tsunami generation/propagation for this earthquake source up to $t = 60$ s and then simulations were continued in FUNWAVE for $t > 60$ s. To compare results with the other 2 sources, the surface elevation computed with NHWAVE is plotted at $t = 15$ s in Fig. 3.8c, where, we see that while the deformation is aligned along a fault trace similar to that of Socquet et al. (2019), the polarity of deformation is opposite, i.e., there are large negative elevations (subsidence) on the east side of the Bay and large positive elevations to the SW and NE of the Bay.

Figure 3.9: (a,c) Runups R simulated with NHWAVE/FUNWAVE for landslide sources only (Table 2), compared with field measurements (see Fig. 3.3 for definitions). Landslide tsunami generation is first simulated with NHWAVE in grid BG (figure footprint) up to $t_f = 150$ s, assuming all slides are triggered at $t_S = 75$ s. At this time, NHWAVE results (surface elevation and horizontal velocity a 0.531 times the local depth) are passed onto FUNWAVE to continue simulations for $t > t_f$, in grid BG and then in nested grids EG/SG (white footprints in figure (b)). Solid lines indicate results in 30 m grid BG and dashed lines in 7.5 m grids EG/SG. (b) Maximum surface elevations computed during landslide tsunami simulations.

3.3 Results

Simulations were performed with NHWAVE and FUNWAVE following the methodology detailed above, by considering first each of the 3 coseismic sources (Fig. 3.8), then the seven parameterized landslide sources (Table 2), and finally each coseismic source in combination with all the landslide sources (dual earthquake/landslide sources). Simulations for earthquake or landslide sources alone were performed with or without dispersive effects in the models. All simulations were performed up to $t = 1,200$ s, which was determined to be long enough for maximum runup to be achieved along the Palu Bay shores.

Similar results were computed for each type of simulation, i.e.: (i) the envelope of maximum surface elevations over the study area and runups along the Palu Bay coastline, to be compared with measurements from field surveys (Mikami et al., 2019; Omira et al., 2019; Pribadi et al., 2018; Putra et al., 2019; Widiyanto et al., 2019), in Figs. 3.3, 3.9, and 3.11, respectively (runups computed in both the coarser BG and finer SG/EG grids are provided, whereas envelopes are computed using the finer resolution results, wherever available); and (ii) time series of surface elevations computed at locations where various observations or measurements were made or inferred, i.e., Wani, Pantoloan, Dupa, Talise, KN Hotel, and Grand Mall (Fig. 3.1b), in Figs. 3.4, 3.10, and 3.12, respectively.

Animations of tsunami propagation simulations for: (1) coseismic sources alone (*Jamelot.mp4*, *Soquet.mp4*, and *Ulrich.mp4*); (2) landslide sources alone (*slides.mp4*); and (3) dual coseismic/landslide sources (*dual.mp4*) are given in supplementary material, together with an animation of the slide LS-F* and its corresponding wave generation (*slide-LS-F*.mp4*.)

First, regarding the effects of grid resolution, Figs. 3.3, 3.9, and 3.11 show that, for all types of sources, the largest runups simulated in the finer grids

Figure 3.10: Similar results as in Fig. 3.4 (same vertical scale kept in figures for comparison), but for simulations with NHWAVE/FUNWAVE of landslide sources only (Table 2). Simulation results include a landslide trigger delay of $t = 75$ s. Solid/dashed colored lines are results with dispersion turned on/off.

(EG/SG) are larger than in the coarser grid (BG), which justifies using nested grids in FUNWAVE. Then, regarding the effects of dispersion, time series of surface elevation in Fig. 3.4 show that, in coseismic sources simulation, only small absolute differences occur at most times and locations, whether frequency disper-

sion is turned on or off in the models. This could have been expected for the longer waves of coseismic tsunamis, but relative differences between the two simulations are nevertheless as large as 25-35% for some times/locations. In contrast, in landslide source simulations, time series in Fig. 3.10 show that, at most times/locations, much larger absolute or relative differences in surface elevations, and larger phase shifts, occur between the dispersive and non-dispersive results. This indicates significant dispersive effects on phase celerity and again is expected, based on earlier work, for the shorter waves of landslide tsunami, particularly considering the large depth of the bay. These results justify using dispersive wave models in this work.

