

Hydro- and morphodynamic tsunami simulations for western Greece compared with sedimentary field traces

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*When the sea turns its enormous power against us, our best defense is to get out of its way.
But to do that we must first be able to predict when and where the sea will strike.*

(PARKER 2010, p. 1)

Abstract

Oceans are an important resource for modern civilisation but at the same time entail high risks for coastal population. Major oceanic threats include storm surges and tsunamis, which have taken innumerable lives through history. Despite great advances in the scientific understanding of such wave phenomena in recent decades, coasts still may suffer heavy casualties in the case of extraordinary storm surges or tsunamis, in particular in developing countries. Whereas modern technology allows a precise and early forecast of storm surges, the prediction of tsunamis and associated inundation is a particularly demanding challenge. Similar to storm surges, tsunamis are dominated by the dynamic of long gravity waves. Generated by earthquakes, gravitational mass wastings, volcanic activity, atmospheric disturbances or cosmic impacts, tsunamis can cross entire oceans in the form of shallow water waves without major energy losses. Due to the large wavelength of up to several hundred kilometres, tsunamis strike the mainland with tremendous momentum resulting in considerable wave run-up and inland penetration and by this in a significant sedimentary and geomorphological response at the coast. Tsunami deposits or so-called tsunamites are characterised by a number of sedimentary features (e.g. basal erosional contacts, normal grading and rip-up clasts) reflecting the hydrodynamic processes during inundation. Tsunamites in the coastal geological record allow not only to derive the frequency of pre-/historical tsunamis in a specific area but also the calibration and validation of hydro- and morphodynamic numerical models as a valuable tool to study local tsunami risks. A basic wave theory describing the propagation of tsunamis in numerical models is the Boussinesq approximation, which adequately reflects the non-linear, dispersive wave behaviour of tsunamis.

Based on sedimentary field traces and hydro- and morphodynamic numerical simulations, this PhD thesis focuses on pre-/historical tsunamis in the eastern Mediterranean Sea in western Greece. Associated with the nearby subduction zone of the African and Eurasian Plate, the west coast of Greece is located in the most active seismo-tectonic and therefore the most tsunami-genic area of the Mediterranean. Palaeotsunamis in two marine inlets—the northern Gulf of Kyparissia (western Peloponnese) and the Ambrakian Gulf (coastal Epirus/Akarnania)—are investigated. For the first study area, different digital elevation models (DEM) are created, which represent the recent topography and the contemporary palaeotopography of prehistorical tsunami impacts indicated by field data. Considering the various palaeo-/DEMs and two different tsunami waveforms (solitary and leading-depression *N*-waves (LDN-waves)), the effects of both, different palaeo-/topographies and various hydrodynamic boundary conditions, on the onshore tsunami response in the numerical model are methodologically investigated. The results show that LDN-waves cause considerably stronger flooding in the study area than solitary waves with the same wave amplitude. This phenomenon, known as the *N*-wave effect, becomes stronger with increasing wave amplitude and is demonstrated for the first time in a specific study area. However, differences in the inundation dynamics are even more significant with regard to the various palaeo-/topographies. This relates to substantial vertical crust movements in the northern Gulf of Kyparissia known from the past, which are considered in the palaeo-DEMs. By applying the different boundary conditions and palaeo-/DEMs, the numerical model achieves close agreement with sedimentary field observations, which allows the reconstruction of local palaeotsunami inundation dynamics for three key locations in the northern Gulf of Kyparissia—the Katakolo Pass, the Epitalio Pass and the valley of Kato Samiko. The best match

between simulations and field data is found for those scenarios that are based on the first created palaeo-DEM considering a 13 m tectonic throw along the major fault zones of the area. In these scenarios, solitary or LDN-waves with amplitudes of between 3 m and 6 m are required to cause significant inundation of the three sites and therefore are assumed to have a sedimentary response in accordance with field observations.

For the second study area, the Ambrakian Gulf, 151 vibracores are analysed with regard to sedimentary tsunami signals in the geological record. Using a hydro- and morphodynamic numerical tsunami model, various LDN-waves are simulated in order to find scenarios that match the spatial distribution of detected tsunami deposits in the field. Both, field traces and simulations indicate a decreasing tsunami influence from the western to the eastern Ambrakian Gulf, although various scenarios are needed to reproduce field observations in different parts of the gulf, as is also implied by the field data itself. Tsunami deposits detected in the western gulf show a better agreement with the simulated short-period tsunami waves from a southern or western direction, whereas long-period tsunamis from the west favourably reproduce field observations in the more inner gulf. Tsunamis from a south-western direction generally do not match field data. LDN-wave amplitudes of between 4 m and 6 m are needed to numerically reproduce the sediment distribution observed in the field, which is nearly the same tsunami magnitude as found for the northern Gulf of Kyparissia.

As demonstrated in this PhD thesis, numerical simulations that take account of sedimentary palaeotsunami field traces allow not only to reconstruct the dynamics, the extent and associated frequency of past tsunami impacts in a specific area, but also to derive plausible tsunami wave scenarios for the future in view of risk assessment. Although this approach may demand a large number of assumptions which have to be made in the numerical model, reliable simulation results can be achieved if the local palaeotopography, the morphodynamics of tsunami inundation and a wide variety of plausible hydrodynamic boundary conditions are considered, as performed here. The conducted studies reveal high tsunami risks for the northern Gulf of Kyparissia and the Ambrakian Gulf, which is in contrast to the results of conventional tsunami simulation studies for western Greece without comparison and calibration using field data.

Zusammenfassung

Ozeane sind eine wichtige Ressource der modernen Gesellschaft, sie können jedoch auch große Risiken für Küstenbewohner darstellen. Sturmfluten und Tsunamis zählen zu den größten Naturgefahren des Meeres, die in der Vergangenheit unzählige Leben gefordert haben. Trotz großer Fortschritte in der wissenschaftlichen Erforschung solcher Wellenphänomene können extreme Sturmfluten und Tsunamis nach wie vor zu hohen Opferzahlen führen, insbesondere in Entwicklungsländern. Während Sturmfluten dank moderner Technologie präzise und frühzeitig vorhergesagt werden können, ist die Warnung vor Tsunamis und betroffener Küstenbereiche eine besondere Herausforderung. Tsunamis sind wie Sturmfluten durch die Dynamik von Schwerewellen gekennzeichnet. Sie werden durch Erdbeben, gravitative Massenbewegungen, vulkanische Aktivität, atmosphärische Störungen oder Meteoriteneinschläge ausgelöst und können ohne wesentliche Energieverluste ganze Ozeane in Form von Flachwasserwellen durchqueren. Aufgrund ihrer großen Wellenlänge von bis zu mehreren hundert Kilometern treffen Tsunamis mit einem enormen Impuls auf die Küste, verursachen beachtliche Wellenaufbauhöhen und Eindringtiefen und hinterlassen deutliche sedimentologische sowie geomorphologische Spuren an der Küste. Tsunami-Ablagerungen oder sogenannte Tsunamite weisen eine Reihe typischer sedimentologischer Merkmale auf, wie z. B. basale Erosionsdiskordanzen, eine nach oben abnehmende Korngröße und Erosionsgerölle. Diese Merkmale sind Ausdruck der wirkenden hydrodynamischen Prozesse während der Überflutung. Tsunamite in der Sedimentstratigraphie von Küsten lassen nicht nur auf die Häufigkeit prä-/historischer Tsunamis in einer Region schließen, sondern erlauben auch eine Kalibrierung und Validierung hydro- und morphodynamischer numerischer Modelle, die ein geeignetes Hilfsmittel sind, um regionale Tsunami-Risiken zu untersuchen. Eine gebräuchliche Wellentheorie in numerischen Modellen, um das nicht lineare, dispersive Wellenverhalten von Tsunamis während ihrer Ausbreitung zu beschreiben, ist die Boussinesq-Approximation.

Basierend auf sedimentologischen Geländebefunden sowie hydro- und morphodynamischen numerischen Simulationen beschäftigt sich die vorliegende Dissertation mit prä-/historischen Tsunami-Ereignissen im östlichen Mittelmeer in Westgriechenland. Die Westküste Griechenlands liegt aufgrund ihrer Nähe zur Subduktionszone der Afrikanischen und Eurasischen Platte im seismisch-tektonisch aktivsten und damit im Tsunami gefährdetsten Bereich des gesamten Mittelmeerraums. Die Studie untersucht Paläo-Tsunami-Ereignisse in zwei ausgedehnten Buchten der griechischen Westküste — der nördliche Golf von Kyparissia (westliche Peloponnes) und der Ambrakische Golf (küstennahes Epirus/Akarnanien). Für das erste Untersuchungsgebiet wurden verschiedene digitale Höhenmodelle (DHM) erstellt, welche die heutige Topographie und die Paläo-Topographie zur Zeit der aus Geländedaten abgeleiteten prähistorischen Tsunami-Ereignissen wiedergeben. Unter Berücksichtigung der verschiedenen Paläo-/DHMs und zwei verschiedener Tsunami-Wellenarten (solitäre Wellen und *N*-Wellen mit führendem Wellental) wurden zunächst die Auswirkungen sowohl der unterschiedlichen Paläo-/Topographien als auch der verschiedenen hydrodynamischen Grenzbedingungen auf die Tsunami-Überflutung im numerischen Modell methodisch untersucht. Die Ergebnisse zeigen, dass die *N*-Wellen eine stärkere Überflutung der Küste verursachen als solitäre Wellen mit derselben Wellenamplitude. Dieses Phänomen, bekannt als *N*-Wellen-Effekt, nimmt mit steigender Wellenamplitude zu und wird zum ersten Mal für ein bestimmtes Untersuchungsgebiet demonstriert. Die verschiedenen Paläo-/Topographien verursachen noch deutlichere Unterschiede in der Überflutungsdynamik

als die verschiedenen Wellenarten. Dies ist auf die erheblichen vertikalen Krustenbewegungen im nördlichen Golf von Kyparissia zurückzuführen, die in den Paläo-DHMs berücksichtigt wurden. Mithilfe der unterschiedlichen Grenzbedingungen und Paläo-/DHMs lässt sich eine gute Übereinstimmung des numerischen Modells mit den sedimentologischen Geländebeobachtungen erreichen. Dies erlaubt es, die lokale Überflutungsdynamik von Paläo-Tsunamis an drei Schlüsselstellen des nördlichen Golfs von Kyparissia zu rekonstruieren — dem Katakolo Pass, dem Epitalio Pass und dem Tal von Kato Samiko. Die beste Übereinstimmung zwischen den Simulationen und Geländedaten besteht für solche Szenarien, die auf dem ersten erstellten Paläo-DHM beruhen, das von einem tektonischen Vertikalversatz von 13 m entlang der Hauptstörungszonen der Region ausgeht. In diesen Szenarien sind solitäre oder *N*-Wellen mit Wellenamplituden zwischen 3 m und 6 m erforderlich, um eine deutliche Überflutung der drei Gebiete in Übereinstimmung mit den sedimentologischen Geländebefunden zu verursachen.

Für das zweite Untersuchungsgebiet, dem Ambrakischen Golf, wurden 151 Bohrkerne der Region hinsichtlich sedimentologischer Anzeichen vergangener Tsunami-Ereignisse ausgewertet. Ziel war es, die sich hieraus ergebende räumliche Verteilung der detektierten Tsunami-Sedimente in einem hydro- und morphodynamischen numerischen Tsunami-Modell auf der Grundlage verschiedener *N*-Wellen-Szenarien zu reproduzieren. Sowohl die Geländebefunde als auch die Simulationen zeigen eine Abnahme der Tsunami-Wellenenergie im Ambrakischen Golf von West nach Ost. Zugleich sind jedoch mehrere Szenarien erforderlich, um die Geländebeobachtungen in verschiedenen Bereichen des Untersuchungsgebietes erklären zu können, wie es auch die Geländebefunde selbst andeuten. Während die detektierten Tsunami-Ablagerungen im westlichen Golf besser mit den simulierten kurzwelligen Tsunamis aus südlicher und westlicher Richtung übereinstimmen, sind langwellige Tsunamis aus Richtung Westen besser geeignet, um die Geländebefunde im inneren Golf zu erklären. Dagegen zeigen Tsunamis aus südwestlicher Richtung grundsätzlich kaum Übereinstimmungen mit den Geländedaten. Vergleichbar mit der Größenordnung, wie sie für den nördlichen Golf von Kyparissia festgestellt wurde, sind *N*-Wellenamplituden zwischen 4 m und 6 m erforderlich, um die im Gelände beobachtete Verteilung von Tsunami-Sedimenten im numerischen Modell zu reproduzieren.

Die Ergebnisse der vorliegenden Dissertation zeigen, dass sich mithilfe von numerischen Simulationen, die sedimentologische Geländebefunde von Paläo-Tsunamis berücksichtigen, nicht nur die Dynamik, die Ausmaße und die Häufigkeit prä-/historischer Tsunamis rekonstruieren lassen, sondern auch plausible Szenarien möglicher zukünftiger Tsunami-Ereignisse abgeleitet werden können. Dies stellt eine wichtige Grundlage für Risikoeinschätzungen dar. Wenngleich dieser Ansatz zahlreiche Annahmen im numerischen Modell erfordert, so können dennoch verlässliche Simulationsergebnisse erzielt werden, sofern die regionale Paläo-Topographie, die morphodynamischen Prozesse der Tsunami-Überflutung sowie eine Vielzahl möglicher hydrodynamischer Grenzbedingungen Berücksichtigung finden, wie es in dieser Studie erfolgt ist. Auf dieser Grundlage kann für den nördlichen Golf von Kyparissia und den Ambrakischen Golf ein hohes Risiko durch Tsunamis abgeleitet werden. Diese Einschätzung steht im Gegensatz zu den Ergebnissen konventioneller Simulationsstudien für Westgriechenland, die nicht auf einem Vergleich und einer Kalibrierung mit Geländebefunden beruhen.

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¹This chapter will be submitted as a review paper.

²This chapter is based on the paper by RÖBKE *et al.* (2016a) submitted to *Continental Shelf Research*.

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³This chapter is based on the paper by RÖBKE *et al.* (2015).

⁴This chapter is based on the paper by RÖBKE *et al.* (2016b) submitted to *Geophysical Journal International*.

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List of Abbreviations and Symbols

AD	after Christ (from Latin <i>Anno Domini</i>)
a.s.l.	above sea level
BC	before Christ
DART	Deep-ocean Assessment and Reporting of Tsunamis
b.s.l.	below sea level
cal	calibrated years according to the radiocarbon dating method
CFL	Courant-Friedrichs-Lewy
DEM	digital elevation model
DGPS	differential global positioning system
DHM	digitales Höhenmodell
EFZ	Epitalio fault zone
GEBCO	General Bathymetric Chart of the Oceans
KdV	Korteweg-de-Vries
LDN-wave	leading-depression <i>N</i> -wave
LEN-wave	leading-elevation <i>N</i> -wave
M_s	surface wave magnitude
NLSW equations	non-linear shallow water equations
sech	hyperbolic secant
SRTM	Shuttle Radar Topography Mission
tanh	hyperbolic tangent

1 Introduction

1.1 Significance of coastal hazards and risk assessment

Human beings are still inferior to the enormous power of the sea. The close interaction of atmosphere and solid earth with the ocean regularly results in disastrous storm surges and tsunami impacts, which have claimed innumerable lives through history. Braving the violent motions of the sea, engineering work early enabled coastal life. As reported by Pliny the Elder, already in the first century AD, German settlers built artificial dwelling hills in the North Sea to rise above the highest flood to be expected (PARKER 2010, p.62). Since then, coastal management has seen major advances, nowadays offering solid protection against coastal hazards worldwide. Accordingly, economic and leisure activities increasingly concentrate on the coast having resulted in a percentage of more than 10 % of the world's population that currently lives in coastal lowlands less than 10 m a.s.l. (above sea level) (MCGRANAHAN *et al.* 2007, p.22). This being more than 700 million people—most of them living in megacities with more than 10 million inhabitants each—the vulnerability of coasts to oceanic impacts is higher than ever before. This has become particularly obvious from catastrophes in recent history, such as the 1970 Bhola cyclone in Bangladesh with more than 300 000 victims, the Indian Ocean tsunami in 2004 claiming almost 230 000 lives, Hurricane Katrina in the southern USA in 2005 (more than 1800 casualties), Cyclone Nargis in 2008 in Myanmar, which killed more than 140 000 people, and the 2011 Tōhoku tsunami in Japan with a death toll of over 17 000 (PARKER 2010, p.1; NDGC/WDS 2015a).

Prediction is the very essence of coastal management. The knowledge of forthcoming storm surges or tsunamis does not avert the impact itself, but human loss can significantly be reduced when prepared for the striking sea. This is demonstrated by comparing the death tolls associated with the above-mentioned natural disasters, which are clearly smaller for industrial, well-engineered countries than for developing countries despite a similar intensity of storm surges or tsunamis. Whereas forecast techniques and safety constructions are highly advanced with regard to storm surges, mitigation of tsunami impacts is a particularly demanding challenge due to the sudden and infrequent nature of this phenomenon. Owing to the low frequency of tsunamis, information on past tsunami events is essential in view of a reliable future risk assessment. Such information is available from written accounts on the one hand and from sedimentary deposits and other geomorphological field traces of tsunamis in the coastal geological record on the other hand. The latter may allow the reconstruction of detailed inundation dynamics, the intensity and frequency of palaeotsunamis having occurred thousands of years ago, provided that sedimentary signatures are preserved. Synthesising field findings with hydro- and morphodynamic numerical simulation techniques has proven particularly effective. Numerical models allow to go through various possible tsunami scenarios in order to find scenarios that compare favourably with sedimentary tsunami field traces, which are then considered as a reliable base in respect to future tsunami risk assessment. This, for instance, was impressively demonstrated in view of the 2011 Tōhoku tsunami in Japan. Historical accounts as well as sedimentary and numerical investigations conducted prior to this event had already revealed an extraordinary tsunami impact in north-east Japan in 869 AD—the so-called Jōgan tsunami (ABE *et al.* 1990; MINOURA *et al.*

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2001; SAWAI *et al.* 2007, 2008; SATAKE *et al.* 2008; NAMEGAYA *et al.* 2010; SUGAWARA *et al.* 2011). Unfortunately, the findings of these studies have never been considered in regional zoning plans—a glaring omission of the coastal management, since later investigations revealed a close similarity between both tsunami events (SHISHIKURA *et al.* 2011; SAWAI *et al.* 2012; SUGAWARA *et al.* 2013; NAMEGAYA & SATAKE 2014).