When simulating coseismic sources alone, the observed runups are significantly underpredicted by the 3 sources in the southern part of Palu Bay (Fig. 3.3; south of ~ -0.75 Lat. N), with a maximum of 5 m whereas observed runups reached up to 10.5 m. In contrast, in the northern part of the Bay, runups are relatively well predicted by Jamelot et al. (2019)'s and Ulrich et al. (2019)'s coseismic sources, with a slight advantage for the latter. This agrees with earlier studies, showing that coseismic sources alone could not explain the tsunami coastal impact (e.g. Sepúlveda et al., 2020; Nakata et al., 2020), particularly in the south. Consistent with these results, Fig. 3.4 shows that measured or inferred time series of surface elevation at many locations are not well reproduced by the coseismic source simulations, particularly in the southern part of the Bay. Exceptions are Wani, the northern location (Fig. 3.1b) where results of the Jamelot et al. (2019) source agree well with the partial reconstruction based on a video recording, and Talise in the SE where results for Ulrich et al. (2019)'s source partly agree with the reconstructed surface elevation.

When only landslide tsunami sources are modeled, the observed runups are well predicted in the SW part of Palu Bay (Fig. 3.9), particularly in the area

Figure 3.11: Runups R simulated with NHWAVE/FUNWAVE for dual earthquake/landslide (Table 2) sources, compared with field measurements (see Fig. 3.3 for definitions of data, 3 coseismic sources, line colors and types). Landslide tsunami generation is first simulated with NHWAVE in grid BG (figure footprint) up to $t_f = 150s$. At this time, NHWAVE results are linearly combined (surface elevation and velocity at 0.531 times the local depth) with those computed with FUNWAVE in grid BG for each of the 3 coseismic sources. Simulations are then continued with FUNWAVE for $t > t_f$, in grid BG and then nested EG/SG grids (white footprints in figures (b,d,e)).

of the largest landslide sources (LS-E, LS-F*). However, a few of the largest observed runups are still underpredicted in the area of Dupa on the SE of the Bay (around ~ -0.85 Lat. N). In the northern part of the bay, on the western side,

Figure 3.12: Similar results as in Figs. 3.4 and 3.10 (same vertical scale kept in figures for comparison), for simulations with NHWAVE/FUNWAVE of combined (dual) coseismic/landslide (Table 2) sources. See Fig. 3.4 caption for definition of coseismic sources. All results are computed here with dispersion turned on.

observed runups are nearly as well predicted as for coseismic sources, but because of the timing of the event, maximum runups caused by coseismic or landslide sources would have likely occurred at different times here (see animations of model results). Hence, it is difficult to identify their primary source, which may explain the mitigated conclusions or even confusion in some earlier studies. On the NE side

of the Bay, around Wani and Pantoloan, the landslide tsunami impact is predicted to be quite large and explains the large runups observed better than for coseismic sources. This is confirmed in time series of surface elevation (Fig. 3.10), where there is a much better agreement in Wani and Pantaloan of model results with the reconstituted time series. In the SE of the Bay, however, consistent with the underpredicted runups, the landslide tsunami simulations do not explain well the time series reconstructed in Dupa, Talise, and KN Hotel. Finally, at Grand Mall in the south, while the shorter period landslide tsunami waves agree better in timing with those of the reconstructed time series, their amplitude is still underpredicted, despite using a very fine model grid that could have enhanced wave shoaling.

Results of the dual earthquake/landslide source simulations (Fig. 3.11) are consistent with the above observations. For all earthquake sources, but particularly for the combination of Ulrich et al. (2019)'s with the landslide sources, the observed runups on the entire west side of the bay are well reproduced in the simulations at most locations, and this is also the case on the east side of the bay, except essentially for the area around Dupa (around ~ -0.85 Lat. N) where another local source of waves would be needed, perhaps from another landslide not yet identified in the SE of the Bay. Other studies, have pointed out this discrepancy and this need (e.g. Liu et al., 2020). In the time series of Fig. 3.12, we see a good agreement of the dual source simulations with the reconstructed time series in Wani and Pantoloan, particularly when using Ulrich et al. (2019)'s or Socquet et al. (2019)'s source with the landslides. A reasonable agreement is also found in the SE of the Bay, in Dupa and Talise, when combining the landslide sources with Ulrich et al. (2019)'s source. At the KN Hotel, however, in the same area, none of the simulations agree well with the short reconstructed time series. Finally, at Grand Mall, in view of the uncertainty (and fairly arbitrary manner) of reconstructing the observed time

series, one could argue that combining the landslide sources with Ulrich et al. (2019)’s coseismic source also provides a reasonable agreement with observations, at least in amplitude, and more or less in phase.