1.2 Tsunami risks in the Mediterranean

Based on the combined use of sedimentary field traces as well as hydro- and morphodynamic numerical models, this PhD thesis focuses on pre-/historical tsunamis in the Mediterranean Sea in western Greece, which is regularly affected by tsunami impacts due to high seismic activity (e.g. AMBRASEYS 1962; ANTONOPOULOS 1979; SOLOVIEV 1990; SOLOVIEV *et al.* 2000; PAPADOPOULOS & FOKAEFS 2005; SCHIELEIN *et al.* 2007; AMBRASEYS & SYNOLAKIS 2010; HADLER *et al.* 2012; NDGC/WDS 2015a; NDGC/WDS 2015b; NOAA 2015a). About 13.3 % of all tsunamis recorded in the world's oceans have occurred in the Mediterranean excluding the Sea of Marmara and the Black Sea (for this and the following information on tsunami reports see SCHIELEIN *et al.* 2007, pp.159, 160). Since 1900, 53 tsunamis have been recorded in the Mediterranean Sea, which is 7.1 % of all 745 reports worldwide. This is about twice the percentage compared to the Indian Ocean (3.5 %), the place of the devastating 2004 Indian Ocean

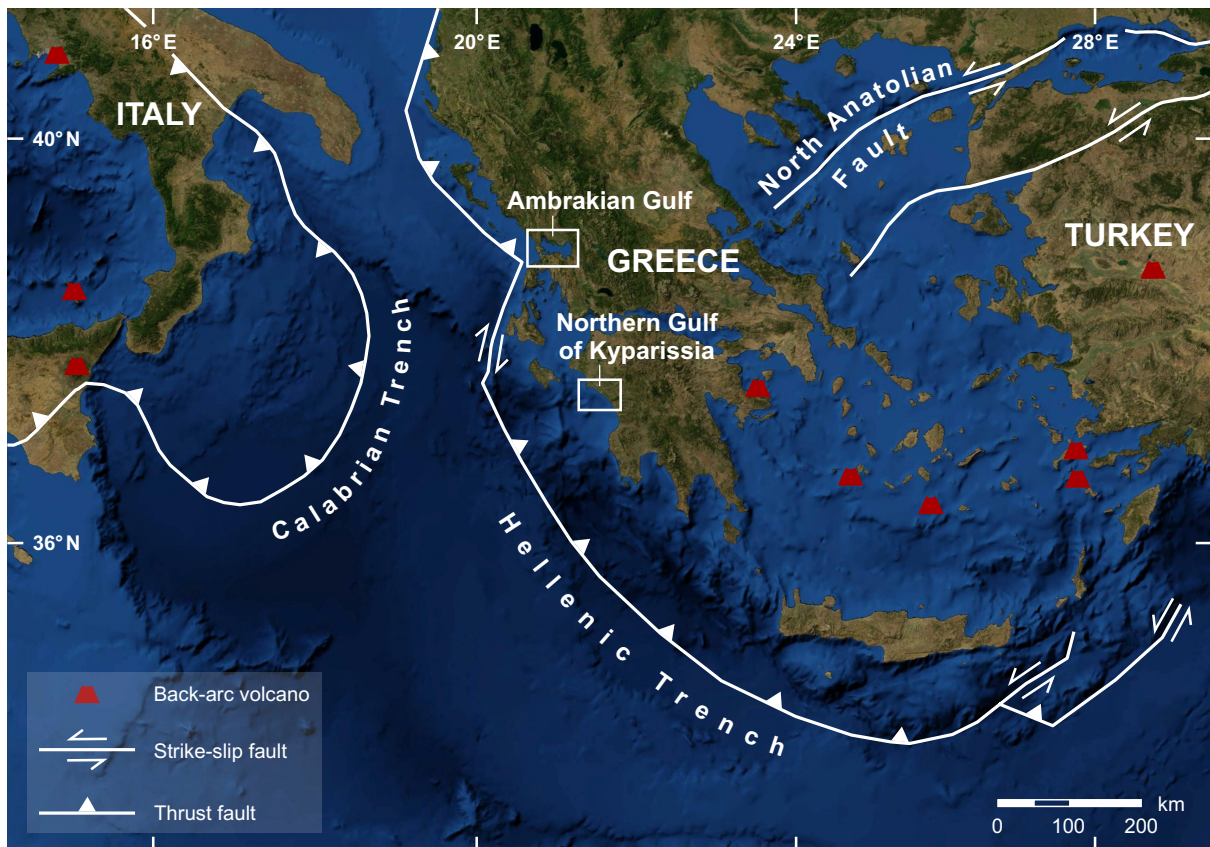


Figure 1.1: Satellite picture and shaded offshore topography (Google Inc. 2013) of the central and eastern Mediterranean, including the simplified tectonic regime according to FACCENNA *et al.* (2001, p. 810), KOKKALAS *et al.* (2006, p. 100) and POLONIA *et al.* (2011, p. 2). White boxes show the two study areas of this thesis — the northern Gulf of Kyparissia and the Ambrakian Gulf.

tsunami. Although the number of reports is not indicative of the consequences of the impacts, statistics clearly demonstrate tsunamis being a frequent phenomenon in the Mediterranean world and a major risk to million residents and tourists.

Besides written accounts, sedimentary and geomorphological field traces give evidence of numerous tsunami events in the Mediterranean during the Holocene (last 10 000 years), e.g. in southern Italy and Sicily (e.g. MASTRONUZZI & SANSÒ 2000, 2004; GIANFREDI *et al.* 2001; SCICCHITANO *et al.* 2007; DE MARTINI *et al.* 2010; SMEDILE *et al.* 2011, 2012), in the Ionian Islands as well as coastal Epirus/Akarnania in Greece (e.g. VÖTT *et al.* 2006, 2007, 2008, 2009a,b, 2010a, 2011a; MAY *et al.* 2012a,b; WILLERSHÄUSER *et al.* 2013; FISCHER *et al.* 2015), along the western and southern Peloponnese (e.g. SCHEFFERS *et al.* 2008a; RÖBKE *et al.* 2012, 2013; VÖTT 2013; VÖTT *et al.* 2010a, 2011b, 2015; HADLER *et al.* 2015a; WILLERSHÄUSER *et al.* 2015a,b), in the Aegean (e.g. PIRAZZOLI *et al.* 1999; DOMINEY-HOWES 2002), on Crete (e.g. DOMINEY-HOWES *et al.* 1999; SCHEFFERS & SCHEFFERS 2007; BRUINS *et al.* 2008), on Cyprus (KELLETAT & SCHELLMANN 2002), along the Levantine coast (e.g. MORHANGE *et al.* 2006; REINHARDT *et al.* 2006) and in the Nile Delta (STANLEY & BERNASCONI 2006).

Indicated by both, the geographical distribution of tsunami field traces as well as written accounts, the eastern Mediterranean is particularly affected by tsunamis. Most tsunami reports are known from the west coast of Greece with 44 and 47 events respectively since 373 BC (SOLOVIEV *et al.* 2000, p. 14; SCHIELEIN *et al.* 2007, pp. 164, 193–199). Accordingly, the average recurrence interval of tsunamis in this region for the period since 1900 is 8 years only (SCHIELEIN *et al.* 2007, p. 164). The high tsunami risk in western Greece is closely related to the high seismicity of the nearby subduction zone of the Hellenic Trench, located less than 100 km off the coast and about 5267 m deep (Fig. 1.1; BARALE 2008, p. 14). Here, the African Plate is being subducted beneath the Eurasian Plate and Aegean Microplate at a rate of 35 mm per year (SHAW & JACKSON 2010, p. 966). This plate movement regularly results in shallow-focus earthquakes with a magnitude of $M_s \geq 6.5$ at focal depths of less than 40 km along shallow dipping thrust faults as long as 300 km (PAPAZACHOS 1996; BENETATOS *et al.* 2004; USGS 2015; cf. RÖBKE *et al.* 2013, pp. 74, 75). Associated with significant co-seismic vertical crust movements, such earthquakes are believed to be the favourite earthquake type to generate tsunamis (Chap. 2.3.1). Earthquakes of similar magnitude and with comparable geometry are also known from the nearby Calabrian Trench as the second major subduction zone of the African and Eurasian Plate in the Mediterranean Sea (Fig. 1.1; BOCCALETTI *et al.* 1984; TORTORICI *et al.* 1995; MONACO & TORTORICI 2000; POLONIA *et al.* 2011). Located in the central Ionian Sea offshore southern Italy and Sicily, the Calabrian Trench is therefore another tsunami source with regard to the west coast of Greece.

Owing to the distinct offshore topography and coastal relief conditions in the central and eastern Mediterranean, also subaerial and submarine gravitational mass wastings are known as a trigger of tsunamis in western Greece (SCHIELEIN *et al.* 2007, pp. 193–199; cf. RÖBKE *et al.* 2013, pp. 75, 76). This is particularly true for mass wastings associated with Mount Etna in Sicily (Fig. 1.1; PARESCHI *et al.* 2006a,b) and the Hellenic Trench (FERENTINOS 1991, 1992; PAPANIKOLAOU *et al.* 2007, p. 21). Tsunamis in western Greece may further arise from volcanic activity of Mount Etna or volcanoes in the southern Aegean due to, e.g. pyroclastic flows, submarine volcanic explosions or caldera formation (SCHIELEIN *et al.* 2007, pp. 193–199; cf. RÖBKE *et al.* 2013, p. 76). Moreover, atmospheric disturbances may produce local meteotsunamis in marine inlets in western Greece (PAPADOPOULOS 1993). Meteorite impacts, however, are not known to have caused tsunamis in the Mediterranean Sea so far (VÖTT & MAY 2009, p. 42) but must still be regarded as a potential trigger mechanism.

1.3 Study area, aims and outline

This study focuses on tsunami events in two marine inlets of the eastern Ionian Sea in western Greece—the Gulf of Kyparissia and the Ambrakian Gulf (Fig. 1.1). The Gulf of Kyparissia is a 80 km long bay of the western Peloponnese between the village of Katakolo in the north and the town of Filiatra in the south with a maximum water depth of 1850 m (POULOS *et al.* 2002, p. 250). The main area of interest is the northern part of the gulf, which is characterised by the Alpheios River mouth and extensive lowlands bounded by hilly coastlands (Fig. 3.1a). The regional capital of the area is Pyrgos counting 25 180 inhabitants (ELSTAT 2011). The second study area, the Ambrakian Gulf, is located almost 135 km to the north of the Gulf of Kyparissia in coastal Epirus/Akarnania to the north of Lefkada Island. The Ambrakian Gulf has a maximum N-S- and W-E-extent of about 20 km and 35 km respectively and a maximum water depth of 65 m only (Fig. 5.1; UKHO 2010). The gulf is connected to the Ionian Sea by the very narrow Strait of Preveza. The largest cities of the Ambrakian Gulf are Preveza with a population of 17 724 and Arta with 21 642 inhabitants (ELSTAT 2011).

Based on numerical forward simulations and sedimentary field traces found in both study areas, the current PhD thesis deals with tsunami events in the northern Gulf of Kyparissia and the Ambrakian Gulf of the last 8000 years with the main aims to

- investigate the dynamics, the extent and associated frequency of palaeotsunami inundation using sedimentary vibracore data,
- create numerical tsunami forward models that take account of the contemporary palaeotopography of detected tsunami events and consider tsunami sediment processes,
- study the effects of different hydrodynamic boundary conditions and tsunami waveforms and of different palaeo-/topographies on the onshore tsunami response in a numerical model in order to derive their implications for the comparison of simulation results with palaeotsunami field data,
- create and simulate plausible tsunami wave scenarios that closely accord with sedimentary field data,
- assess local tsunami risks considering information on past tsunami events as derived from field data and simulation results.

The thesis is divided into four main chapters, each of which forming a separate, stand-alone paper published, submitted or planned to be submitted in/to international journals. Chapter 2 provides a geoscientific literature review of the tsunami phenomenon (for this and the following see Fig. 1.2). Here, the fundamental wave physics and the various generation mechanisms of tsunamis as earthquakes, mass wastings or volcanic eruptions are clarified. Furthermore, sedimentary and geomorphological field traces of tsunamis as well as numerical tsunami modelling and simulation techniques are presented and discussed, both these topics forming the basis of the studies in the following research chapters.

Chapter 3 presents a methodological investigation in regard to the comparison of sedimentary tsunami field traces and tsunami simulations in the first study area, the northern Gulf of Kyparissia. Three different digital elevation models (DEM) are created, one reflecting the recent topography and two reflecting possible palaeotopographies associated with prehistorical tsunami landfall indicated by field data. Based on the three DEMs, solitary and leading-depression *N*-waves are simulated in order to methodologically analyse the effects of different hydrodynamic

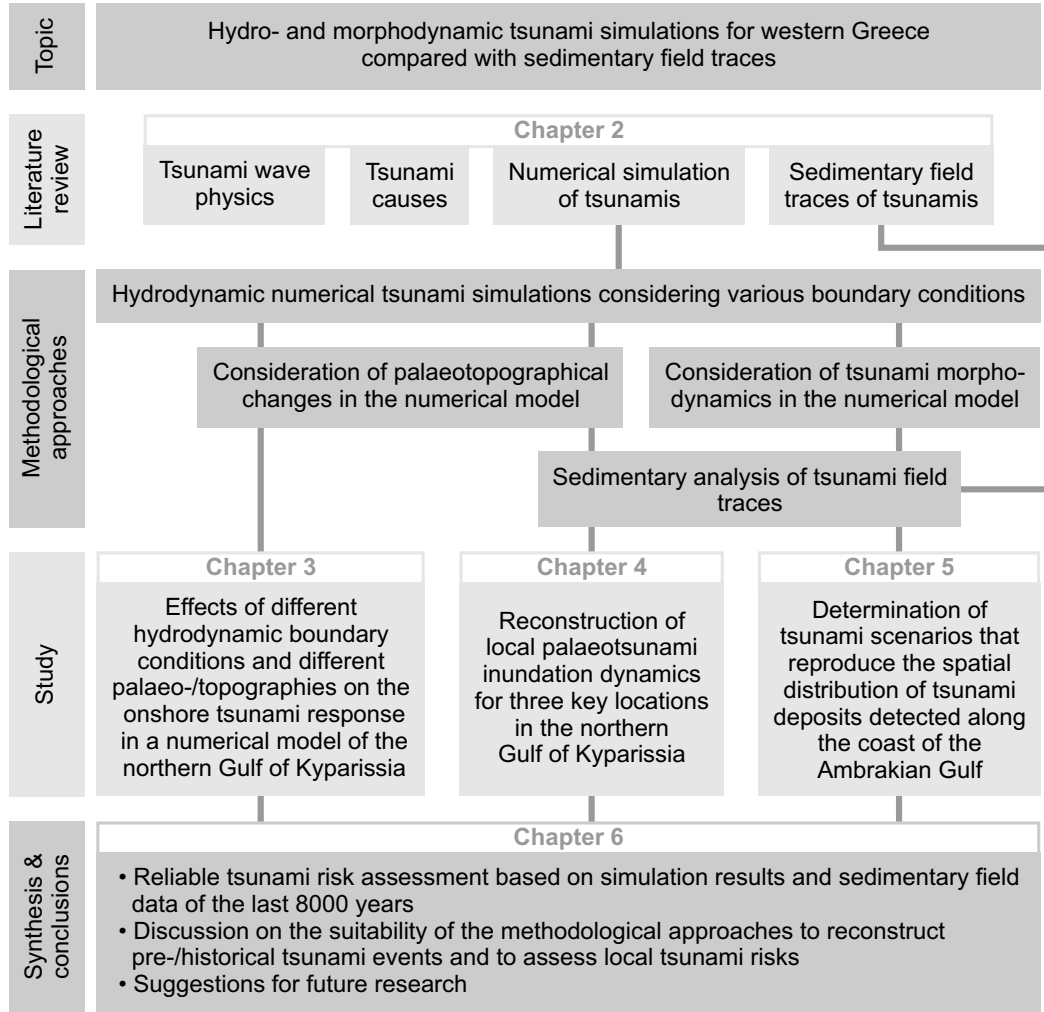


Figure 1.2: Conceptual framework of the current PhD thesis, including the contents, methodological approaches and main objectives of the conducted studies.

boundary conditions and tsunami wave theories as well as different palaeo-/topographies on the onshore tsunami response in the numerical model. Moreover, the possibility and difficulty of numerical tsunami models to be calibrated and compared with palaeotsunami field traces as well as the regional tsunami risk are discussed.

Based on the methodology and simulation results of Chap.3, Chap.4 provides a detailed comparison of sedimentary tsunami field traces and tsunami simulations for three key locations in the northern Gulf of Kyparissia — the Katakolo Pass, the Epitalio Pass and the valley of Kato Samiko. Taking account of the recent topography and two possible palaeotopographies of the time of prehistorical tsunami impacts, those tsunami wave scenarios are determined that can explain tsunami deposits detected in each location. These scenarios and the field data are used to reconstruct small-scale palaeotsunami inundation dynamics for all three key sites.

Chapter 5 concentrates on tsunami impacts on the Ambrakian Gulf, the second study area of this thesis. A total of 151 vibracores taken along the gulf are systematically analysed in view of potential tsunami event deposits in order to derive the scale of pre-/historical tsunami inundation in the area. A hydro- and morphodynamic numerical tsunami model is created that takes account of various tsunami wave scenarios and considers sediment processes. By comparing the computed

1 Introduction

tsunami sedimentation in the numerical model with the spatial distribution of detected tsunami deposits in the entire study area, those scenarios are determined that reproduce field data most accurately, which are therefore the favourite scenarios that may occur also in present times.

Chapter 6 summarises and synthesises the main findings of the three research chapters (Chaps. 3–5). Based on these findings, a reliable tsunami risk assessment for the northern Gulf of Ky-parissia and the Ambrakian Gulf is derived considering tsunami field data of the last 8000 years and tsunami simulation results. The chapter further discusses the suitability of the chosen methodological approaches to investigate pre-/historical tsunami impacts in the study areas and gives suggestions for future research.

2 The tsunami phenomenon¹

Abstract With human activity increasingly concentrating on coasts, tsunamis (from Japanese *tsu* = *harbour*, *nami* = *wave*) are a major natural hazard to today's society. Stimulated by disastrous tsunami impacts in recent years, for instance in south-east Asia (2004) or in Japan (2011), tsunami science has significantly flourished, which has brought great advances in risk assessment and mitigation plans. Based on tsunami research of the last decades, this chapter provides a thorough treatise on the tsunami phenomenon from a geoscientific point of view. Starting with the wave physics, tsunamis are introduced as shallow water waves or wave trains crossing entire oceans without major energy loss. At the coast, tsunamis typically show wave shoaling, funnel and resonance effects as well as a significant run-up and backflow. Tsunami waves are caused by a sudden vertical displacement of the water column due to a number of various trigger mechanisms. Such are earthquakes as the main trigger, submarine and subaerial mass wastings, volcanic activity, atmospheric disturbances (meteotsunamis) and cosmic impacts, as is demonstrated by giving corresponding examples from the past.

Tsunamis are known to have a significant sedimentary and geomorphological off- and onshore response. So-called tsunamites form allochthonous high-energy deposits that are left at the coast during tsunami landfall. Tsunami deposits show typical sedimentary features, as basal erosional unconformities, fining-upward and -landward, a high content of marine fossils, rip-up clasts from underlying units and mud caps, all reflecting the hydrodynamic processes during inundation. Tsunami field traces can be used for the calibration and validation of numerical tsunami models, which are dealt with in the last section of this chapter. First, the difference between modelling and simulation is pointed out followed by a short description of the basic modelling techniques, including discretisation, guidelines for appropriate temporal and spatial resolution as well as the nesting method. A basic theory behind numerical tsunami models is the Boussinesq approximation — a simplification of the three-dimensional Navier-Stokes equations —, which adequately reflects the non-linear, dispersive wave behaviour of tsunamis. The fully non-linear Boussinesq equations allow the simulation of tsunamis, e.g. in the form of solitary or *N*-waves, in numerical models.

2.1 Introduction

Tsunamis are long water waves or sets of long water waves, usually observed in oceans or in large lakes, that originate from a sudden vertical displacement of the water column. Triggering mechanisms comprise earthquakes, submarine or subaerial gravitative mass wastings, volcanism, cosmic impacts or atmospheric disturbances. Mistakenly called *oshio* at first (from Japanese *high tide*), later the term tsunami was introduced for this wave phenomenon by the Japanese (PARKER 2010, p. 140), meaning *harbour (tsu) wave (nami)*. This term relates to the fact that tsunamis are often not perceived in the open sea but form their destructive power near the coast, especially in marine inlets or in harbours.

With coasts becoming increasingly popular as places of residence, tourism and industry, tsunamis are a major threat to today's society. Recent events in south-east Asia (2004) and

¹This chapter will be submitted as a review paper.

Japan (2011) have demonstrated the tremendous power of tsunamis on coastal landscapes resulting in huge human and economical losses. According to the global historical tsunami database (NDGC/WDS 2015a), altogether about 940 000 deaths have been counted in connection with tsunami impacts worldwide. Adding an unknown number of unrecorded victims especially associated with pre-/historical tsunami events, the actual death toll must be assumed to be well above one million people. The total amount of economic loss is even more difficult to estimate. Solely regarding the two most destructive events in recent history, estimates of the total financial damage range from \$15 billion to \$80 billion for the Indian Ocean tsunami in 2004 (BRYANT 2008, p.177) and up to \$335 billion with respect to the Tōhoku tsunami and earthquake in Japan (2011) for the year 2011 alone (DANIELL & VERVAECK 2012, p.17). The latter event was the costliest natural disaster ever recorded.

Stimulated by recent tsunami events, geoscientific research of this natural phenomenon has significantly increased during the last decades, in particular since the Indian Ocean tsunami in 2004 (e.g. KEATING 2006, pp.392, 393; RABINOVICH *et al.* 2015, pp.615, 616). The main topics of tsunami science are numerical simulation of tsunami generation, propagation and inundation, field surveys in search of on- and offshore sedimentary and geomorphological traces of pre-/historical tsunami impacts, research on tsunamigenic earthquakes and gravitational mass wastings as well as the observation of tsunami waves in the oceans and along coasts using, e.g. DART (Deep-ocean Assessment and Reporting of Tsunamis)-buoys, satellites or other instruments (KEATING 2006, pp.388, 389). Recent years have seen major advances in these research fields, which have considerably improved the understanding of the complex tsunami phenomenon (RABINOVICH *et al.* 2015, pp.615, 616). Accordingly, also tsunami forecast techniques have extended and significantly gained in reliability. This is particularly true when tsunami risk assessment is based on modern warning systems in combination with hydro- and morphodynamic numerical models that are calibrated with sedimentary field data and — if available — deep ocean tsunami measurements and information on the trigger mechanism of past tsunami events (cf. Chaps. 3–5). Although a reliable tsunami risk assessment still does not prevent the impact itself, the progress in tsunami science and forecast techniques during the last years has certainly reduced the relative number of tsunami victims since 26 December 2004 (RABINOVICH *et al.* 2015, pp.615, 616).

Based on tsunami research of the last decades, this literature review provides a thorough treatise on the tsunami phenomenon from a geoscientific point of view. Starting with the fundamental wave physics of tsunamis in Chap. 2.2, the main physical features of tsunami wave propagation and interaction with the coast are dealt with. Chapter 2.3 gives detailed insights into the various generation mechanisms of tsunamis ranging from earthquakes as the main trigger to submarine and subaerial mass wastings, volcanic activity, atmospheric disturbances (meteotsunamis) and cosmic impacts by giving corresponding examples from the past. Based on the insights into the wave physics of tsunamis, Chap. 2.4 deals with the sedimentary and geomorphological imprint of tsunami waves with the main focus on onshore tsunami deposits and their differences to storm-related sediments. Chapter 2.5 provides an outline of modern numerical tsunami modelling and simulation techniques, including a brief treatise on (i) the basic guidelines of tsunami modelling, (ii) the governing hydrodynamic equations of numerical tsunami models and (iii) the various waveforms used to mathematically describe tsunami waves in numerical models. Finally, a summary of the main findings of this review is given in Chap. 2.6.

2.2 Wave physics of tsunamis

Despite the fundamental physical differences between tsunami waves and ordinary wind-generated waves, both types of waves mostly use the same terminology. Thus, wavelength, phase

velocity, wave period and wave height are the main parameters describing tsunami waves, too. Furthermore, tsunamis—like wind-generated—are affected by refraction and diffraction and undergo shoaling and breaking when approaching the coast (BRYANT 2008, p.27). During tsunami formation, the body of water is tipped away from its equilibrium state due to the impulsive vertical displacement of the water column. Water particles are caused to oscillate because gravity is acting as restoring force against this movement. Consequently, tsunamis are described as gravity waves (KRUISZ & HITZENBERGER 2011, p.179).

Compared to wind-generated waves (typical wavelengths 50 m to 400 m; PARKER 2010, p.105), tsunamis are long-wave phenomena, their wavelength frequently ranging from 100 km to 500 km in the open sea (LAUTERJUNG *et al.* 2009, p.39; KUHLMANN 2014, p.148). Even in coastal areas, the crest-to-crest distance of tsunami waves hardly falls below 10 km (WARD 2002, pp.2, 3; BRYANT 2008, p.29). With their wavelengths being significantly larger than the water depth of the sea, averaging 3800 m (MARTIN & EIBLMAIER 2002, keyword *hypsographische Kurve*), tsunamis usually travel as so-called shallow water waves. Shallow water waves satisfy the following condition (WIEGEL 1992, p.13; BRYANT 2008, p.31):

$$d < \frac{\lambda}{2} \quad (2.1)$$

where d = water depth [m]
 λ = wavelength [m].

This condition is met by most tsunami waves. Considering Eq.2.1 and the fact that the wave base of water waves is about half the wavelength, the whole water column is influenced by shallow water waves or tsunamis, even in the deep sea (WIEGEL 1992, pp.13, 17–19; KOPP & WEINREIBE 2009, p.20). This makes tsunamis stand out from wind-generated waves, which only affect the upper water column due to their shorter wavelength.

The phase velocity or celerity of shallow water waves and tsunamis respectively is expressed by the following equation (HELENE & YAMASHITA 2006, p.856):

$$c = \sqrt{gd} \quad (2.2)$$

where c = phase velocity or celerity of the wave [m s^{-1}]
 g = gravitational acceleration (9.81 m s^{-2}).

Consequently, the phase velocity of tsunami waves is positively correlated to gravity and water depth (cf. Fig.2.1) and is independent of wave height and wavelength. For example, tsunamis generated in a 10 000 m deep oceanic trench travel at an impressive velocity of 313 m s^{-1} in this area. In the 4000 m to 6000 m deep abyssal plain, tsunami wave velocity amounts to 221 m s^{-1} on average. Even when approaching the shallow waters of the shelf (maximum 200 m deep), tsunami waves still move at a considerable speed of 44 m s^{-1} . Finally, in close vicinity to the coast, tsunami wave velocity is reduced to 10 m s^{-1} in 10 m deep water.

The loss of energy of a propagating water wave is inversely proportional to its wavelength (HELAL 2010, p.226). Considering the enormous wavelength and phase velocity of tsunami waves, they are thus able to cross entire oceans in short time spans without major energy loss. Such tsunamis are called *teletsunami*. A popular example is the teletsunami that was generated on 22 May 1960 next to the central Chilean coast due to a strong earthquake measuring $M_s = 8.9$. The resulting wave arrived at the east coast of Japan almost 21 h later, some 17 000 km distant (BRYANT 2008, p.144).

2 The tsunami phenomenon

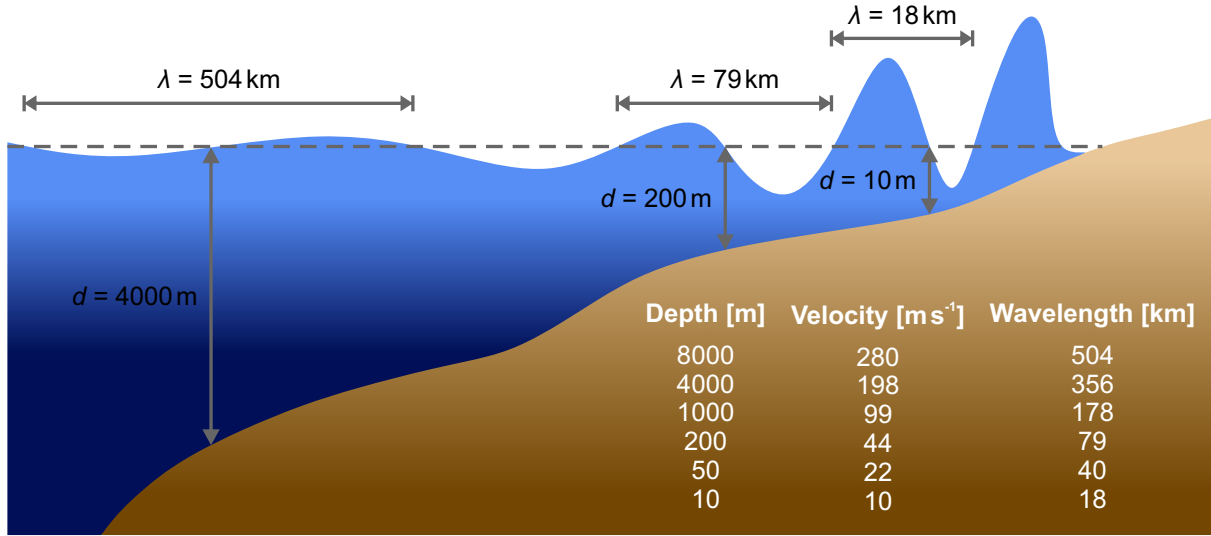


Figure 2.1: Tsunami wave shoaling and change in wave parameters (phase velocity, wavelength and wave height) due to decreasing water depth from the deep ocean towards the coast (after UNESCO-IOC 2006, p. 4). Wavelengths are based on a typical tsunami wave period of constantly 30 min.

The wave period of tsunamis or wind-generated waves, defined as the time span between two consecutive wave crests, is calculated using the following relation:

$$T = \frac{\lambda}{c} \quad (2.3)$$

where T = wave period [s].

In general, the wave period of tsunami waves, also called *tsunami window*, varies from a few minutes to two hours (BOLT *et al.* 1975, p. 134; HELAL & MEHANNA 2008, p. 788).

The wave height of tsunamis, meaning the vertical distance between crest and trough, may reach several metres at the wave's origin, depending on the intensity of the triggering mechanism (Chap. 2.3). Receding from the source, tsunami waves rapidly flatten in the open sea, their wave height usually falling below 1 m (BOLT *et al.* 1975, p. 133; DAWSON & STEWART 2007a, pp. 167–172). Therefore, and because of the large wavelength, tsunamis are often not perceived in the open sea. Approaching the coast and shallower waters, tsunami waves slow down, their wavelength being reduced up to 10 km (Fig. 2.1). As wavelength and phase velocity decrease in a similar degree, the wave period is assumed to be constant (KLUG 1986, p. 96). In contrast, height and steepness of the wave increase near the coast due to fact that the amount of energy of the wave is concentrated to a decreasing water volume (KLUG 1986, p. 96).

Besides the amount of energy input during generation as well as the weakening during propagation (by chain of islands, for instance) and the distance to the source, the final wave height of a tsunami at the shoreline mainly depends on the nearshore bathymetry (KLUG 1986, pp. 96, 97). In the case of a gently sloping seafloor, all other conditions remaining equal, tsunami wave height and phase velocity will be moderate compared to a steep seafloor owing to the increased dissipation of wave energy during propagation across wide shallow waters (BRYANT 2005, pp. 218, 219). Furthermore, the coastline's shape is of major importance concerning the nearshore wave height and phase velocity (KLUG 1986, pp. 96): Especially in marine inlets as bays, fjords or harbours, tsunami waters are concentrated and accelerated due to strong funnel effects. This results in an

increased wave height and phase velocity compared to coasts showing a straight outline. Funnel effects may even occur in the case that the inlet is not opened in travel direction of a tsunami as diffraction and refraction around promontories or harbour moles deflect the waves into them (BOLT *et al.* 1975, p. 134). Besides funnelling, tsunami waves may be amplified in marine inlets due to internal resonance effects. This happens if the inlet's eigenperiod is some harmonic of the wave period of the incoming tsunami wave (Chap. 2.3.4). The nearshore bathymetry and the shape of the coastline frequently being irregular at one and the same coast, here tsunami wave heights may vary enormously from place to place related to the above-mentioned effects (KLUG 1986, p. 97; CAMFIELD 1994, p. 178; BORMANN 2012, p. 4).

Tsunami waves do not necessarily near the coast in the form of a wave crest, but sometimes it is the wave trough that arrives first (for this paragraph see BOLT *et al.* 1975, p. 134). Such tsunami waves are often mathematically described as leading-depression *N*-waves (Chap. 2.5). In this case, the sea abruptly retreats, sometimes for many 100 m and usually several minutes before the arrival of the first wave crest, depending on the wave period. In spite of this *draw down* being an infallible sign of nearing tsunami waves, ignorant people do often follow the receding sea and therefore even approach the hazardous wave crest.

Due to frequency dispersion (Chap. 2.5), tsunamis generally do not travel as single waves but form so-called *tsunami wave trains*, comprising up to more than 10 separate waves (BRYANT 2008, pp. 27, 28; BORMANN 2012, pp. 5, 6). Depending on the wave period, up to 2 h may lie between the arrival of two consecutive wave crests at the coast, often making people go back to the sea too early after the first wave made landfall (HELAL 2010, p. 230). Rarely is the first wave crest the highest of the wave train, but mostly it “occurs among the first 10 or so” (MURTY 1977, p. 2).

Nearing the coast, tsunami waves usually form spilling breakers but do not break in a plunging way (BRYANT 2005, p. 216; Fig. 2.2). Shortly after the landfall, the so-called *run-up*—meaning the inundating tsunami waters on land—typically moves at 5 m s^{-1} to 8 m s^{-1} (BRYANT 2008, p. 34; FRITZ *et al.* 2006; HAYASHI & KOSHIMURA 2012). On rare occasions, even velocities of up to 20 m s^{-1} were observed (NANAYAMA & SHIGENO 2006, p. 153 and references therein; MORTON *et al.* 2007, p. 187). The run-up velocity can approximately be calculated as follows (CAMFIELD 1994, p. 179; BRYANT 2008, pp. 46, 47):

$$v_r = d_f^{0.7} \sqrt{\tan \beta_w} n^{-1} \quad (2.4)$$

where v_r = flow velocity of the run-up [m s^{-1}]
 d_f = flow depth or inundation depth [m]
 β_w = inclination of the water surface [$^\circ$]
 n = Manning's roughness coefficient [s].

During the run-up, the *flow depth* or *inundation depth* of tsunami waters, i.e. the vertical distance between the ground surface and the water surface, may exceed 20 m (MORTON *et al.* 2007, p. 187), but decreases with the run-up propagating on the mainland, as does the flow velocity. This is due to the increased friction on vegetated and developed land, decelerating the tsunami waters until the *backflow* or *backwash* is formed in the direction of the sea. Concentrated in channel-like depressions, the backflow velocity may be exceptionally high, not uncommonly in the order of the run-up or even higher (LE ROUX & VARGAS 2005; DAWSON & STEWART 2007a, pp. 175, 176).

An appropriate measure to determine the wave height of a tsunami is the *run-up height*, defined as the maximum height above sea level the waves reach onshore (HELAL 2010, p. 231).

2 The tsunami phenomenon



Figure 2.2: Photographic record of tsunami landfall at Laie Point on the island of Oahu, Hawaii on 9 March 1957 (photos taken by Henry Helbusch (NDGC/WDS 2014)). A typical feature is the spilling manner of the nearshore wave breaking. The tsunami was caused by an earthquake of $M_s = 8.3$ in the Aleutian Islands, some 3600 km distant.

Contrary to the true wave height of tsunami waves on the shoreline, which is hardly measurable, the run-up height can simply be estimated after the water has receded by using flood marks at buildings and vegetation or the height of transported debris (KLUG 1986, p.97). The (maximum) run-up height is calculated depending on the tsunami waveform and the manner of wave breaking. In the case of solitary tsunami waves, which are a common mathematical description of tsunami waves (Chap. 2.5), the (maximum) run-up height is defined as (SYNOLAKIS 1987, p. 529; SYNOLAKIS *et al.* 2008, p. 2202):

$$H_r = 2.831 \sqrt{\cot \beta_c} H^{\frac{5}{4}} \quad (2.5)$$

where H_r = (maximum) run-up height [m]
 β_c = inclination of the coast [$^\circ$]
 H = tsunami wave height before shoaling [m].

This relation is true, when tsunami waves do not form plunging breakers during the run-up but break in a spilling manner and when $\sqrt{H} \gg 0.288 \tan \beta_c$.

Ordinarily, the run-up height clearly exceeds the actual tsunami wave height on the shoreline. Nevertheless, both may strongly vary in neighbouring coasts (see above; CAMFIELD 1994, p. 178). Run-up heights of several decametres are no rare occurrence. A current example is the tsunami forced by the Tōhoku earthquake measuring $M_s = 8.3$ on 11 March 2011 in Japan (USGS 2012a), causing run-up heights of almost 39 m in the north-east of the country (NDGC/WDS 2015a). Furthermore, run-up heights of nearly 51 m have been recorded during the landfall of the 2004 Indian Ocean tsunami at the Indonesian west coast (NDGC/WDS 2015a), which was triggered by an earthquake of $M_s = 8.8$ (USGS 2005). In the Mediterranean, the maximum run-up of 25 m was observed on the Island of Amorgos in the Aegean Sea on 9 July 1956 (PAPAZACHOS & DIMITRIU 1991, pp. 162, 163). This tsunami was related to the combined effects of a $M_s = 7.5$ earthquake and an associated submarine landslide in the Amorgos Basin (PAPAZACHOS *et al.* 1985; PERISSORATIS & PAPADOPOULOS 1999). The maximum run-up height recorded during tsunami inundation directly caused by seismic crust movement amounts to 64 m, namely on the Peninsula of Kamchatka in 1737 (BRYANT 2005, p. 216). However, the maximum run-up height ever recorded was about 524 m in Lituya Bay in Alaska on 9 July 1958, when — forced by a nearby earthquake — a volume of about $30.5 \cdot 10^6 \text{ m}^3$ of consolidated rock was released by the precipitous mountainside in the north-eastern bay and struck the water body from an elevation of between 600 m and 900 m (FRITZ *et al.* 2009).

The area on land inundated by the run-up approximately agrees with the cross-sectional area under the tsunami wave crest close to shore (Fig. 2.3; for this paragraph see BRYANT 2008, pp. 45, 46). The same applies to the water volume. The longer the wavelength and the larger the wave height — resulting in an increased tsunami wave energy — the larger the water volume and area of inundation onshore. The maximum distance to the shoreline reached by tsunami waters onshore is indicated by the *inland penetration*, which is strongly influenced by the wave energy, the slope, the local topography as well as the density of vegetation and buildings on land. Thus, a large inland penetration is especially expected for flat coasts sparsely vegetated and developed and therefore showing low values of friction. Furthermore, funnel effects may locally increase the inland penetration, in particular when tsunami waters are concentrated and accelerated in coastal valleys. Owing to the multitude of influencing factors, the inland penetration may vary significantly at one and the same coast, as was already said for the wave height and run-up height. In extreme cases, tsunami waves may penetrate several 10 km into the mainland, but in the majority of cases penetration is much weaker. For instance, the maximum inland penetration

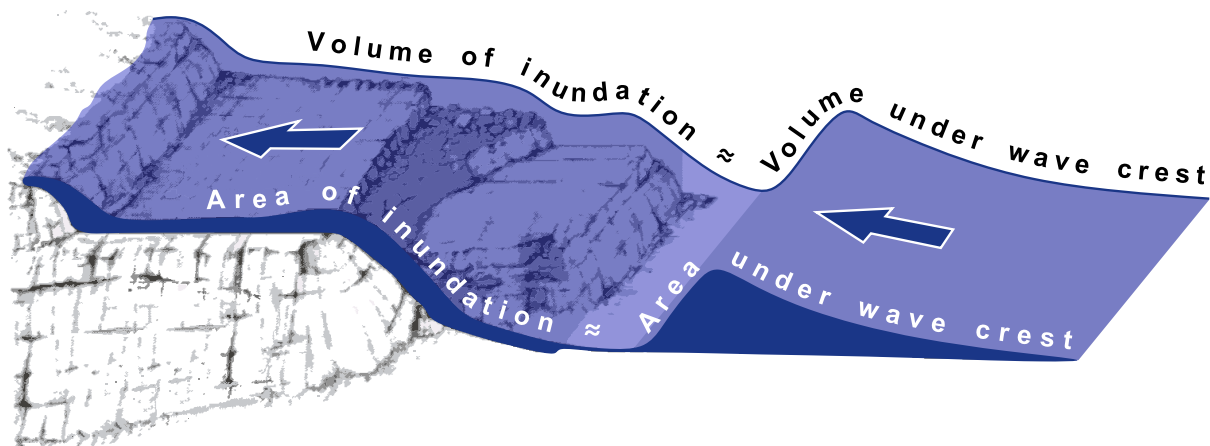


Figure 2.3: Schematic drawing showing that the cross-sectional area and the water volume under the tsunami wave crest close to shore approximately corresponds with the cross-sectional area and water volume of the run-up (after BRYANT 2008, p. 45).

reached by the 2004 Indian Ocean tsunami on the flat north-western tip of Indonesia at Banda Aceh amounted to nearly 5 km, related to a local wave height of almost 10 m.