3.4 Discussion

In this work, unlike earlier studies, our goal was to model the 2018 Palu event for a combination of published earthquake sources and observed (rather than hypothetical) landslide sources (i.e., measured in bathymetric surveys and confirmed by video evidence to have generated significant waves), using an actual landslide tsunami model for the latter. Dispersive models were used to simulate tsunami propagation, which was found particularly important for the shorter landslide tsunami waves.

Published earthquake models for the 2018 Palu event predict different vertical deformation, so we selected and simulated 3 different models of varying complexity to estimate the epistemic uncertainty associated with modeling the coseismic tsunami (Jamelot et al., 2019; Socquet et al., 2019; Ulrich et al., 2019). While the initial surface elevation caused by each coseismic source is quite different (Fig. 3.8), the generated tsunamis can all explain runups observed in the northern section of the bay, but they all underpredict the larger runups in the southern part of the bay. On the southwest side, large runups occurred just onshore of confirmed coastal landslides, indicating the importance of simulating landslide tsunamis for this event. Without a comparison of pre- and post- earthquake leveling data, it is not clear which, if any, of the earthquake models is most appropriate.

Liu et al. (2020)’s shallow water bathymetric survey identified multiple coastal landslides, which video evidence (Carvajal et al., 2019) showed to have generated large waves (e.g. Fig. 3.2), and estimated cross-/along-shore extents and volumes of landslides (LS-B,C,D,E,L,M; Table 2), which allows for the first time model-

ing their associated wave generation. Detailed surveying of slide LS-F* by Takagi et al. (2019), where large wave generation was observed (see Fig. 3.2), provided additional information allowing to model landslide tsunami generation in this location. Both motion and wave generation by these 7 landslides were simulated using the state-of-the-art 3D non-hydrostatic wave model NHWAVE, with an underlying layer of granular slide material. Once waves were fully generated, results were passed into the 2D dispersive long wave model FUNWAVE-TVD, for simulation tsunami propagation and coastal impact. A landslide trigger delay of 75 s from the start of the earthquake was estimated based on the pilot video as well as time series of surface elevation at Wani and Pantoloan (Fig. 3.10 a,b). The landslides alone are found to be capable of generating runups on the same order as those observed onshore of their locations, except on the southeast side of the bay where runups are still underpredicted in the Dupa area (simulated 2-4 m, vs observed 8-10.5 m). Compared to the reconstructed time series, good agreement is obtained in Wani and Pantoloan, but not in locations in the southeast of the bay. At Grand Mall, wave arrival matches what was observed, however, the amplitudes are not as large.

Our results of dual earthquake/landslide source simulations are all in general agreement with observed runups, again with the exception of the Dupa to Talise area southeast of the bay. Results of the dual source based on Ulrich et al. (2019)'s coseismic source predict runups slightly better and agree well with observed/reconstructed time series, except at the KN Hotel that none of the results can explain. The time series at Grand Mall is well predicted in amplitude and decently in phase.

To explain the large runups observed SE of the bay, Nakata et al. (2020) modeled a large $700 \times 10^6 \text{ m}^3$ hypothetical underwater landslide off of Talise and obtained

Figure 3.13: (a,b) Maximum surface elevation and (c) Runups simulated for: (a and magenta in c) hypothetical SE landslide, and (b and green in c) dual Ulrich et al. (2019)'s earthquake source with all the slides in Table 2 plus SE slide sources, compared with field measurements (black bullets). (d-f) Time series of surface elevation computed for each case (same color coding), compared to reconstructed time series at: (d) Talise, (e) KN Hotel, (f) Grand Mall. (See Fig. 3.4 for definitions). Results are from 7.5 m resolution grid SG.