2.3 Causes of tsunamis

2.3.1 Seismic activity

Seismicity is the leading cause of tsunamis. Based on the tsunami catalogue by the GeoForschungsZentrum Potsdam of the year 2005, almost 82.3 % of worldwide 1323 recorded tsunamis since the 17th century BC (Late Bronze Age Minoan eruption of Santorini Volcano), i.e. 1089 events, are directly ascribed to earthquakes (SCHIELEIN *et al.* 2007, pp.158–199). The same situation is found in the Mediterranean (without taking account of the Sea of Marmara and the Black Sea), where about 83.5 % of all 176 observed events are directly related to seismic activity (SCHIELEIN *et al.* 2007, pp.193–199).

Earthquakes may cause tsunami waves as follows:

- In the majority of cases, tsunami waves are excited by the sudden vertical displacement of the water column due to the vertical shift of the seafloor along faults during the focal mechanism (MURTY 1977, p.2; BRYANT 2008, pp.127–139). A downward movement of submarine crust makes the water masses dashing into the newly emerged cavity, while a huge wave trough is generated at the water surface (KLUG 1986, p.85). In return, the water column is pushed to the top, forming a huge wave crest at the water surface, when the submarine crust is shifted upwards (Fig.2.4). In both cases, not only one, but usually a series of waves is generated because of the inertia of the oscillating water particles and due to frequency dispersion (Chap.2.5).
- Rayleigh waves, i.e. the surface waves of earthquakes moving in a retrograde elliptical manner in the upper crust, cause a temporary vertical displacement of the seafloor by several decimetres (BRYANT 2008, pp.138, 139). This may provoke moderate tsunami waves especially in shallow water areas as bays, estuaries or small lakes or may intensify tsunami waves originally generated by seafloor displacement (MURTY 1977, p.2; BRYANT 2008, pp.138, 139).

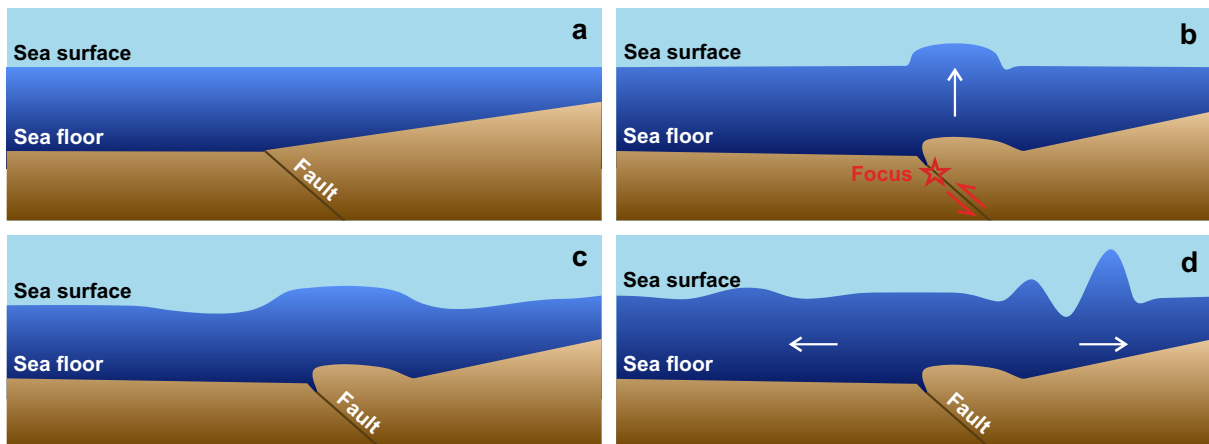


Figure 2.4: Generation of tsunami waves due to the sudden uplift of the seafloor during a submarine shallow-focus earthquake (after UNESCO-IOC 2006, p. 3).

- Finally, indirect tsunami generation by earthquakes can be due to seismogenic subaerial or submarine gravitational mass wastings. Tsunamis of this nature are not considered in the above-mentioned statistics of seismogenic tsunamis and are dealt with in detail in Chap. 2.3.2.

The following remarks concentrate on tsunami generation in connection with vertical crust movement during the focal mechanism of earthquakes, this being the most significant seismogenic triggering process of tsunamis. Only a small portion of the numerous earthquakes occurring worldwide actually cause tsunamis. This is due to the multitude of conditions, tsunamigenic earthquakes have to satisfy:

- Fundamentally, the epicentre of earthquakes must be located in or by the sea, otherwise, the energy of the crust movement cannot be transferred to the water body (KLUG 1986, p. 84).
- For an effective transfer of seismic energy to the water column, the hypocentre of earthquakes must be situated in shallow crust depth (KLUG 1986, pp. 84, 85). According to BRYANT (2008, p. 129), most seismogenic tsunamis are caused by shallow-focus earthquakes at depths between 0 km and 40 km.
- The focal mechanism of earthquakes must be accompanied with a vertical shift of the seafloor. This is the case for earthquakes along dip-slip and oblique-slip faults due to tectonic compression or extension (BRYANT 2008, pp. 135, 136). In contrast, strike-slip faults showing no vertical component of movement do not favour tsunamigenic earthquakes, even if the earthquake's magnitude is high (KLUG 1986, p. 85). For instance, no tsunami was generated during the strong earthquake of San Francisco in 1906 ($M_s = 7.7$; USGS 2012b), although crust segments along the San Andreas Fault, partly running beneath the sea, slipped horizontally by up to 6 m (BOLT *et al.* 1975, p. 137). Earthquakes along shallow dipping thrust faults (dip values between 20° and 30°) are most efficient at generating tsunami waves due to their geometry favouring large vertical crust movements (BRYANT 2008, pp. 135–138; KOPP & WEINREIBE 2009, pp. 21, 22). As shallow dipping thrust faults are most common in the area of subduction zones — basically located beneath the sea and showing a high seismic activity — here, most tsunamis are generated worldwide (BRYANT

2 The tsunami phenomenon

2008, pp. 135–139; KOPP & WEINREIBE 2009, p. 21). Consequently, the highest tsunami-genic potential can be found in the Pacific—the seemingly *peaceful* ocean—, which is enclosed by numerous subduction zones (*Ring of Fire*). In total, 52.9% of all tsunamis recorded worldwide occurred in the Pacific Ocean (BRYANT 2008, p. 14).

- An appropriate fault geometry provided, a minimum earthquake magnitude of $M_s = 6.5$ is generally believed to generate tsunami waves (e.g. BRYANT 2005, p. 219). Of course, this is not an accurate threshold and only gives a vague classification because the generation of tsunami waves does not only depend on the earthquake magnitude but also on the rupture velocity. Usually, the focal mechanism of earthquakes lasts for tenths of seconds, its rupture velocity ranging from 1 km s^{-1} to 10 km s^{-1} , too fast for the water body to respond entirely to this movement (TODOROVSKA & TRIFUNAC 2001, pp. 151–154). Consequently, the impulse can only partly be transmitted to the water. However, there are earthquakes with slower rupture velocities in the order of 0.1 km s^{-1} to 1 km s^{-1} and longer focal durations that may exceed 100s. During so-called *tsunami earthquakes*, the seismic moment is transmitted to the water body more effectively (KANAMORI 1972). This is especially the case, if the rupture velocity roughly corresponds to the local shallow water wave velocity, which depends on the water depth (Eq. 2.2; TODOROVSKA & TRIFUNAC 2001, p. 152). On the one hand, this may lead to tsunami formation even if the earthquake’s magnitude is below the threshold of $M_s = 6.5$. On the other hand, earthquakes assigned $M_s \geq 6.5$ may generate stronger tsunami waves than expected from their magnitude.

Besides the rupture velocity, the initial height of seismogenic tsunamis mainly depends on the mean vertical slip of the seafloor and on the area of displacement (BRYANT 2008, p. 136). With a rise in vertical slip, the initial wave height increases (DAWSON & STEWART 2007a, p. 168). The same is true for the area of seafloor displacement, as “the water has less opportunity to flow sideways” (TODOROVSKA & TRIFUNAC 2001, p. 152) the larger the area shifted. As a rule, the initial tsunami wave height is proportional to the cube of the length of the rupture zone (BRYANT 2008, p. 136), which, in turn, is positively correlated to the area of displacement. According to DAWSON & STEWART (2007a, p. 168), the initial disturbance of the sea surface of seismogenic tsunamis does generally not exceed the vertical slip of the focal mechanism (up to several metres). With increasing distance to the hypocentre, seismogenic tsunami waves rapidly flatten and do often cross the ocean without being noticed, until they near the coast (Fig. 2.1).

By tendency, the mean vertical slip and the area of displacement during an earthquake are proportional to the earthquake magnitude (BOLT 2003, pp. 168–170; GROTZINGER *et al.* 2008, p. 341). Consequently, the earthquake magnitude positively correlates with the tsunami intensity (BRYANT 2005, pp. 219, 220; DAWSON & STEWART 2007a, p. 168). This is expressed by the relationship between the earthquake Richter magnitude scale and the tsunami Imamura-Iida magnitude scale (Tab. 2.1). The Imamura-Iida scale was derived from a circa 100 tsunamis recorded at the Japanese coast between 1700 and 1960, using the following equation (IIDA 1963):

$$m_{II} = \log_2 H_r \quad (2.6)$$

where m_{II} = tsunami Imamura-Iida magnitude scale (dimensionless).

Thus, the tsunami magnitude, ranging from -2 to 5 , is derived only from the maximum run-up height, which, in turn, can be linked to the earthquake Richter magnitude. As both magnitude scales are logarithmic scales, the maximum run-up height grows disproportionately with increasing tsunami and earthquake magnitude (Tab. 2.1). Although the Imamura-Iida scale has become

Table 2.1: Relationship between earthquake Richter magnitude, tsunami Imamura-Iida magnitude and maximum tsunami run-up height, based on the analysis of about 100 tsunamis recorded in Japan between 1700 and 1960 (after IIDA 1963).

Earthquake magnitude	Tsunami Imamura- Iida magnitude	Maximum run-up height [m]
6.0	-2	< 0.3
6.5	-1	0.5–0.75
7.0	0	1.0–1.5
7.5	1	2.0–3.0
8.0	2	4.0–6.0
8.3	3	8.0–12.0
8.5	4	16.0–24.0
8.8	5	> 32.0

very common, it has been refined several times, for instance by SOLOVIEV (1970) considering not the maximum but the mean run-up height along a stretch of coast, and by ABE (1979), who added a specific coefficient to the equation.

Tsunami waves principally propagate at a right angle to the fault line of an earthquake, i.e. into two opposite directions (BRYANT 2008, pp.136, 137). The longer the fault line—which may amount to several 100 km—and the slower the rupture velocity, the more intense is the elongated beaming of wave power into these two directions (DAWSON & STEWART 2007a, p.169; BRYANT 2008, p.137). This explains the high vulnerability of the Hawaiian Islands to teletsunamis because most subduction zones in the Pacific Ocean are located right-angled to the archipelago (BOLT *et al.* 1975, p.137).

There are two prominent examples of recent extraordinary tsunami events of seismogenic origin—the 2004 Indian Ocean tsunami and the 2011 Tōhoku tsunami. The Tōhoku tsunami was caused by a megathrust earthquake measuring $M_s = 8.3$ some 130 km offshore Sendai in north-eastern Japan on 11 March 2011 (HWANG 2014, p.20). Tsunami formation was forced by the shallow focus depth (about 30 km) and the widespread seafloor displacement (nearly 45 000 km²) by up to 9.7 m (LIN *et al.* 2012, pp.126, 127; USGS 2013). Additionally, the rupture velocity of the source mechanism was quite low (1.5 km s⁻¹ to 2 km s⁻¹; HWANG 2014, p.25). This altogether caused high-amplitude tsunami waves with maximum run-up heights of almost 39 m and a maximum inland penetration of about 5 km at the east-northeast coast of Honshu Island (ESTEBAN *et al.* 2013, pp.89–91; NDGC/WDS 2015a). In total, about 17 128 people fell victim to this event (NDGC/WDS 2015a).

The 2004 Indian Ocean tsunami was even more devastating, killing an estimated 226 898 people (NDGC/WDS 2015a). The tsunami was triggered by an earthquake $M_s = 8.8$ about 250 km south-southeast offshore Banda Aceh at the northern tip of Sumatra (USGS 2005). The hypocentre was located in 30 km depth. During the earthquake, a seafloor area of approximately 100 000 km² was shifted by averagely 3 m along a rupture length of about 1000 km. Thereby, a total energy of almost 4.5×10^{15} J was transferred to the water body, the energy produced by an ordinary nuclear power plant during one month (LAUTERJUNG *et al.* 2009, p.39). This explains the enormous power of the generated tsunami waves, penetrating more than 5 km inland and reaching onshore elevations of up to 50.9 m in the north-west of Sumatra Island (NDGC/WDS 2015a; NDGC/WDS 2015b).

Finally, an extraordinary seismogenic tsunami is well known in the history of the Mediterranean world. On 21 July 365 AD, a strong earthquake with an estimated magnitude of $M_s = 8.3$

occurred south-west off the Island of Crete, which probably resulted in a vertical shift of the western part of the island by almost 9 m (STIROS 2009, pp. 480–483; PIRAZZOLI *et al.* 1992; PIRAZZOLI *et al.* 1996, pp. 6085–6092; STIROS 2010, pp. 56, 57). Since the surrounding seafloor was involved in this movement, a teletsunami was triggered, likely to have devastated the entire eastern and central Mediterranean (cf. STIROS 2001; STIROS & DRAKOS 2006; SHAW *et al.* 2008; VÖTT *et al.* 2009a; MAY *et al.* 2012b; SAMARAS *et al.* 2015).

2.3.2 Submarine and subaerial gravitative mass wastings

Tsunamis driven by gravitative mass wastings are less frequent than those generated by seismicity. According to SCHIELEIN *et al.* (2007, pp. 158–160), 116 of all 1323 recorded tsunamis, i.e. 8.8 %, are associated with mass wastings, 48 of them induced by earthquakes. An even smaller proportion of 5.7 % (10 of total 176 tsunamis) can be concluded for the Mediterranean regardless of the Black Sea and the Sea of Marmara (SCHIELEIN *et al.* 2007, pp. 193–199).

Subaerial mass wastings as rockfalls or landslides may cause tsunami waves when striking coastal waters. In this case, water is displaced by the entering material itself and by the air trapped between the material and the water surface (KLUG 1986, p. 89; DAWSON & STEWART 2007a, p. 170). Despite this tandem mechanism, only 4 % of the mass wasting’s energy is transformed into tsunami wave energy, explaining the weak effects such waves have in general (BRYANT 2008, p. 189).

Submarine gravitative mass wastings are more efficient in generating tsunamis significant in the far-field (Fig. 2.5; for this paragraph see BRYANT 2008, pp. 179–184). This is due to the more effective energy transfer to the water body and the larger mass and volume usually moved within submarine mass wastings. The latter is particularly true concerning submarine landslides, i.e. unconsolidated material sliding downslope in closed units, which may comprise some 1000 km³ up to 20 000 km³. The large volume of submarine compared to subaerial mass wastings is due to the seafloor vastly being covered by unconsolidated sediment. This in combination with the absence of vegetation, the increased pore water pressure and the high relief energy along continental margins, deep-sea trenches, mid-ocean ridges, volcanoes, seamounts or guyots make gravitative mass wastings likely to occur in the marine environment. Besides unconsolidated material, submarine mass wastings are also associated with submarine bedrock, especially when stratified parallel to the slope and showing alternating layers permeable and impermeable to water.

There is a variety of mechanisms making unconsolidated sediments or bedrock to destabilise and therefore to cause submarine gravitative mass wastings:

- Most submarine mass wastings are due to the kinetic energy of seismic surface waves or crust dislocation during earthquakes (BRYANT 2008, p. 180). Moreover, seismic surface waves temporarily increase the sediment pore water pressure, which may result in the destabilisation of the slope when a critical threshold is passed (NIEDORODA *et al.* 2007, p. 112). Actually, it is not always possible to ascertain whether tsunami waves are caused by an earthquake or an associated submarine mass wasting or by both of it.
- Besides seismic surface waves, storm surges can critically increase the sediment pore water pressure and thus destabilise the seafloor. This is due to the temporary rise in water level during storm surges, which may increase the weight on the continental shelf by several 1 000 000 t km⁻² (BRYANT 2008, p. 183). Furthermore, this may even cause tectonic stress to be released, sufficient to trigger an earthquake, which, in turn, can generate a tsunami or a tsunamigenic submarine landslide (e.g. MOROZOVA 2006; LIU *et al.* 2009). A popular

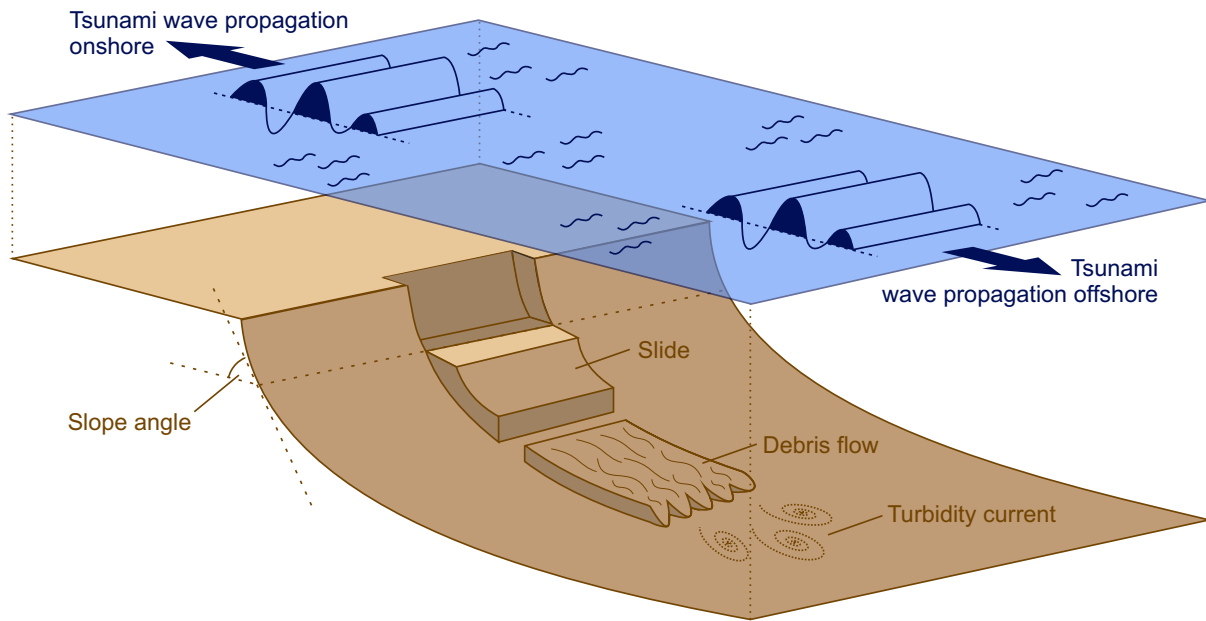


Figure 2.5: Generation of a tsunami and turbidity current caused by a submarine landslide (after BRYANT 2008, p. 185 and TEN BRINK *et al.* 2014, p. 43). Note that most of the wave energy is focused in two opposite directions along the axis of the slide, which is in contrast to the elongated beaming usually found for seismogenic tsunamis.

example for this is the $M_s = 7.9$ Great Kantō earthquake of 1 September 1923 in the Tokyo area, which was probably induced by a strong typhoon making landfall on the eastern Japanese coast at the same time (BRYANT 2008, p. 183; USGS 2012c). The earthquake caused a submarine landslide and destructive tsunami waves with a maximum run-up of 13 m, altogether killing an estimated 142 807 people (NDGC/WDS 2015a).

- Ongoing sediment accumulation can trigger submarine mass wastings if a slope becomes overloaded and oversteepened (NIEDORODA *et al.* 2007, p. 112). This mechanism is most significant for delta areas, where high rates of sedimentation are common (BRYANT 2008, p. 183).
- Submarine mass wastings are known to be caused by erosion of submarine slopes, especially in submarine canyons, forced by currents (COSTA *et al.* 2005, p. 153).
- Volcanic eruptions may induce vast subaerial and submarine landslides or debris avalanches on volcano flanks or even cause complete edifice collapse (caldera formation). This is apparent from huge sliding masses of several 1000 km^3 found at the base of volcanic islands all over the world, for instance in the Canary Islands, Cape Verde, Mascarene Islands or the Hawaiian Islands (e.g. LIPMAN *et al.* 1988; MOORE *et al.* 1994; MASSON 1996; KEATING & MCGUIRE 2000; KOPP & WEINREIBE 2009, pp. 23, 24). Furthermore, repetitive lava flows or deposition of pyroclastic material may oversteepen volcano flanks by time, leading to failure as well (BRYANT 2008, p. 184).
- According to BRYANT (2008, p. 183), gases as methane enclosed in marine sediments can weaken the sediment cohesion and therefore trigger submarine mass wastings. Gases are mainly produced in the continental shelf and slope area, especially adjacent to river deltas due to the anaerobic decay of terrestrial organic material covered by sediment. If the in-

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ternal sediment pressure raises above the local hydrostatic pressure, the sediment becomes unstable leading to failure. In greater depth, solid gas hydrates are enclosed in the sediment due to low temperatures and the high hydrostatic pressure. Gas hydrates, however, are liable to transform into gas, for instance in the case of rising water temperatures, which again may result in a critical increase of the internal sediment pressure and slope failure. In addition, large volumes of gas suddenly released upwards may cause a so-called *blow out*, capable to generate tsunami waves by displacing a large water volume (SCHEFFERS & KELLETAT 2006, p. 3).

- Lastly, anthropogenically induced mass wastings, attributable, for instance to explosion, detonation or engineering work, have to be mentioned. A prime example is the submarine mass wasting triggered by the refill of an excavation at the airport of Nice (French Riviera) on 16 October 1979. The mass wasting caused tsunami waves 3 m high at the coast of Antibes, some 10 km distant in a south-western direction (ASSIER-RZADKIEWICZ *et al.* 2000).

According to PELINOVSKY & POPLAVSKY (1996, p. 15), tsunami generation by submarine landslides — the most significant trigger among mass wastings — mainly depends on the volume and the mass of the moving material as well as the slide velocity, the slope angle and the water depth (Fig. 2.5). In general, a rise in volume and mass means an increase in tsunami wave height and energy (WARD 2001, p. 11 203). A more complex correlation holds true for the sliding velocity (WARD 2001, pp. 11 203, 11 204; DAWSON & STEWART 2007a, p. 169): Slow submarine landslides are likely to generate weak or no tsunamis, while very fast ones cause tsunamis with wave heights in the order of the slide’s thickness. However, tsunamis of exceptional size are generated in phase, when the slide velocity approximates the local tsunami wave velocity depending on the water depth (cf. rupture velocity in Chap. 2.3.1). The velocity of the slide itself is positively correlated to the mass and density of the moving material and the slope angle (BRYANT 2008, p. 185). If $\lambda \gg d$, the maximum initial wave height of slide-induced tsunamis in the source region can be approximated using the following equation (PELINOVSKY & POPLAVSKY 1996, p. 15):

$$H_{max} = \frac{\pi a^2}{4 d_i} \quad (2.7)$$

where H_{max} = maximum initial tsunami wave height above sea level [m]
 a = thickness of submarine landslide [m]
 d_i = initial water depth [m].

As it takes some time for the slide to fully develop and to reach its final velocity, there is a significant gap between the landslide and tsunami generation (BRYANT 2008, p. 185), which is different from tsunamis directly caused by earthquakes. In the early stage, a submarine landslide moves downslope in coherent blocks but usually disintegrates by time forming a debris flow (Fig. 2.5; BRYANT 2008, pp. 179–186). If the debris flow mixes and incorporates water, a turbidity current emerges, leaving a specific depositional signature at the seafloor — a so-called *turbidite*. Consequently, turbidites give evidence of former submarine landslides and by this of possibly related tsunamis. Turbidites are further indicative of palaeotsunami events because turbidity currents can directly be caused by tsunami waves. For example, the Augias megaturbidite described from the Ionian and Syrtic Abyssal Plain in the central Mediterranean Sea documents a huge turbidity current, that might had been triggered by tsunami waves caused

by a debris avalanche on the flanks of Mount Etna on Sicily (PARESCHI *et al.* 2006a,b), by the tsunami associated with the Minoan Eruption of Santorini (KASTENS & CITA 1981; HIEKE & WERNER 2000; PARESCHI *et al.* 2006c; Chap. 2.3.3) or by the AD 365 Cretan earthquake and resultant tsunami (POLONIA *et al.* 2013).

Whereas turbidity currents are irrelevant in view of tsunami generation, submarine landslides or debris flows cause a significant vertical displacement of the water body, which is crucial for tsunami wave excitation (BRYANT 2008, p. 185). Upslope, at the back of the mass wasting, water is rushing into the cavity constantly built by the moving material forming a depression at the water surface, whereas water is pushed up by the accelerated sediment at the front (KLUG 1986, p. 89; MCMURTRY *et al.* 2004, p. 221). This causes a significant dipole character and an *N*-waveform (Chap. 2.5) of slide-generated tsunamis (OKAL & SYNOLAKIS 2004, pp. 901, 902; DAWSON & STEWART 2007a, p. 169). Trough and crest radiate from the source area, but most wave energy is focused in two opposite directions along the axis of the mass wasting (Fig. 2.5). This effect becomes strongest when the slide velocity approximates the local tsunami wave velocity (WARD 2001, p. 11 203). The focused character of slide-induced tsunamis is in contrast to the elongated beaming usually found for seismogenic tsunamis (Chap. 2.3.1). Furthermore, wavelength and wave period are shorter in the case of landslide tsunamis (BRYANT 2008, p. 185). The wave period increases the larger the mass wasting and the smaller the slope angle but is independent from water depth and the mass of the moving material (WIEGEL 1955, p. 764, 771; WATTS 1998, p. 135).

A prominent example of a submarine landslide causing a teletsunami is the Storegga slide on the continental slope off Norway, nearly 8000 years ago (e.g. BUGGE *et al.* 1988; CANALS *et al.* 2004; HAFLIDASON *et al.* 2004, 2005). The Storegga slide is one of at least 5 huge landslides that occurred along the Norwegian continental slope during the past 30 000 years and was probably connected to seismic activity or decomposition of gas hydrates in slope sediments. The volume of the Storegga slide was about 2400 km^3 to 3200 km^3 with an average thickness of 114 m and a maximum reach of more than 800 km. According to numerical simulations (HENRY & MURTY 1992; HARBITZ 1992; BONDEVIK *et al.* 2005a; HILL *et al.* 2014), the slide triggered tsunami waves that spread across the entire Norwegian Sea, the North Sea and also reached the north-eastern Atlantic. The initial wave height in the source area was more than 20 m, which closely agrees with the outcome of Eq. 2.7 of approximately 20.4 m, assuming an average initial water depth of about 500 m. Reaching the North Sea and the open Atlantic, tsunami waves flattened down to few metres. Simulations revealed maximum run-up heights of 5.3 m to 18 m in eastern and northern Scotland respectively, 13.7 m in eastern Iceland, 11.5 m in eastern Greenland and up to 15 m in western Norway, which is in good accordance with tsunami deposits found in corresponding elevations onshore (LONG *et al.* 1989; DAWSON *et al.* 1993; BONDEVIK *et al.* 1997a,b, 2005a,b).

A more recent example of slide-induced tsunamis is known from the Grand Banks of Newfoundland in November 1929 (e.g. LONG *et al.* 1989, pp. 64, 65; BENT 1995; PIPER *et al.* 1999; FINE *et al.* 2005; MOORE *et al.* 2007). Due to an earthquake on the continental slope off Newfoundland measuring $M_s = 7.2$, hundreds of rotational slides were triggered with a total volume of almost 200 m^3 . As a result, three major tsunami waves, 3 m to 8 m high, were generated, which ran up the nearby coast of southern Newfoundland by maximum 13 m and killed 28 people. Although there were no fatalities beyond Newfoundland, tsunami waves spread along the North American east coast and were even recorded in South Carolina and along the Portuguese coast 4 h to 6 h after tsunami generation.

Also subaerial mass wastings are known to have triggered significant tsunami waves, although not significant in the far-field. The rockslide in the north-eastern Lituya Bay in Alaska on 9 July

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1958 is a prominent example (e.g. PARARAS-CARAYANNIS 1999; FRITZ *et al.* 2009; WEISS *et al.* 2009): The rockslide was caused by a nearby earthquake with a surface magnitude (M_s) of between 7.9 and 8.3. A total volume of almost $30.5 \times 10^6 \text{ m}^3$ was released into the bay from the adjacent precipitous mountainside at an elevation of between 600 m and 900 m. Comparable to a meteorite impact, consolidated rock struck the water surface with an estimated velocity of 110 m s^{-1} and together with the trapped air triggered a huge solitary-like tsunami wave, which ran up the opposite mountainside by about 524 m. Due to the shallow water, tsunami wave heights rapidly decreased in the bay causing run-ups of only 11 m at the seaward entrance 11 km distant. Related to the distinct relief and high seismic activity, Lituya Bay is generally prone to slide-generated tsunamis, which is demonstrated by two former large rockslides in 1936 and 1853, when tsunami run-up heights of 150 m and 120 m were recorded.

Landslide tsunamis are also known from the Mediterranean, which is particularly prone to gravitative mass wastings due to the distinct on- and offshore topography as well as the high seismic and volcanic activity in the central and eastern part (FERENTINOS 1991, pp. 266, 267; FERENTINOS 1992). Huge mass wastings, for instance, occurred on the slopes of Mount Etna on Sicily in southern Italy (e.g. PARESCHI *et al.* 2006a,b). Here, an extraordinary subaerial and submarine debris avalanche with a total volume of almost 20 km^3 to 25 km^3 was caused by volcanic activity in the early Holocene, which probably caused a devastating tsunami in the entire central and eastern Mediterranean. More recently, on 9 July 1956, the combined effects of an earthquake and associated submarine landslide in the Amorgos Basin (Aegean Sea) caused tsunami waves that ran up the nearby Island of Amorgos by almost 25 m (cf. Chap. 2.2; PAPAACHOS *et al.* 1985; PAPAACHOS & DIMITRIU 1991, pp. 162, 163; PERISSORATIS & PAPADOPOULOS 1999).

2.3.3 Volcanism

With a percentage of 4.8 % only ($\cong 64$ events), volcano-induced tsunamis are rare occurrence in the world's oceans (SCHIELEIN *et al.* 2007, pp. 158, 159). This is also true for the Mediterranean (regardless the Black Sea and the Sea of Marmara), where 5.7 %, i.e. 10 tsunamis, are found in connection with volcanism (SCHIELEIN *et al.* 2007, pp. 193–199). There are various mechanisms causing volcanogenic tsunamis, sometimes even in combination (BRYANT 2008, pp. 217–221; DAWSON & STEWART 2007a, p. 169):

- Volcanic earthquakes during eruption are most significant in generating tsunami waves. Here, the same mechanisms apply as described for earthquakes in general in Chap. 2.3.1.
- Pyroclastic flows, i.e. dense dispersions out of ejected hot gas, lava and clasts streaming downslope with up to 55 m s^{-1} , tend to cause tsunamis when striking the water body. This holds particularly true, when the density of the flow is higher than of the water, causing a huge water volume to be displaced. Due to the high velocity, pyroclastic flows may trigger tsunami waves even in great distance to their origin. This was demonstrated, e.g. in Indonesia in 1815, when a huge pyroclastic flow of Mount Tambora travelled 15 km onshore before it reached the sea and caused wave run-ups of 4 m. Besides pyroclastic flows, tsunamigenic mass wastings on volcano flanks may also originate from sediment or bedrock destabilisation during eruption. Such mass wastings comprise subaerial and/or submarine debris avalanches, debris flows, landslides (Chap. 2.3.2) or so-called lahars, which are mud or debris flows made out of waterlogged volcanic material as ash and clasts.
- Occasionally, submarine volcanic explosions trigger tsunamis. This mechanism is restricted to shallow waters less than 500 m deep. In greater depth, explosions are suppressed due

to the increased hydrostatic pressure, making volcanic activity more effusive. Submarine volcanic explosions are caused when seawater gets in contact with the magma chamber and quickly evaporates. Water vapour rises up and displaces the surrounding water, which may result in significant tsunami wave excitation.

- During eruption, volcanoes loose substantial amount of magma, leading to failure and collapse of the magma chamber with time (caldera formation), especially in the case of stratovolcanoes. As this causes water to rush into the newly emerged cavity, submarine caldera formation is often associated with tsunami generation.
- Finally, tsunamis are found in connection with atmospheric pressure waves that are released within strong volcanic explosions. The relevant trigger mechanism is the same as described for meteotsunamis in Chap. 2.3.4. Volcanic atmospheric pressure waves only cause weak tsunami waves but even thousands of kilometres distant to the volcano. This was observed, e.g. during the final explosion of Mount Krakatau in 1883 (see below).

Volcanogenic tsunamis may be devastating in the near-field but usually do not affect more remote areas, which is due to their limited point source (KLUG 1986, p. 87). Nevertheless, few extraordinary eruptions are known that have caused far-reaching teletsunamis attributable to numerous fatalities. The most prominent examples are the eruption of Mount Krakatau in 1883 and the Minoan Eruption of Santorini in the 17th century BC, two of the strongest eruptions ever experienced by humanity.

The occurrences of Mount Krakatau are well documented in written accounts (e.g. VERBEEK 1884; LATTER 1981; SELF & RAMPINO 1981; NOMANBHOY & SATAKE 1995; CAREY *et al.* 2001; CHOI *et al.* 2003a; BRYANT 2005, pp. 241–243; MAENO & IMAMURA 2011), which is unique in view of volcano-induced tsunamis. After several months of increased activity, the main eruption of the volcano took place on the morning of 27 August 1883 in the form of three strong explosions, which all were accompanied by caldera formation and tsunami waves. The most horrific tsunami was caused by the third and strongest explosion, that formed a huge caldera with a diameter of 6 km and a depth of 270 m. The tsunami devastated the entire Sunda Strait with all its villages and major towns and killed more than 36 000 people. Run-up heights vastly exceeded 20 m with a maximum of 42 m. In coastal lowlands, tsunami waters penetrated 5 km inland. In north-eastern Java, coral boulders of up to 600 t were carried onshore. The tsunami was recorded in the entire Indian Ocean. Four hours after the final explosion and some 2000 km distant, tsunami waves ran up the north-west coast of Australia by 4 m. Another five hours later, the tsunami reached the Ganges River near Calcutta, 4000 km away, where it sunk 300 boats. Numerical simulations revealed that caldera formation within the third explosion was not the only trigger mechanism of the tsunami because wave heights are clearly underestimated in this scenario. Thus, the combination of caldera formation, pyroclastic flow and submarine explosion that occurred altogether during the third explosion is hold responsible for the remote effects of the resulting tsunami. Due to the strong atmospheric pressure wave that was released within the final explosion and travelled around the world seven times, weak tsunamis were excited worldwide. Within the next 37 hours, tsunami wave heights of between 0.1 m and 0.5 m were recorded in the Bay of San Francisco and Tokyo, in Honolulu, in the English Channel and—even separated from the oceans—in Lake Taupo, in the centre of the North Island of New Zealand.

About 3500 years before Mount Krakatau, an extraordinary tsunamigenic eruption struck the Mediterranean world—the Minoan eruption of Santorini in the 17th century BC (e.g. YOKOYAMA 1978; PICHLER & FRIEDRICH 1980; KASTENS & CITA 1981; HIEKE & WERNER

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2000; BRYANT 2005, pp. 238–240; DRUITT 2014; PERÍÁÑEZ & ABRIL 2014a). Santorini is one of several volcanoes of the volcanic island chain in the southern Aegean and was home to the Minoan civilisation centred on the nearby island of Crete. The Minoan eruption has been linked to the ten plagues of Egypt described in the Bible and is assumed to have introduced the gradual decline of the Minoans 50 to 100 years later. With a duration of probably four days, the eruption of Santorini went through several distinct phases. Beginning with Plinian eruptions, the volcano ejected a huge volume of tephra and pumice, followed by multiple pyroclastic flows and lahars, which probably caused locally destructive tsunamis. Sea water increasingly made contact with the magma chamber, finally causing a phreatomagmatic explosion and the formation of a huge caldera. The resulting tsunami had considerable dimensions. According to numerical simulations, initial wave heights ranged between 46 m and 68 m. This closely agrees with sea-borne pumice deposits nowadays found up to 60 m above the contemporary sea level on the neighbouring island of Anafi, some 20 km to the east. About 30 min after the final explosion, tsunami waves with an estimated height of approximately 11 m reached the island of Crete. As the explosion and caldera formation particularly affected the western part of Santorini, most of the wave energy was focused westwards into the central Mediterranean Sea. Here, the increased hydrostatic pressure and oscillatory water movement associated with the passage of the tsunami is assumed to have triggered a huge turbidity current in the Gulf of Sidra, which might have resulted in the deposition of the Augias megaturbidite described from the Ionian and Syrté Abyssal Plain in the central Mediterranean Sea (Chap. 2.3.2). Although most of the wave energy was beamed in a western direction, calculation results and field evidence indicate considerable wave run-up heights of at least 7 m even along the Turkish and Levantine coasts.

2.3.4 Atmospheric disturbances

The preceding chapter briefly dealt with the generation of tsunamis due to volcanic atmospheric pressure waves. Such tsunamis associated with atmospheric disturbances are generally called *meteotsunamis*. They show the same wave physics as tsunamis related to any other mechanism mentioned so far but are less energetic and only locally destructive in principle (MONSERRAT *et al.* 2006, p. 1041). Meteotsunamis are caused by small-scale atmospheric disturbances approaching the coast from the sea (for this paragraph see MONSERRAT *et al.* 2006, pp. 1037, 1041, 1042; ŠEPIĆ *et al.* 2009, p. 1; GEIST *et al.* 2014, pp. 124, 125). Such disturbances are atmospheric fields or waves in the troposphere causing abrupt pressure fluctuations, usually of only few hectopascals. Whereas volcanogenic atmospheric disturbances are rare occurrence, they appear ubiquitously in the form of fronts, gravity waves (e.g. undulating lee waves after the passage of mountain ridges), shelf clouds or tornadoes. According to the inverse barometer effect (change in water level by 1 cm per 1 hPa), the passage of an atmospheric disturbance over the sea is accompanied by a corresponding fall or rise in water level by several centimetres, forming a very shallow and long water wave that travels in the same direction and with the same velocity as the atmospheric disturbance (Fig. 2.6). If the celerity of this water wave is the same as the celerity of general shallow water waves in the area (depending on the local water depth according to Eq. 2.2) or if both show a similar wave period and/or wavelength, then so-called external resonance effects may increase wave heights by more than factor 10 (Fig. 2.6). External resonance effects only appear in wide shallow water areas approximately 40 m to 160 m deep because here, shallow water waves show the same celerity as typical atmospheric and associated water level disturbances (mostly 20 m s^{-1} to 40 m s^{-1}).