a good agreement with observations near this location. There is no indication on the seafloor that such a large failure occurred or would even be possible, but this work confirms that a SMF here could potentially explain the large runups near Talise and improve the agreement with time series at Talise and the KN Hotel. Based on our stated goal of only considering proven landslide sources, we did not include such a hypothetical slide in our modeling of dual sources. However, as

part of this discussion and to support the need for future marine geology work to identify this source, we modeled a landslide at a location similar to Nakata et al. (2020)'s (119.8675 Lon. E, -0.8540 Lat. N), but in part based on Liu et al. (2020)'s surveys, with a more realistic footprint ($b = 500$ m and $w = 1,000$ m), and thickness ($T = 150$ m), leading to a much smaller volume $V_S = 26.3 \cdot 10^6$ m³. As before, the slide was first simulated in NHWAVE and then propagated with FUNWAVE in grids BG and SG. Results in Fig. 3.13a,c show that generated waves focus on 2 areas onshore of the slide. One, south of the slide, is the area in between the KN Hotel (-0.866 Lat.) and where runup were reported (around -0.873 Lat.); here the slide causes a 3 m runup, consistent with these measurements. The second area is north of Talise (-0.876 Lat.), where the largest runups were measured (around -0.845 Lat.); Here, the slide causes a large runups ~ 5.9 m, albeit still lower than the 8-10.5 m that were reported. Figs. 3.13b,c shows results for a dual source resulting from the combination of Ulrich et al. (2019)'s earthquake source with all the slides in Table 2, plus the hypothetical SE slide sources. Some wave interferences slightly reduce the runup south of the KN Hotel but north of Talise, the combined runup is still nearly 6 m, whereas without the SE slide it was only 3 m (Fig. 3.11c).

Finally, Figs. 3.13d,e,f show time series of surface elevation computed at Talise, KN Hotel, and Grand Mall, respectively for the SE slide and for the complete dual source based on Ulrich et al. (2019). Compared to the earlier results in Fig. 3.12, we see that the new dual source provides an improved general agreement with the reconstructed time series; at the KN Hotel, in particular, only the inclusion of the SE slide able is able to explain the first elevation wave observed at $t \simeq 125$ s (here still underestimated but arriving at the right time).

The 2018 Palu tsunami was a complicated event, which this work suggests

consisted of dual sources of potentially equal importance, earthquake and coastal landslides. The importance of considering all available information to discern how the tsunami waves were generated is also shown, as the time series reconstruction from video evidence is valuable for distinguishing between models, rather than just runup. A proper understanding and modeling of such destructive dual source tsunami event can help mitigate tsunami coastal hazard resulting from future similar events, here or in other tsunami prone areas.

Acknowledgements

Support was provided to the University of Rhode Island authors from Grant CMMI-15-35568 from the Engineering for Natural Hazards Program, National Science Foundation. Numerical simulations reported in this work used HPC resources, as part of the Extreme Science and Engineering Discovery Environment (XSEDE) (project BCS-170006), which is supported by the National Science Foundation (NSF) grant number ACI-1548562.

Special thanks to Dr. Anne Socquet for providing their earthquake slip distribution model, and to Dr. Thomas Ulrich for providing the time series of horizontal and vertical ground motion from their earthquake model. DRT publishes with the permission of the CEO of the British Geological Survey United Kingdom Research and Innovation. Dr. Boma Kresning is acknowledged for helping to navigate and translate Indonesian data sources.

References

- Arikawa, T., Muhari, A., Okumura, Y., Dohi, Y., Afriyanto, B., Sujatmiko, K.A., Imamura, F., 2018. Coastal subsidence induced several tsunamis during the 2018 sulawesi earthquake. *Journal of Disaster Research* 13, 1–3.
- Bao, H., Ampuero, J.P., Meng, L., Fielding, E.J., Liang, C., Milliner, C.W., Feng, T., Huang, H., 2019. Early and persistent supershear rupture of the 2018 magnitude 7.5 palu earthquake. *Nature Geoscience* 12, 200–205.
- BIG, 2018. Real time tidal observation/ pengamatan pasang surut real time. tides.big.go.id (last accessed April 30, 2020).
- BNPB, 2019. Gempabumi dan tsunami sulawesi tengah. <https://bnpb.go.id/infografis/infografis-gempabumi-m74-tsunami-sulawesi-tengah> (last accessed May 5, 2020).
- Bradley, K., Mallick, R., Andikagumi, H., Hubbard, J., Meilianda, E., Switzer, A., Du, N., Brocard, G., Alfian, D., Benazir, B., et al., 2019. Earthquake-triggered 2018 palu valley landslides enabled by wet rice cultivation. *Nature Geoscience* 12, 935–939.
- Carvajal, M., Araya-Cornejo, C., Sepúlveda, I., Melnick, D., Haase, J.S., 2019. Nearly instantaneous tsunamis following the mw 7.5 2018 palu earthquake. *Geophysical Research Letters* 46, 5117–5126.
- Cipta, A., Omang, A., Supartoyo, Minarno, P.A., Solilkhin, A., Falah, F.N., Efendi, R., 2018. Anomali perilaku gelombang tsunami. Di balik pesona Palu , 133–141.

- Dumbser, M., Käser, M., 2006. An arbitrary high-order discontinuous galerkin method for elastic waves on unstructured meshes—ii. the three-dimensional isotropic case. *Geophysical Journal International* 167, 319–336.
- Enet, F., Grilli, S.T., 2007. Experimental study of tsunami generation by three-dimensional rigid underwater landslides. *J. Waterway, Port, Coastal, Ocean Eng.* 133(6), 442–454.
- Fang, J., Xu, C., Wen, Y., Wang, S., Xu, G., Zhao, Y., Yi, L., 2019. The 2018 mw 7.5 palu earthquake: A supershear rupture event constrained by insar and broadband regional seismograms. *Remote Sensing* 11, 1330.
- Frederik, M.C., Adhitama, R., Hananto, N.D., Sahabuddin, S., Irfan, M., Moefti, O., Putra, D.B., Riyalda, B.F., et al., 2019. First results of a bathymetric survey of palu bay, central sulawesi, indonesia following the tsunamigenic earthquake of 28 september 2018. *Pure and Applied Geophysics* 176, 3277–3290.
- Goda, K., Mori, N., Yasuda, T., Prasetyo, A., Muhammad, A., Tsujio, D., 2019. Cascading geological hazards and risks of the 2018 sulawesi indonesia earthquake and sensitivity analysis of tsunami inundation simulations. *Frontiers in Earth Science* 7, 261.
- Goto, C., Ogawa, Y., Shuto, N., Imamura, F., 1997. Numerical method of tsunami simulation with the leap-frog scheme. *IUGG/IOC TIME Project Intergovernmental Oceanographic Commission of UNESCO, Manuals and Guides* 35, 126.
- Grilli, S.T., O'Reilly, C., Harris, J.C., Bakhsh, T.T., Tehranirad, B., Banihashemi, S., Kirby, J.T., Baxter, C.D., Eggeling, T., Ma, G., et al., 2015. Modeling of smf tsunami hazard along the upper us east coast: detailed impact around ocean city, md. *Natural Hazards* 76, 705–746.

- Grilli, S.T., Shelby, M., Kimmoun, O., Dupont, G., Nicolsky, D., Ma, G., Kirby, J.T., Shi, F., 2017. Modeling coastal tsunami hazard from submarine mass failures: effect of slide rheology, experimental validation, and case studies off the us east coast. *Natural hazards* 86, 353–391.
- Grilli, S.T., Tappin, D.R., Carey, S., Watt, S.F., Ward, S.N., Grilli, A.R., Engwell, S.L., Zhang, C., Kirby, J.T., Schambach, L., et al., 2019. Modelling of the tsunami from the december 22, 2018 lateral collapse of anak Krakatau volcano in the Sunda Straits, Indonesia. *Scientific reports* 9, 1–13.
- Gusman, A.R., Supendi, P., Nugraha, A.D., Power, W., Latief, H., Sunendar, H., Widiyantoro, S., Wiyono, S.H., Hakim, A., Muhari, A., et al., 2019. Source model for the tsunami inside Palu Bay following the 2018 Palu earthquake, Indonesia. *Geophysical Research Letters* 46, 8721–8730.
- He, L., Feng, G., Li, Z., Feng, Z., Gao, H., Wu, X., 2019. Source parameters and slip distribution of the 2018 Mw 7.5 Palu, Indonesia earthquake estimated from space-based geodesy. *Tectonophysics* 772, 228216.
- Hébert, H., Heinrich, P., Schindel, F., Piatanesi, A., 2001. Far-field simulation of tsunami propagation in the Pacific Ocean: impact on the Marquesas Islands (French Polynesia). *Journal of Geophysical Research: Oceans* 106, 9161–9177.
- Heidarzadeh, M., Muhari, A., Wijanarto, A.B., 2019. Insights on the source of the 28 September 2018 Sulawesi tsunami, Indonesia based on spectral analyses and numerical simulations. *Pure and Applied Geophysics* 176, 25–43.
- Heinrich, P., Schindele, F., Guibourg, S., Ihml, P.F., 1998. Modeling of the February 1996 Peruvian tsunami. *Geophysical research letters* 25, 2687–2690.