External resonance effects by itself are not sufficient to generate significant tsunamis (for this paragraph see MONSERRAT *et al.* 2006, pp. 1037–1040; ŠEPIĆ *et al.* 2009, p. 1; GEIST *et al.* 2014, pp. 124, 125). Even if there is an amplification by factor 10, an atmospheric disturbance

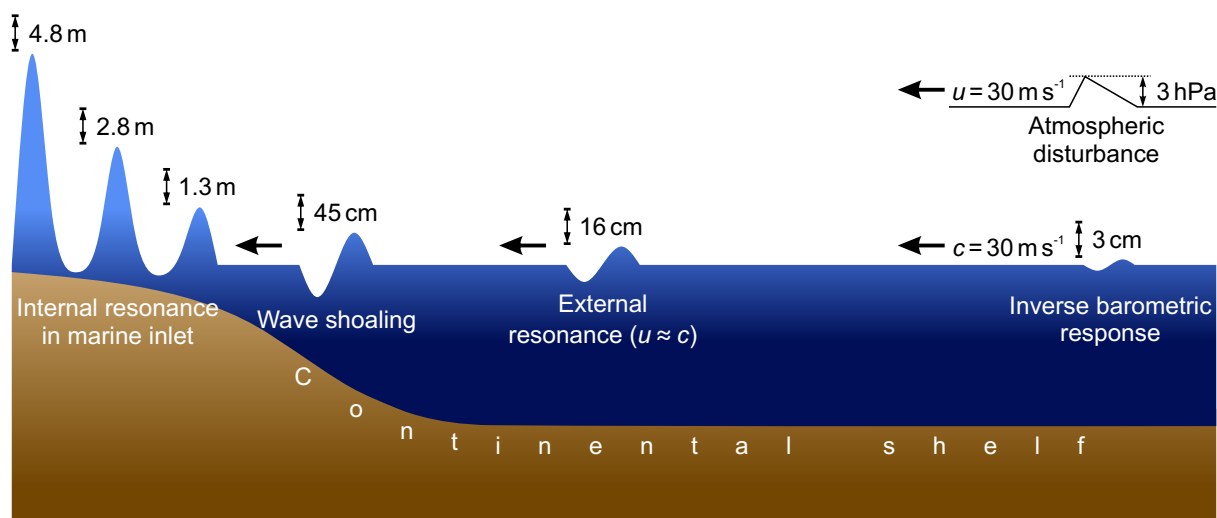


Figure 2.6: Stepwise generation of a meteotsunami in the continental shelf and a marine inlet associated with the passage of an atmospheric disturbance (e.g. coldfront) of few hectopascals moving with the velocity u (after MONSERRAT *et al.* 2006, p. 1039). The diagram is based on a meteotsunami that occurred in the Nagasaki Bay in south-west Japan on 31 March 1979. Meteotsunamis are closely connected with the appearance of external and internal resonance effects as well as wave shoaling and funnelling.

of, for instance 4 hPa would result in maximum 0.4 m high tsunami waves only. Consequently, further, so-called internal resonance effects are required to produce extraordinary wave heights. Such effects may appear when water waves associated with the atmospheric disturbance enter narrow bays, harbours, fjords or other inlets. If the eigenperiod of the inlet (depending on the water depth and shape) approximates or is a multiple of the period of entering water waves already amplified by external resonance effects, a further increase in wave height by several times is possible (Fig. 2.6). Wave shoaling due to decreasing water depth and funnelling further amplifies the meteotsunami in the inlet, finally causing wave heights of more than 5 m at the shore.

Due to the required double resonance, meteotsunamis are a phenomenon only observed in certain marine inlets fulfilling a couple of conditions (MONSERRAT *et al.* 2006, pp. 1038, 1039). On the one hand, inlets must be surrounded by wide continental shelf waters to allow external resonance effects. On the other hand, inlets are required to be narrow in shape and to be opened in the direction atmospheric disturbances and associated water waves use to travel in order to cause internal resonances. If all conditions are met, it is a question of time for the next significant atmospheric disturbance to generate a meteotsunami. In regions frequently struck by tropical or extratropical cyclones, meteotsunamis occasionally coincide with storm surges, causing even stronger destruction on coasts (VENNELL 2010; THIEBAUT & VENNELL 2011; DRAGANI *et al.* 2014). This is due to such cyclones often being associated with atmospheric disturbances, especially in the form of cold fronts.

Meteorologically induced tsunamis have been observed worldwide (VILIBIĆ *et al.* 2014a), which caused numerous local names to have emerged. For example, meteotsunamis in the Balearic Islands and in New Zealand are referred to as *rissaga*, in Sicily *marubbio*, along the Dalmatian coast *šćiga*, in the German North Sea and Baltic Sea *Seebär*, in the United Kingdom and Ireland *death waves*, in Madeira *lavadiads*, in the Azorean Islands *inchas* and in Japan *abiki* or *yota* (RABINOVICH *et al.* 2009, p. 891; NEWIG & KELLETAT 2011). A prominent example of a destructive meteotsunami is known from Nagasaki Bay in south-west Japan of 31 March

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1979 (Fig. 2.6; MONSERRAT *et al.* 2006, pp. 1038–1040). The meteotsunami was triggered by an atmospheric disturbance that moved over the East China Sea in an almost eastern direction towards the bay at a velocity of about 31 m s^{-1} . The disturbance was associated with a sudden rise and gradual fall in air pressure by 3 hPa over a horizontal scale of almost 200 km, causing an inverted barometer response of the water level by 0.03 m and a corresponding long water wave. Owing to the local water depth of 50 m to 150 m, the velocity of the atmospheric disturbance and associated water wave approximated the celerity of shallow water waves in this area (external resonance) resulting in an amplification of the initial wave height from 0.03 m to 0.16 m. Shoaling effects in the shelf area further increased the wave height to 0.45 m to the entrance of Nagasaki Bay. Here, the wave period was 36 min and by this nearly matched the eigenperiod of the bay of 35 min (internal resonance), which is 6 km long and 1 km wide. Besides wave shoaling and funnelling in the bay, the internal resonance caused the decisive amplification of the initial water disturbance, resulting in a destructive meteotsunami wave of finally 4.8 m at the landward tip of the bay, which claimed three lives.

Frequently affected by meteotsunamis is the harbour of Ciutadella in western Menorca in the Balearic Islands (MONSERRAT *et al.* 1991; RABINOVICH & MONSERRAT 1998; MONSERRAT *et al.* 2006, pp. 1044–1046; RABINOVICH *et al.* 2009, p. 891). The harbour is located at the landward tip of an elongated bay, 1 km long and 100 m wide, opened in a south-western direction. The eigenperiod of the bay is about 10.5 min, which closely matches the wave period of shallow water waves in the wide shelf directly offshore. Both, the coinciding period and the wide shallow waters favour internal and external resonances. Local meteotsunamis mostly appear in the time between June and September, when the macro weather situation frequently causes atmospheric disturbances travelling in a north-eastern direction across the Balearic Islands, i.e. in the direction of the bay of Ciutadella. Destructive meteotsunamis with wave heights of more than 2 m occur every four to five years. Extraordinary wave heights of 6 m were observed in the harbour of Ciutadella in the evening of 15 June 2006. The meteotsunami was related to an atmospheric pressure jump of 7 hPa passing the Balearic Islands with almost 25 m s^{-1} in a north-eastern direction. As a consequence, 40 boats sank and a total economic loss of tens of million euros was reported.

Another hot spot for meteorological induced tsunamis is the Dalmatian coast (BELUŠIĆ & MAHOVIĆ 2009; ŠEPIĆ *et al.* 2009; VILIBIĆ & ŠEPIĆ 2009). This area being liable to meteotsunamis is mainly due to the shallow water depth of the Adrian Sea in connection with the plethora of narrow elongated bays found along the Dalmatian canale-coast, both favouring external and internal resonance effects. In addition, atmospheric disturbances in the form of undulating lee waves are common in the Adrian Sea due to the regular passage of air masses from the nearby Apennine Mountains in the west.

Besides the meteotsunami events described so far, there are further detailed reports on this wave phenomena, e.g. for the Bay of Buenos Aires (DRAGANI *et al.* 2014), the Great Lakes and the US East Coast (BECHLE & WU 2014; HORVATH & VILIBIĆ 2014; ŠEPIĆ & RABINOVICH 2014; VILIBIĆ *et al.* 2014b), the United Kingdom and western Europe (HASLETT *et al.* 2009; FRÈRE *et al.* 2014), the German North Sea (NEWIG & KELLETAT 2011), the Finnish coast (PELLIKKA *et al.* 2014), the Ionian and Aegean Sea (PAPADOPOULOS 1993), western South Africa (OKAL *et al.* 2014) and south-west Australia (PATTIARATCHI & WIJERATNE 2014).

2.3.5 Meteorite impacts

Only terrestrial tsunami trigger mechanisms have been considered so far. However, tsunamis may also be caused by cosmic impacts. Such events being rare occurrence, the knowledge on cosmically induced tsunamis is mainly based on numerical and experimental simulations, asso-

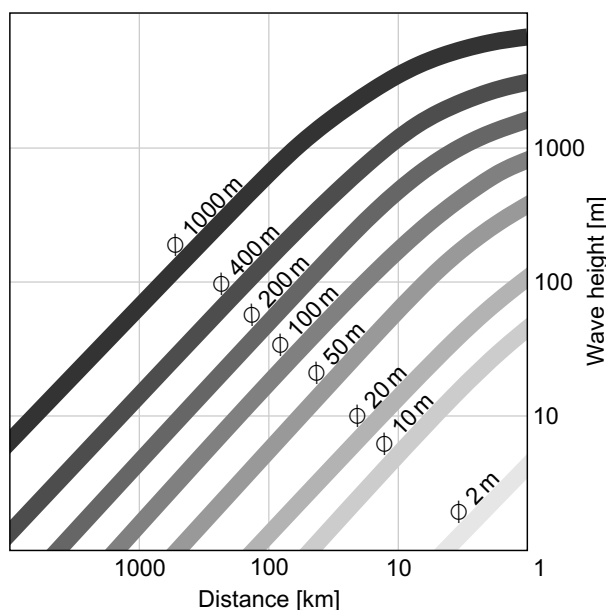


Figure 2.7: Calculated wave heights of cosmically induced tsunamis in the open sea depending on the meteorite diameter (2 m to 1000 m) and the distance from the impact site (after WARD & ASPHAUG 2000, pp. 65–70 and DAWSON & STEWART 2007a, p. 168). Calculations are based on an impact velocity of 20 km s^{-1} , a density of the meteorite of 3 g cm^{-3} and a constant water depth of 4000 m excluding wave shoaling effects.

ciated with corresponding uncertainty (DAWSON & STEWART 2007a, p. 170). Meteorite impacts cause tsunami waves in a similar manner as subaerial mass wastings, especially rockslides, as was already described from Lituya Bay (BRYANT 2008, pp. 236–240; Chap. 2.3.2). Analogously, meteorites displace a huge water volume during the impact, on the one hand because of the entering mass itself, on the other due to the air trapped between the meteorite and the water. As a result of this and especially due to the collapse of the displaced water back into the newly emerged cavity, a multitude of tsunami waves is excited spreading radially from the impact site. Initial wave heights may be gigantic and depend on the impact velocity (typically between 10 km s^{-1} and 45 km s^{-1}), the density of the meteorite, the water depth, the impact angle and, most importantly, the meteorite diameter (WARD & ASPHAUG 2000, pp. 65–70; BRYANT 2008, pp. 238–245). For instance, the impact of a 1 km meteorite in 4000 m deep water would result in initial tsunami wave heights of just as 4000 m (Fig. 2.7). Such impressive wave heights in the order of the water depth are due to the meteorite pushing forward up to the sea floor and temporarily displacing the entire water column upwards (WARD & ASPHAUG 2003, p. F8).

Due to the multitude of wave periods and wavelengths excited during a meteorite impact, tsunami waves quickly disperse from the source area and rapidly flatten (WARD & ASPHAUG 2003, p. F8). Nevertheless, extraordinary wave heights may appear even far distant from the impact site depending on the meteorite diameter (Fig. 2.7). Taking up again the above mentioned example of a 1 km meteorite, tsunami wave heights of several 100 m would be observed 100 km distant to the source and of still several decametres in a distance of 1000 km. For comparison, a 20 m meteorite, causing initial wave heights of 100 m, would result in 1 m tsunami waves only after spreading a 100 km through the ocean.

Cosmic impacts are rare occurrence with large meteorites appearing less frequently than smaller ones. For example, a meteorite 1 km or larger in diameter strikes the earth every 150 000 years on average, whereas objects $\geq 100 \text{ m}$ occur at least every 1000 years (HILLS & GODA 1999,

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p.190). Due to the land-sea ratio on the earth surface, meteorites statistically strike the ocean and may result in tsunami generation in two-third of all cosmic impacts. According to HILLS & MADER (1997, p. 393), the probability of a meteorite causing a teletsunami in one of the world's big oceans and producing wave run-ups of averagely 100 m on adjacent coasts is a few times 10^{-4} per year, that means every few 1000 years.

Albeit there are no written records of fateful cosmic impacts and resultant tsunamis, numerous large impact craters give evidence of such events. Of all 188 or so craters identified on the earth surface, seven are located in the oceans, with another 20 craters found in areas that were covered by the sea at the time of impact (DAWSON & STEWART 2007a, p. 170; PASSC 2015). One of the strongest cosmic impact ever occurred on earth and that was probably accompanied by giant tsunami waves is the impact of Chicxulub in the northern Yucatán Peninsula in Mexico about 65 million years ago (ALVAREZ *et al.* 1980; ALVAREZ 1997; BOURGEOIS *et al.* 1988; SMIT *et al.* 1996; WARD & ASPHAUG 2002, p. 1078). The impact is hold responsible for the extinction of life, especially of the dinosaurs, globally observed for the Cretaceous/Tertiary boundary. With an estimated diameter of 10 km to 15 km, the meteorite has left an impact crater of almost 180 km in diameter at the northern coast of Yucatán, nowadays covered by Tertiary sediment 300 m to 1000 m thick. Initial tsunami wave heights were relatively moderate compared to the impressive size of the meteorite, probably in the order of several 100 m. This was due to the contemporary shallow shelf waters of the impact site. If the water depth had been 1000 m, initial wave heights would have been of the same amount (see above) and still 100 m some 4000 km distant to the source. Nevertheless, the tsunami of Chicxulub probably had far-reaching effects, which is indicated by associated marine high-energy deposits found at the Cretaceous-Tertiary boundary in and around the Gulf of Mexico.

2.4 Sedimentary and geomorphological traces of tsunamis

In the littoral zone, strong laminar and turbulent currents connected with large flow depths may occur with the arrival of tsunami waves (Chap. 2.2), which result in significant erosion and transport of sediment and bedrock. Consequently, tsunamis are assumed to leave characteristic deposits and geomorphological features both on- and offshore, provided that unconsolidated sediment is available and the coast is susceptible to erosion. Geoscientific field investigations into tsunamis are a quite young area of tsunami science. Only few related publications existed prior to 1990, whereas a significant increase, both in number and variety, can be observed since then (SCHEFFERS & KELLETAT 2003, p. 84; KEATING 2006, pp. 392, 393). The knowledge on the sedimentary and geomorphological tsunami imprint is mainly based on field studies along coasts that recently have been affected by tsunamis or known to have experienced significant tsunami inundation in the past (DAWSON & STEWART 2007b, p. 578). Most of these studies concentrate onshore, with the main focus on *tsunamites*, i.e. boulders dislocated by tsunamis or sedimentary event layers of tsunamigenic origin in the geological record. Besides, the landscaping effects of tsunamis have been investigated, meaning in particular large-scale sedimentary and erosive tsunamigenic landforms, e.g. chevrons or dune bedforms as well as sculptured rocky coasts showing whirlpool and cavitation structures (e.g. ANDRADE 1992; YOUNG & BRYANT 1992; BRYANT & YOUNG 1996; BRYANT *et al.* 1996; AALTO *et al.* 1999; KELLETAT & SCHELLMANN 2002; KELLETAT & SCHEFFERS 2003; SCHEFFERS & KELLETAT 2003, pp. 88, 89; BRYANT 2008, pp. 60–123; KENCH *et al.* 2008; GOFF *et al.* 2009; VÖTT *et al.* 2009a, 2011a,b; MAY *et al.* 2012a,b). Due to difficulty of access, studies into the submarine geomorphological effects of tsunamis, including the formation of turbidities, are comparatively rare (e.g. BOURGEOIS *et al.* 1988; POAG *et al.* 1992; SMIT *et al.* 1996; DAWSON & STEWART 2007a, pp. 176–180; SCHEFFERS

et al. 2008b; DAWSON & STEWART 2008; SMIT *et al.* 2012; JONATHAN *et al.* 2012; SAKUNA *et al.* 2012; SMEDILE *et al.* 2012; IKEHARA *et al.* 2014; GOODMAN-TCHERNOV & AUSTIN Jr. 2015; Chap. 2.3.2). The following remarks concentrate on the sedimentary onshore traces of tsunamis, these being most important for the comparison of tsunami simulation results and tsunami field traces presented in Chaps. 3–5.

Tsunami onshore deposits basically derive from the upper subtidal, the beach and a narrow stripe further landwards, which altogether is the erosion zone of tsunamis (e.g. MORTON *et al.* 2007, pp. 187–194; APOTSOS *et al.* 2011b, p. 8; JAGODZIŃSKI *et al.* 2012; SZCZUCIŃSKI *et al.* 2012; SUGAWARA *et al.* 2014a, p. 299). Sediment erosion in the upper subtidal and the intertidal zone is especially significant in the case of *N*-waves making landfall (Chap. 2.5), when coastal waters first recede and expose unconsolidated sediment to be then picked up by the subsequent wave crest (DAWSON & STEWART 2007b, p. 578; MORTON *et al.* 2007, pp. 186, 187). Tsunami waves entrain sediment of all grain sizes ranging from clay to boulders (DAWSON & STEWART 2007a, p. 173), most of which is transported in suspension, whereas bedload is of minor importance, except for larger clasts (MORTON *et al.* 2007, p. 187). With tsunami waters propagating onshore, flow velocity and inundation depth gradually drop due to increasing sediment load, rising topography and the raised surface roughness on vegetated and developed land (DAWSON & STEWART 2007a, p. 173; MORTON *et al.* 2007, pp. 189). This reduces the erosive and transport power of tsunami waves with time, leading to a zone where erosion and deposition balance out. Finally, tsunami wave energy falls below a critical value and sediment is increasingly being released in the deposition zone.

After the run-up and maximum onshore extent of tsunami waters, the flow direction gradually reverses towards the sea forming the backflow. Where the coast steeply slopes into the sea, the backflow velocity can be considerably high, sometimes even higher than the run-up velocity (Chap. 2.2). Frequently, the backflow is channelled following topographic depressions, where erosion is particularly strong (LE ROUX & VARGAS 2005; DAWSON & STEWART 2007a, p. 175). When tsunami waves penetrate deeply into the coast or if wave periods are short, the backflow of the preceding wave may still be active when the next wave arrives. In this case, the run-up is significantly decelerated by the backflow, resulting in a premature deposition of transported sediment (MORTON *et al.* 2007, pp. 189, 202).