- Jamelot, A., Gailler, A., Heinrich, P., Vallage, A., Champenois, J., 2019. Tsunami simulations of the sulawesi m w 7.5 event: Comparison of seismic sources issued from a tsunami warning context versus post-event finite source. *Pure and Applied Geophysics* 176, 3351–3376.
- Kirby, J.T., Shi, F., Nicolsky, D., Misra, S., 2016. The 27 april 1975 kitimat, british columbia, submarine landslide tsunami: a comparison of modeling approaches. *Landslides* 13, 1421–1434.
- Liu, P.F., Higuera, P., Husrin, S., Prasetya, G., Prihantono, J., Diastomo, H., Pryambodo, D., Susmoro, H., 2020. Coastal landslides in palu bay during 2018 sulawesi earthquake and tsunami. *Landslides* .
- Ma, G., Kirby, J.T., Hsu, T.J., Shi, F., 2015. A two-layer granular landslide model for tsunami wave generation: Theory and computation. *Ocean Modelling* 93, 40–55.
- Ma, G., Shi, F., Kirby, J.T., 2012. Shock-capturing non-hydrostatic model for fully dispersive surface wave processes. *Ocean Modelling* 43, 22–35.
- Madsen, P., Fuhrman, D., Schaffer, H., 2008. On the solitary wave paradigm for tsunamis. *J Geophys Res.* 113, C12012.
- Mikami, T., Shibayama, T., Esteban, M., Takabatake, T., Nakamura, R., Nishida, Y., Achiari, H., Marzuki, A.G., Marzuki, M.F.H., Stolle, J., et al., 2019. Field survey of the 2018 sulawesi tsunami: inundation and run-up heights and damage to coastal communities. *Pure and Applied Geophysics* 176, 3291–3304.
- Miyajima, M., Setiawan, H., Yoshida, M., Ono, Y., Kosa, K., Oktaviana, I.S., et al., 2019. Geotechnical damage in the 2018 sulawesi earthquake, indonesia. *Geoenvironmental Disasters* 6, 1–8.

- Muhari, A., Imamura, F., Arikawa, T., Hakim, A.R., Afriyanto, B., 2018. Solving the puzzle of the september 2018 palu, indonesia, tsunami mystery: clues from the tsunami waveform and the initial field survey data. *Journal of Disaster Research* 13, sc20181108.
- Nakata, K., Katsumata, A., Muhari, A., 2020. Submarine landslide source models consistent with multiple tsunami records of the 2018 palu tsunami, sulawesi, indonesia. *Earth, Planets and Space* 72, 1–16.
- Okada, Y., 1985. Surface deformation due to shear and tensile faults in a half-space. *Bulletin of the seismological society of America* 75, 1135–1154.
- Omira, R., Dogan, G., Hidayat, R., Husrin, S., Prasetya, G., Annunziato, A., Proietti, C., Probst, P., Paparo, M., Wronna, M., et al., 2019. The september 28th, 2018, tsunami in palu-sulawesi, indonesia: A post-event field survey. *Pure and Applied Geophysics* 176, 1379–1395.
- Pakoksung, K., Suppasri, A., Imamura, F., Athanasius, C., Omang, A., Muhari, A., 2019. Simulation of the submarine landslide tsunami on 28 september 2018 in palu bay, sulawesi island, indonesia, using a two-layer model. *Pure and Applied Geophysics* 176, 3323–3350.
- Paulik, R., Gusman, A., Williams, J.H., Pratama, G.M., Lin, S.I., Prawirabhakti, A., Sulendra, K., Zachari, M.Y., Fortuna, Z.E.D., Layuk, N.B.P., et al., 2019. Tsunami hazard and built environment damage observations from palu city after the september 28 2018 sulawesi earthquake and tsunami. *Pure and Applied Geophysics* 176, 3305–3321.
- Pelties, C., Gabriel, A.A., Ampuero, J.P., 2014. Verification of an ader-dg method

- for complex dynamic rupture problems. *Geoscientific Model Development* , 847–866.
- Prasetya, G., De Lange, W., Healy, T., 2001. The makassar strait tsunamigenic region, indonesia. *Natural Hazards* 24, 295–307.
- Pribadi, S., Gunawan, I., Nugraha, J., Haryono, T., Ermawan, Basri, C., Romadon, I., Mustarang, A., Heriyanto, Yatimantoro, T., 2018. Survey tsunami teluk palu 2018.
- Putra, P.S., Aswan, A., Maryunani, K.A., Yulianto, E., Kongko, W., 2019. Field survey of the 2018 sulawesi tsunami deposits. *Pure and Applied Geophysics* 176, 2203–2213.
- Sassa, S., Takagawa, T., 2019. Liquefied gravity flow-induced tsunami: first evidence and comparison from the 2018 indonesia sulawesi earthquake and tsunami disasters. *Landslides* 16, 195–200.
- Schambach, L., Grilli, S., Tappin, D., Gangemi, M., Barbaro, G., 2020. New simulations and understanding of the 1908 messina tsunami for a dual seismic and deep submarine mass failure source. *Marine Geology* 421, 106093.
- Schambach, L., Grilli, S.T., Kirby, J.T., Shi, F., 2019. Landslide tsunami hazard along the upper us east coast: effects of slide deformation, bottom friction, and frequency dispersion. *Pure and Applied Geophysics* 176, 3059–3098.
- Sepúlveda, I., Haase, J.S., Carvajal, M., Xu, X., Liu, P.L.F., 2020. Modeling the sources of the 2018 palu, indonesia, tsunami using videos from social media. *Journal of Geophysical Research: Solid Earth* .
- Shi, F., Kirby, J.T., Harris, J.C., Geiman, J.D., Grilli, S.T., 2012. A high-order

- adaptive time-stepping tvd solver for boussinesq modeling of breaking waves and coastal inundation. *Ocean Modelling* 43, 36–51.
- Socquet, A., Hollingsworth, J., Pathier, E., Bouchon, M., 2019. Evidence of supershear during the 2018 magnitude 7.5 palu earthquake from space geodesy. *Nature Geoscience* 12, 192–199.
- Song, X., Zhang, Y., Shan, X., Liu, Y., Gong, W., Qu, C., 2019. Geodetic observations of the 2018 mw 7.5 sulawesi earthquake and its implications for the kinematics of the palu fault. *Geophysical Research Letters* 46, 4212–4220.
- Sunny, R.C., Cheng, W., Horrillo, J., 2019. Video content analysis of the 2018 sulawesi tsunami, indonesia: Impact at palu bay. *Pure and Applied Geophysics* 176, 4127–4138.
- Syamsidik, Benazir, Umar, M., Margaglio, G., Fitrayansyah, A., 2019. Post-tsunami survey of the 28 september 2018 tsunami near palu bay in central sulawesi, indonesia: Impacts and challenges to coastal communities. *International Journal of Disaster Risk Reduction* 38, 101229.
- Takagi, H., Pratama, M.B., Kurobe, S., Esteban, M., Aránguiz, R., Ke, B., 2019. Analysis of generation and arrival time of landslide tsunami to palu city due to the 2018 sulawesi earthquake. *Landslides* 16, 983–991.
- Tanioka, Y., Satake, K., 1996. Tsunami generation by horizontal displacement of ocean bottom. *Geophysical Research Letters* 23, 861–864.
- Tappin, D., Watts, P., Grilli, S.T., 2008. The papua new guinea tsunami of 17 july 1998: anatomy of a catastrophic event. *Natural Hazards and Earth System Sciences* 8, 243–266.