The hydrodynamic characteristics of the tsunami run-up and backflow cause a number of sedimentary features typical of onshore tsunami deposits. Such criteria have repeatedly been observed in tsunami deposits on coasts worldwide and are summarised in the following according to the compilations of DAWSON (1994), DAWSON & SHI (2000), DAWSON & STEWART (2007a, pp. 173–180), DAWSON & STEWART (2007b, pp. 578–581), DOMINEY-HOWES *et al.* (2006), MORTON *et al.* (2007), FUJIWARA (2008), SHIKI *et al.* (2008), VÖTT & MAY (2009, pp. 45, 46) and SHANMUGAM (2012).

Owing to the large onshore reach of tsunamis, their sediments are often widely spread across the coast and are found in considerable distance and elevation to the sea. Tsunamites frequently form a single stratum or only few strata draping the former ground separated by an erosional contact or unconformity (Fig. 2.8). In general, the maximum thickness of tsunami sediments is observed in the lower deposition zone. From here, event layers usually thin out seawards as landwards depending on the local conditions, including the local sediment supply, topography, coastline configuration and the individual hydrodynamics. It has to be considered that the landward boundary of tsunami deposits does not give the maximum reach of tsunami waves, which may be far more inland.

Depending on the availability of sediment, tsunami deposits consist of fine- and coarse-grained material with an increasing degree of sorting to the top and a general *fining-upward* tendency

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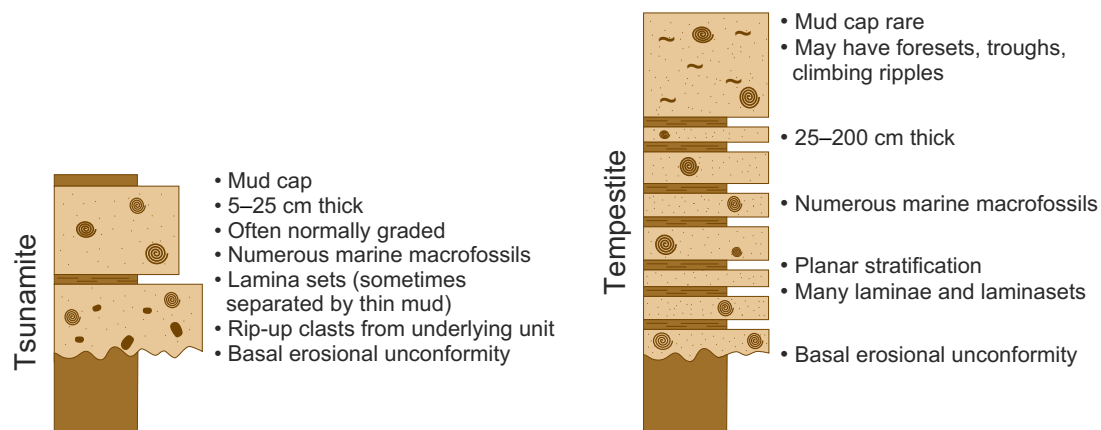


Figure 2.8: Sedimentary features of tsunami deposits (tsunamites) versus storm deposits (tempestites) according to MORTON *et al.* (2007, p. 204) and SHIKI *et al.* (2008, p. 324). The grain size increases with horizontal extent of the illustrated layers.

(normal grading), laminar flow provided. The sediment becomes finer in a landward direction (*fining-landward* or *fining-inland*) due to decreasing transport power with progressive wave propagation. In their lower parts, tsunami deposits frequently show so-called *rip-up clasts*, i.e. fine-grained material eroded from the former land surface and incorporated into event layers during inundation (Fig. 2.9). Tsunami deposits are often capped by a thin layer of clay, silt and terrestrial plant remnants (*mud cap*), which are released right between run-up and backflow when the flow velocity is almost nil. Internal mud layers in tsunamites (Fig. 2.9) have been used to derive the number of individual waves during the tsunami landfall. However, this interpretation is problematic, in that mud layers are prone to erosion by following waves in the tsunami wave train. Moreover, the time between successive tsunami waves (usually < 2 h) may be too short for clay and silt to settle out of the water (cf. BOURGEOIS *et al.* 1988, pp. 241, 242; DAWSON & STEWART 2007a, p. 176). Consequently, not every wave leaves a mud layer and not every mud layer refers to the quiescent phase between two successive waves, but may be connected to other variation in the hydrodynamics of inundation. Another physical feature of tsunami sediments finally relates to the concentrated erosion of the backflow, which may result in channel-like structures in the geological record filled with coarse-grained backflow deposits.

Besides those sedimentary criteria, tsunami deposits are further characterised by the following features:

- Tsunamites often show remnants of diverse micro- and macrofauna entrained by tsunami waves in marine, shallow marine, lagoonal, terrestrial or limnic environments during off- and onshore propagation (DAWSON *et al.* 1996; HEMPHILL-HALEY 1996; HASLETT *et al.* 2000; DAWSON & SHI 2000, pp. 891, 892; DAWSON & STEWART 2007a, p. 174; BRYANT 2008, pp. 57–59; VÖTT *et al.* 2009a, pp. 118–120; VÖTT *et al.* 2015, pp. 96–100; PILARCZYK *et al.* 2014; RAZJIGAEVA *et al.* 2014). Macrofossils especially occur in the form of shell debris but also as fish remnants or corals. Typical microfossils are various ostracods, diatoms and foraminifera. Many fossils are found broken due to turbulent off- and onshore hydrodynamics. Being more resistant, spherical-shaped species are dominant in tsunami deposits.
- Sediment transport by sea water leaves a characteristic geochemical fingerprint in tsunami deposits, which can be detected using various geochemical parameters, as electrical conductivity, pH level, organic content/loss on ignition and especially the (relative) concentration

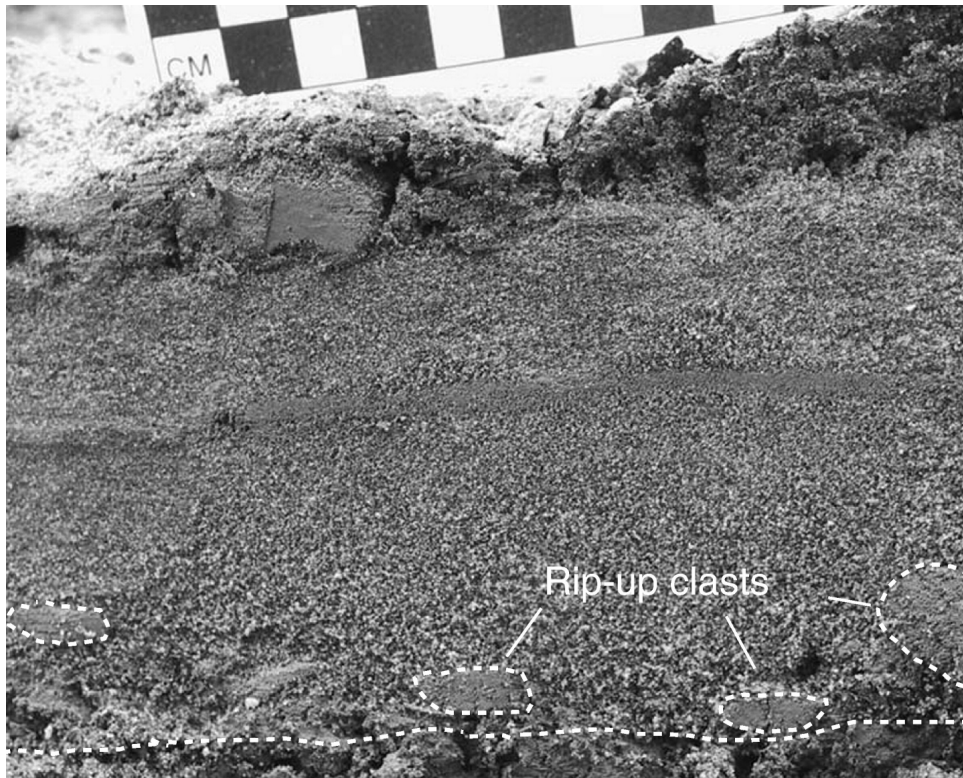


Figure 2.9: Photograph taken in La Quinta (Peru) showing a tsunamite, including a lower erosional contact, rip-up clasts from underlying units, fining-upward sequences and an internal mud layer (MORTON *et al.* 2007, p. 193).

of various chemical elements and compounds as well as their ratio (e.g. SCHROLL 1976, p. 168; MINOURA & NAKAYA 1991; GOFF & CHAGUÉ-GOFF 1999; KORTEKAAS 2002, pp. 23, 24; VÖTT *et al.* 2002, pp. 133–144; VÖTT *et al.* 2003, pp. 126–128; VÖTT *et al.* 2009a, p. 121; VÖTT *et al.* 2011b, pp. 271–273; VÖTT *et al.* 2011c, pp. 260–262; VÖTT *et al.* 2015, pp. 93–96; CHAGUÉ-GOFF *et al.* 2004, pp. 93–96; CHAGUÉ-GOFF *et al.* 2012; SZCZUCIŃSKI *et al.* 2006, pp. 801–803; BLUME *et al.* 2010, p. 27; RÖBKE *et al.* 2013, pp. 90–94). Compared to lower and upper sediment units, tsunami deposits frequently show a higher (relative) concentration of sodium, sulphur, chlorine and boron, which are all major elements of sea water. Due to the production of calcium carbonate of marine organisms, sea water is also enriched with calcium and is therefore higher concentrated in tsunami event layers. With titanium and iron acting as possible terrestrial indicators, this may result in increased calcium-titanium or calcium-iron ratios in tsunami sediments. Whereas the sedimentary iron signal may disappear in time, especially titanium is a reliable indicator of terrestrial conditions as it cannot precipitate secondarily. A wide range of geochemical parameters should be used to detect tsunami deposits because individual parameters can be strongly influenced by the local environment. For example, high calcium concentrations may also originate from calcifying organisms in limnic or fluvial environments or be connected with erosion of carbonate rock onshore.

- Lastly, tsunami layers are distinguished from adjacent stratigraphic units by distinct changes in the pollen spectra and a relative low pollen content (e.g. GOFF *et al.* 2000, pp. 237–246; HUGHES & MATHEWES 2003; TUTTLE *et al.* 2004, pp. 124, 125; VÖTT *et al.*

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2009a, pp. 118–120; CHAGUÉ-GOFF *et al.* 2012; MILLER *et al.* 2013; RAZJIGAEVA *et al.* 2014). The latter is due to high deposition rates during tsunami inundation and sedimentation causing a relative dilution of pollen in the geological record.

Ultimately, the actual characteristics of tsunamites in the field depend on the sediment supply, the local hydrodynamics and the preservation of deposits and therefore are often found modified from the idealised description above. For example, successive waves in a tsunami wave train or the backflow(s) may rework or even completely erode deposits left by the preceding wave(s) (see above; DAWSON & SHI 2000, pp. 882, 883; DAWSON & STEWART 2007b, p. 579). The same may occur during subsequent tsunami or storm events. Finally, event deposits are subject to weathering processes, which can result in significant sedimentary modification (e.g. WHEATCROFT 1990; WHEATCROFT & DRAKE 2003; SZCZUCIŃSKI 2012; SPISKE *et al.* 2013). However, such processes may also preserve tsunami sediments in the form of beachrock, as described, e.g. from western Greece (e.g. VÖTT & MAY 2009; VÖTT *et al.* 2010a,b, 2011b; HADLER *et al.* 2013, 2015a; WILLERSHÄUSER *et al.* 2015b): Originally interpreted as cemented autochthonous beach sediments, detailed analyses of beachrock in several coastal settings revealed a tsunamigenic origin of the sediment. This was derived from several typical tsunamigenic features found in the beachrock in cemented form, including basal erosive unconformities, rip-up clasts, scouring structures, fining-upward sequences and an increasing degree of sorting to the top. The post-depositional cementation of such beachrock-type calcarenitic tsunamites is related to the solution and redeposition of marine biogenic calcium carbonate by percolating rainwater from upper to lower units in tsunami sediments and covering layers. The redeposition of calcium carbonate particularly occurs in ground water or impermeable clay layers. In several locations of western Greece, surf has eroded the calcarenitic covering layers and by this revealed beachrock-type tsunamites, especially in areas of rising sea level.

Many sedimentary and other characteristics of tsunami deposits directly ascribe to the high-energetic nature of tsunami waves and the sea water influence during sediment transport (KORTEKAAS & DAWSON 2007, pp. 208, 209). As such conditions are also connected with storm surges, the question arises, as how far tsunami and storm deposits (so-called *tempestites*) are similar or can be distinguished. Actually, many parallels between both types of event deposits have been observed, but at the same time also several distinct structural differences, strongly depending on the individual tsunami and storm surge intensity as well as the local geographical conditions. Consequently, no universal criteria can be derived in order to distinguish tsunami from storm deposits worldwide. Instead, the individual physiogeographical setting, including in particular the potential tsunami trigger mechanisms, climate, fetch length, sediment supply as well as the off- and onshore topography, has to be considered. For example, areas where both, extreme tsunamis and extreme storm surges may occur, tsunamites and tempestites may show similar characteristics due to a comparable intensity of both wave phenomena (KORTEKAAS 2002, p. 153; MORTON *et al.* 2007, p. 205). This is particularly true for areas affected by tropical cyclones, which generally produce stronger inundation compared to most extratropical cyclones. In other areas, however, the energy of storm surges lags behind that of tsunamis (particularly outside the sphere of cyclones or in enclosed marine basins with short fetch), which allows a better differentiation between both types of deposits.

Differences between tsunamites and tempestites may arise from the individual hydrodynamic characteristics of tsunamis and storm surges. Although both phenomena are dominated by the dynamic of long gravity waves, storm surges contrary to tsunamis form only one large wave driven by the inverted barometer response and the wind-induced rise in water level at the coast connected with the passage of extra-/tropical cyclones. Storm surges cause a slower and gradual rise in water level until the maximum is reached (frequently several metres above sea

level or mean high tide respectively), which may last for several days and affect coasts more than 500 km wide (MORTON *et al.* 2007, p. 194). Besides the rising water level, storm surges are accompanied by high-energetic wind waves and swell, striking incessantly the coast with wave heights of again several metres depending on the wind conditions and local bathymetry. Related to these hydrodynamic differences, tsunami and storm deposits may show a number of individual sedimentary features which have repeatedly been recorded during field studies with the main focus on low latitudes affected by extreme tropical cyclones.

Although (tropical) storm surges can reach several kilometres inland especially on flat coasts, in many cases, the erosion zone is comparatively small and the zone of deposition is restricted to few 100 m within the coastline, with a lower maximum elevation compared to tsunamis (for this paragraph see e.g. DAWSON & SHI 2000, pp. 889–891, KORTEKAAS 2002; TUTTLE *et al.* 2004; KORTEKAAS & DAWSON 2007; MORTON *et al.* 2007). This is due to the slower and only stepwise inundation caused by storm surges and the lower energy of wind waves and swell, accompanied by lower flow velocities and inundation depths. Similar to tsunamis, storm surges move sediment of all grain sizes ranging from clay to boulders, most of which is transported as bedload but not in suspension, as known from tsunamis. While tsunami layers were often observed to form thin drapes over the previous ground, storm deposits rather tend to backfill topographic depressions, but in both cases separated from the lower unit by erosional contacts (Fig. 2.8). The maximum thickness of tempestites usually exceeds the thickness observed for tsunamites. This relates to the longer inundation period during storm surges with innumerable waves picking up sediment from the ground and leaving it on the coast. Moreover, the smaller erosion and deposition zone usually associated with storm surges is accompanied by shorter transport distances and result in stronger concentration of deposits to a thick but narrow stripe. The latter is further connected with a relatively abrupt landward end of storm deposits. Most tempestites show a clearer stratification than tsunamites and are characterised by numerous subhorizontal planar laminae organised into sets due to the high frequency of wind waves and swell (Fig. 2.8). Storm deposits can be normally or inversely graded. They further exhibit foreset or backset bedding and climbing ripples owing to the dominating bedload transport, lower inundation depths and variable flow velocities. Rip-up clasts are rare in tempestites because most fine-grained material picked up from the ground decays before deposited due to the long inundation period and continuous hydrodynamic turbulence during storm surges. Also mud caps are rarely found capping storm deposits.

Where event deposits are found outside the possible reach of storm surges or in areas protected from storm waters, the identification of tsunamites is more clear. This, for instance, may be the case in lagoons, coastal freshwater lakes, harbours or marine inlets cut off from the open sea. Such places are further advantageous for the detection of tsunamites, in that allochthonous high-energy deposits are often found encased in autochthonous fine-grained sediment and therefore clearly stand out as a temporary disturbance of the local sedimentary environment (cf. VÖTT *et al.* 2006, 2008, 2009a,b; HADLER *et al.* 2015a,b; WILLERSHÄUSER *et al.* 2015a,b). Nevertheless, it has to be considered that tsunami deposits in lagoonal or limnic settings may show modified physical features (e.g. thickness or lateral extent) due to the deposition in shallow water (MORTON *et al.* 2007, p. 205).

As mentioned before, field studies into the sedimentary onshore response of tsunamis also focus on boulder deposits, which are briefly discussed in the following. Dislocated boulders or megaclasts have been observed on coasts worldwide and are seen in connection with extraordinary, high-energetic wave events, as presented by tsunamis (e.g. VÖTT *et al.* 2006, pp. 149–151; GOTO *et al.* 2007; SCHEFFERS 2008; SCHEFFERS *et al.* 2008a; IMAMURA *et al.* 2008; ETIENNE *et al.* 2011; ENGEL & MAY 2012). Such waves are able to remove megaclasts (either loose or

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as part of the bedrock) from the littoral zone and transport them—separated or in groups—landwards to higher ground, sometimes several 100 m inland and many metres above sea level. Frequently, remnants of marine fauna, boreholes related to piddocks, or *rock pools*—circular structures in rock caused by bioerosion due to marine gastropods—indicate the littoral origin of dislocated boulders. Megaclasts appear separated or arranged in walls, clusters or so-called *boulder trains* and often show imbrication when found in groups. Their weight ranges from few tons to several thousands of tons, with a diameter of between few and numerous metres. Despite these dimensions, megaclasts are assumed to be dominantly transported in suspension by tsunami waves, as is true for fine-grained material. The transport of boulders depends on a number of factors, including the mass, shape and volume of the boulder, the solidity and initial water depth of the boulder source as well as the slope and bottom roughness in the littoral zone. Moreover, the wave dynamics are crucially important, i.e. in particular the wave height, flow velocity, inundation depth, waveform, manner of breaking (spilling, plunging or non-breaking) as well as the angle of wave approach.

As was already seen for sedimentary event layers, the interpretation of dislocated megaclasts has to consider not only tsunamis but also storm surges as a possible transport mechanism (e.g. NOTT 1997, 2003a,b; NOORMETS *et al.* 2004; BENNER *et al.* 2010; GOTO *et al.* 2010; SCHEFFERS *et al.* 2010; LORANG 2011; NANDASENA *et al.* 2011; ERDMANN *et al.* 2015). Although, wind-generated waves are far less energetic than tsunami waves, such displacements were repeatedly observed in the field. Several formulae have been created in order to determine the critical wave amplitude of storm and tsunami waves to dislocate boulders depending on their size and shape, the bottom roughness, the inclination and many other parameters (e.g. NOTT 1997, 2003a,b; NOORMETS *et al.* 2004; BENNER *et al.* 2010; LORANG 2011; NANDASENA *et al.* 2011). The formulae showed that storm waves demand far higher amplitudes than tsunamis to cause boulder displacement. However, the determination of critical wave amplitudes is problematic in that a specific wave amplitude is not restricted to a specific flow velocity and flow depth necessary for dislocation, but also strongly depends on the wave period, waveform, the manner of breaking and the flow direction.