- Tappin, D.R., Grilli, S.T., Harris, J.C., Geller, R.J., Masterlark, T., Kirby, J.T., Shi, F., Ma, G., Thingbaijam, K., Mai, P.M., 2014. Did a submarine landslide contribute to the 2011 tohoku tsunami? *Marine Geology* 357, 344–361.
- Ulrich, T., Vater, S., Madden, E.H., Behrens, J., van Dinther, Y., Van Zelst, I., Fielding, E.J., Liang, C., Gabriel, A.A., 2019. Coupled, physics-based modeling reveals earthquake displacements are critical to the 2018 palu, sulawesi tsunami. *Pure and Applied Geophysics* 176, 4069–4109.
- Uphoff, C., Rettenberger, S., Bader, M., Madden, E.H., Ulrich, T., Wollherr, S., Gabriel, A.A., 2017. Extreme scale multi-physics simulations of the tsunami-genic 2004 sumatra megathrust earthquake, in: *Proceedings of the international conference for high performance computing, networking, storage and analysis*, pp. 1–16.
- USGS, 2018. M 7.5 - 70 km n of palu, indonesia. <https://earthquake.usgs.gov/earthquakes/eventpage/us1000h3p4/executive> (last accessed April 30, 2020).
- Valkaniotis, S., Ganas, A., Tsironi, V., Barberopoulou, A., 2018. A preliminary report on the m7.5 palu 2018 earthquake co-seismic ruptures and landslides using image correlation techniques on optical satellite data. doi:10.5281/zenodo.1467128. report submitted to EMSC on 19 October 2018 12:00 UTC.
- VOA-News, 2018. Crew recounts riding tsunami that dumped ferry in village. <https://www.voanews.com/east-asia-pacific/crew-recounts-riding-tsunami-dumped-ferry-village> (last accessed June 30, 2020).
- Watkinson, I.M., Hall, R., 2017. Fault systems of the eastern indonesian triple

- junction: evaluation of quaternary activity and implications for seismic hazards. Geological Society, London, Special Publications 441, 71–120.
- Watkinson, I.M., Hall, R., 2019. Impact of communal irrigation on the 2018 palu earthquake-triggered landslides. *Nature Geoscience* 12, 940–945.
- Weatherall, P., Marks, K.M., Jakobsson, M., Schmitt, T., Tani, S., Arndt, J.E., Rovere, M., Chayes, D., Ferrini, V., Wigley, R., 2015. A new digital bathymetric model of the world’s oceans. *Earth and Space Science* 2, 331–345.
- Widiyanto, W., Santoso, P.B., Hsiao, S.C., Imananta, R.T., 2019. Post-event field survey of 28 september 2018 sulawesi earthquake and tsunami. *Nat. Hazards Earth Syst. Sci* 1, 1–23.
- Yolsal-Çevikbilen, S., Taymaz, T., 2019. Source characteristics of the 28 september 2018 m w 7.5 palu-sulawesi, indonesia (se asia) earthquake based on inversion of teleseismic bodywaves. *Pure and Applied Geophysics* 176, 4111–4126.
- Zhang, C., Kirby, J.T., Shi, F., Ma, G., Grilli, S.T., 2020a. A two-layer non-hydrostatic landslide model for tsunami generation on irregular bathymetry. 1. theoretical basis. *Ocean Modelling* (submitted).
- Zhang, C., Kirby, J.T., Shi, F., Ma, G., Grilli, S.T., 2020b. A two-layer non-hydrostatic landslide model for tsunami generation on irregular bathymetry. 2. numerical discretization and model validation. *Ocean Modelling* (submitted).
- Zhang, C., Tehranirad, B., Kirby, J.T., Derakhti, M., Nemati, F., Grilli, S.T., Ma, G., Shi, F., 2017. Tsunami Benchmark Results for the Non-Hydrostatic Wave Model NHWAVE, Version 2.0. Technical Report. University of Delaware, DE. URL: <http://personal.egr.uri.edu/grilli/zhang-et al-cacr-17-03.pdf>.