Transport formulae demand reliable information on the shape, volume and the mass of dislocated megaclasts. Whereas simple axis measurements may lead to a considerable overestimation of boulder volume and mass and therefore of the critical wave amplitude for dislocation, reliable results can be gained using terrestrial laser scanning (HOFFMEISTER *et al.* 2013, 2014). This technique further allows the regular monitoring of boulders and to record dislocation due to meanwhile storm or tsunami activity in order to determine the relevant transport mechanism. However, in many cases, storm surges can only be excluded as the cause of dislocation, if (i) the boulder mass clearly exceeds the transport capability of wind-generated waves, (ii) boulders are found outside the reach of storm surges or (iii) wave heights that are required for transport are not given due to the local configuration of the coast or the regional climate. In such cases, tsunamis are the favoured transport mechanism of dislocated megaclasts.

2.5 Numerical modelling and simulation of tsunamis

Tsunami risk assessment and mitigation plans based on numerical modelling and simulation of tsunamis increasingly gain in importance. The terms *modelling* and *simulation* are distinguished as follows (MARIA 1997, pp. 7, 8): Modelling is the process of setting up a (numerical) model as a simplified representation of a system of interacting components in nature or society. A key demand on models is to incorporate the salient features of the represented system but at the same time to be still comprehensible and easy to experiment with. Accordingly, the British

economist Joan Robinson once aptly stated: “A model which took account of all the variegation of reality would be of no more use than a map at the scale of one to one” (ROBINSON 1965, p. 33). In contrast to modelling, simulation is the operation and application of a model in view of a specific problem in the represented system in order to study certain characteristics or interactions to be expected in reality. During this process, parameters that can hardly or even not at all be controlled in the real system are continuously adjusted in the model to investigate the resulting effects and to increase the model’s reliability. In a transferred sense, numerical modelling of tsunamis can be understood as “the application of numerical methods to the equations governing long wave propagation over basin-scale bathymetries” (DALRYMPLE *et al.* 2006, p. 150). The resulting model is then used to calculate tsunami generation, propagation and inundation considering and adjusting different parameters, such as the generation mechanism, tsunami waveform and wave constellation (i.e. hydrodynamic boundary conditions), off- and onshore bottom friction etc., which is all part of the simulation process. Numerical simulations allow to go through various possible tsunami scenarios in order to evaluate local tsunami risks and to work out appropriate mitigation plans.

Numerical models calculate natural flow phenomena only in limited resolution, both in space and time. Therefore, the essential first step among all hydrodynamic numerical modelling techniques is *discretisation*, which is the transfer of continuous objects into a finite number of discrete elements, suitable to be dealt with by numerical methods and digital computers. This process mainly involves the reduction of continuous differential equations into discrete difference equations, e.g. with the *leapfrog integration* method (e.g. SHUTO 1991, pp. 179, 180; DURRAN 2010, pp. 62–67). Once these equations are derived, spatial discretisation is realised based on calculation grids covering the study area or represented system with a certain grid spacing (Fig. 2.10). The centre or one of the edges of each grid cell is representative for the whole area covered by this cell. Synonymously, simulations are not calculated continuously over time but with a fixed time step to achieve temporal discretisation.

Appropriate grid spacing and time step are determined according to the *Courant-Friedrichs-Lewy (CFL) condition* (COURANT *et al.* 1928; DURRAN 2010, pp. 98–100):

$$C = \frac{\Delta t \sqrt{gd}}{\Delta x} \leq 1 \quad (2.8)$$

where C = Courant-Friedrichs-Lewy number (dimensionless)
 Δt = calculation time step [s]
 Δx = grid spacing [m].

The condition gives the *Courant-Friedrichs-Lewy* or *Courant number*, which is the number of grid cells passed by the observed quantity (e.g. wave crest of a tsunami) per time step during simulation. The CFL condition says that the Euler method used for solving ordinary differential equations in hydrodynamic numerical models is only stable for CFL numbers ≤ 1 . As is clear from Eq. 2.8, small grid spacing, i.e. high spatial resolution, requires short time steps, to say high temporal resolution. Furthermore, higher flow velocities demand shorter time steps or larger grid spacing to keep stability. With tsunami waves propagating faster in deep water (Eq. 2.2), here, grid spacing has to be larger or the time step to be shorter compared to shallow water areas. Albeit the CFL condition demands a CFL number ≤ 1 , in practice, many numerical models even run stable for numbers ≤ 10 (e.g. Deltares 2014, pp. 39, 40).

Independent of the CFL condition, two rules of thumb are commonly used to determine appropriate grid spacing and time step for tsunami that means shallow water wave simulations (SHUTO *et al.* 1986; BRYANT 2008, p. 37; Deltares 2014, p. 40):

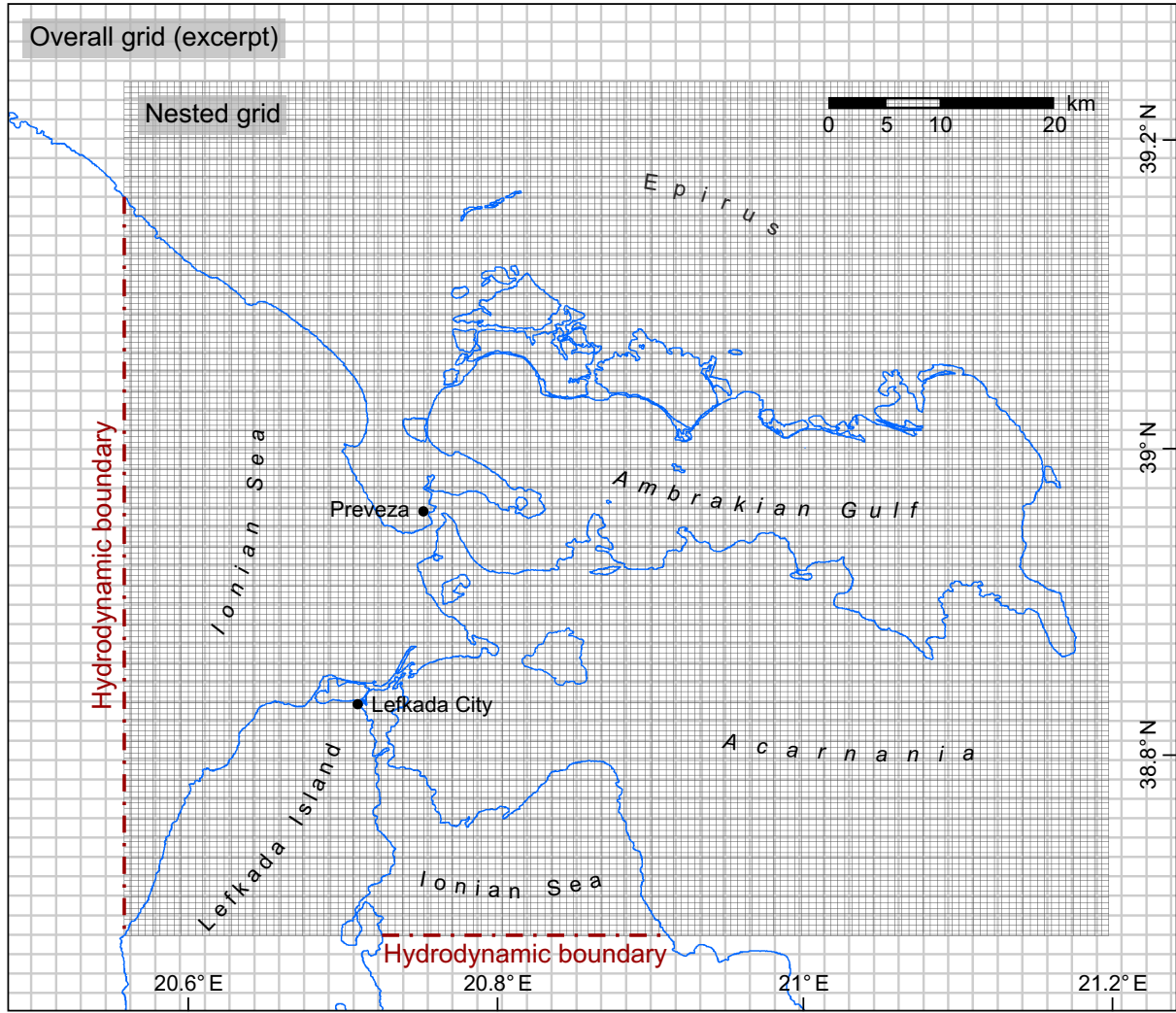


Figure 2.10: Example of a rectangular nested two-part computational grid from the eastern Ionian Sea and the Ambrakian Gulf (western Greece; Chap. 5). Grid spacing is 900 m for the overall grid and 90 m for the nested grid. Computational results of the overall grid are transferred to the nested grid along the hydrodynamic boundaries, e.g. by means of bilinear interpolation. For a better visualisation of the nested grid, grid lines were thinned out.

$$\Delta x \leq \frac{\lambda}{30} \quad (2.9)$$

and

$$\Delta t \leq \frac{T}{40}. \quad (2.10)$$

Consequently, the tsunami wavelength should be covered by at least 30 grid cells and the wave period by at least 40 time steps. However, often more (i.e. smaller) grid cells and more (i.e. shorter) time steps will be chosen to achieve higher spatial and temporal resolution demanded in many tsunami (risk) studies. With high spatial and/or temporal resolution commonly being required locally only but not across the entire model, often several nested calculation grids

with decreasing spacing are used (Fig. 2.10; e.g. GEIST 1997, p. 130; CHOI *et al.* 2003b; SON *et al.* 2011; YAMAZAKI *et al.* 2011; ZHOU *et al.* 2011; JAMELOT & REYMOND 2015). According to this method of *nesting*, simulations are first conducted for the overall grid covering the entire study area. Subsequently, computational results of the overall grid are transferred to the smaller, subordinate nested grid, e.g. by means of bilinear interpolation (Deltares 2014, pp. 677–683). This process can be repeated many times and allows a considerable reduction of the computational cost and thereby of the total calculation duration.

Numerical tsunami models provide only an approximation of the actual flow dynamics in nature with a certain degree of error inherent. Already the first essential step in tsunami modelling, the discretisation, is connected with a clear but acceptable loss of information, known as the *discretisation* or *truncation error*. Once the initial water surface disturbance in the tsunami source area or the initial tsunami waveform is determined, subsequent wave propagation and inundation have to be dealt with by the numerical model using appropriate wave theories — a process which again is subject to approximation. Hardly being influenced by frequency and amplitude dispersion effects, tsunami propagation in the open ocean is adequately reflected by shallow water equations in their linear form, which are derived from the three-dimensional Navier-Stokes equations (SHUTO 1991, pp. 177–181; GEIST 1997, pp. 126–131, DALRYMPLE *et al.* 2006, pp. 145–147). With progressive wave propagation and decreasing water depth, however, non-linear and dispersive effects gain in importance and have to be considered in the numerical model. This is demonstrated by the *Ursell number*, a parameter indicating the degree of non-linearity of longwave gravity waves, as represented by tsunamis (URSELL 1953; OSBORNE & PETTI 1994, p. 1729; cf. LEHFELDT *et al.* 2007, pp. 107, 108):

$$U = \frac{3}{16\pi^2} \left(\frac{A}{d}\right) \left(\frac{\lambda}{d}\right)^2 \quad (2.11)$$

where U = Ursell number (dimensionless)
 A = wave amplitude [m].

Consequently, the Ursell number is the relative wave amplitude times the relative wavelength squared, multiplied by a constant. In large parts of the deep oceans, the Ursell number is small ($U \ll 1$), suggesting linear wave motion in good approximation. With decreasing water depth, the Ursell number becomes larger ($U \gg 1$) and non-linear as well as dispersive effects have to be considered.

Non-linear, dispersive wave behaviour of tsunamis is adequately described by the *Boussinesq approximation* (SHUTO 1991, pp. 177–181; DALRYMPLE *et al.* 2006, pp. 146, 147; TKALICH & DAO 2011, pp. 275–286). This approximation is based on the Boussinesq equations, originally written in the following form (BOUSSINESQ 1872):

$$\frac{\partial^2 \eta}{\partial t^2} - g d \frac{\partial^2 \eta}{\partial x^2} - g d \frac{\partial^2}{\partial x^2} \left(\frac{3}{2} \frac{\eta^2}{d} + \frac{1}{3} d^2 \frac{\partial^2 \eta}{\partial x^2} \right) = 0 \quad (2.12)$$

where η = elevation above water surface [m]
 t = time [s]
 x = horizontal component [m].

The approximation was formulated by Joseph Boussinesq according to the observation of the *wave in translation* by John Scott Russell in 1834 (RUSSELL 1885). In the Union Canal in

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Edinburgh (UK), Russell observed a solitary water elevation originating from the bow of a boat drawn by horses from the banks. When the boat stopped, the water elevation was released from the bow and kept on moving in the canal without a change in shape or velocity, which Russell called the wave in translation. This was the first description of solitary waves, which have often been used to describe tsunami wave behaviour (see below). Russell's observation could not be explained by the contemporary water wave theories and was not fully accepted before the formulation of the approximation by Boussinesq in 1872.

The Boussinesq approximation is a simplification of the three-dimensional Navier-Stokes equations and incorporates dispersive effects into the non-linear shallow equations (HAMM *et al.* 1993, pp.19–25; KIRBY 2003; DALRYMPLE *et al.* 2006, pp.146, 147; DURRAN 2010, pp.20–26; TKALICH & DAO 2011, pp.275–286; Deltares 2014, pp.179–186). The basic idea of this approximation is balancing weak frequency dispersion against weak wave non-linearity, which approximately meets the natural behaviour of tsunami waves propagating towards the coast. The Boussinesq approximation eliminates the vertical coordinate from the flow equations with the assumption of a non-hydrostatic pressure. This is connected with a considerable reduction of the computational effort compared to the full three-dimensional treatment of the corresponding wave dynamics. Since their first formulation in 1872, the Boussinesq equations have been varied and extended many times, yielding a variety of Boussinesq-type equations (HAMM *et al.* 1993, pp.20–22; DALRYMPLE *et al.* 2006, pp.146, 147; TKALICH & DAO 2011, pp.284–286). Such extensions or modifications involve, for instance the consideration of varying water depth (PEREGRINE 1967) or the improvement of the non-linear wave behaviour, the latter resulting in the fully non-linear Boussinesq equations (e.g. WEI *et al.* 1995). A common modification of the Boussinesq equations is the Korteweg-de-Vries (KdV) equation, which is advantageous to describe wave propagation in only one horizontal dimension (KORTEWEG & DE VRIES 1895).

Besides the linear shallow water equations and the Boussinesq approximation, further wave theories exist that may adequately reflect tsunami wave behaviour on certain conditions. For example, also the non-linear shallow water (NLSW) equations, which are the Boussinesq equations with omitted dispersion terms, have widely been applied in numerical tsunami models (HAMM *et al.* 1993, p.23; MURTY & KOWALIK 1993; DALRYMPLE *et al.* 2006, p.145; LEHFELDT *et al.* 2007, pp.107, 108; TKALICH & DAO 2011, pp.285–287; LIANG *et al.* 2013a). Nevertheless, instead of switching from one wave theory to another, each of which describing the local tsunami behaviour most appropriate, the single use of the fully non-linear Boussinesq equations as a comprehensive model is advantageous because it adapts automatically to the particular conditions of each range. This is suggested by DALRYMPLE *et al.* (2006, p.147), who stated: “Although the use of fully nonlinear Boussinesq equations, with dispersive and large amplitude effects, can be considered overkill in the context of a general basin-scale tsunami model, the generality of the modelling framework provided by the model is advantageous in that it automatically covers most of the range of effects of interest, from propagation out of the generation region, through propagation at ocean-basin scale, to run-up and inundation at affected shorelines.”

Mathematically, tsunamis can be described by diverse waveforms depending on the progress in wave propagation and the bathymetric conditions. Due to the wide variety of wave types, this treatise gives only a short overview over the essential and most common waveforms. For further reading see references in the following. According to the linear shallow water wave theory, tsunamis in the open ocean are basically represented by sinusoidal waves, small bathymetric fluctuations provided (GEIST 1997, pp.131, 132; WEISS 2004, p.26; BRYANT 2008, pp.30–34). With tsunamis propagating towards the coast, water depth gradually decreases and non-linear as well as dispersive effects become more influential. This is connected with a progressive steepening of tsunamis, which may result in the formation of solitary waves in time. First described by John

Scott Russell as the wave in translation (see above), solitary waves have often been applied in numerical and experimental tsunami simulations (DAILY & STEPHAN Jr. 1952; SYNOLAKIS 1987; WIEGEL 1992, pp. 65–108; YEH *et al.* 1994; BRIGGS *et al.* 1995; GEIST 1997; WEISS 2004, p. 26; LEHFELDT *et al.* 2007; SYNOLAKIS *et al.* 2008; BALDOCK *et al.* 2009; APOTSOS *et al.* 2011a; TKALICH & DAO 2011, pp. 275–279; RÖBKE *et al.* 2012, 2013; LIANG *et al.* 2013b; Chaps. 3, 4). Based on the Boussinesq approximation and KdV equation, a common solution for solitary waves is given by DEAN & DALRYMPLE (1991, pp. 314–317):

$$\eta = H \operatorname{sech}^2 \left(\sqrt{\frac{3H}{4d^3}} ct \right) \quad (2.13)$$

As can be seen, solitary waves are only defined for positive values of η , meaning, there is no negative wave component below water surface. Thus, wave amplitude and wave height are the same for this waveform.

The suitability of using solitary waves to represent tsunamis has recently been questioned, since their time- and length-scales have been shown to differ from those of real tsunami waves (MADSEN *et al.* 2008; MADSEN & SCHÄFFER 2010; CONSTANTIN 2009; CONSTANTIN & JOHNSON 2008; APOTSOS *et al.* 2011a, p. 2114). Nevertheless, solitary waves are still widely used for tsunami simulations as they describe the ideal waveform to minimise energy loss during propagation and reflect adequately the compensation of frequency dispersion by non-linearity, typical of tsunami waves. Moreover, tsunamis often appear as leading wave crests after long propagation distances (teletsunamis), as can be approximately represented by solitary waves (see references in paragraph above).

However, tsunamis have often been observed to cause coastal waters initially to recede before the first wave crest arrives. That means, tsunamis can also comprise a negative wave component—a wave trough. This is especially the case when tsunamis are generated close to the coast and the propagation distance is too short for the waves to evolve into leading solitary waves (TADEPALLI & SYNOLAKIS 1994, p. 100; TADEPALLI & SYNOLAKIS 1996, p. 2143; MADSEN *et al.* 2008, pp. 2, 3; SYNOLAKIS *et al.* 2008, pp. 2202, 2203; LIANG *et al.* 2013b, pp. 144, 145). This situation is frequently found for coasts directly adjacent to subduction zones, which are highly tsunamigenic (Chap. 2.3.1). Based on these observations, TADEPALLI & SYNOLAKIS (1994, 1996) developed the *N*-wave theory, which describes dipolar tsunami waves showing both, a wave trough and a wave crest. The *N*-wave theory comprises leading-depression (LDN) and leading-elevation (LEN) *N*-waves, both isosceles or asymmetrical. According to the formulation by TADEPALLI & SYNOLAKIS (1994, p. 105), isosceles LDN-waves are expressed by (cf. CARRIER *et al.* 2003, p. 86; CHACÓN-BARRANTES *et al.* 2013, pp. 15, 16):

$$\eta = \frac{3\sqrt{3}H}{2} \operatorname{sech}^2 \left[\frac{3\alpha}{2} \sqrt[4]{\frac{3A}{4}} (t - t_0) \right] \tanh \left[\frac{3\alpha}{2} \sqrt[4]{\frac{3A}{4}} (t - t_0) \right] \quad (2.14)$$

where α = constant modifying the wavelength (dimensionless)
 t_0 = time of the wave's mid-point (half wave period) [s].

Based on theoretical studies (TADEPALLI & SYNOLAKIS 1994, 1996; LIANG *et al.* 2013b) and numerical simulations (Chaps. 3, 4), LDN-waves were found to have a stronger onshore response on the coast compared to solitary waves of the same wave amplitude. This phenomenon is known as the *N*-wave effect, which is mainly due to the steeper wave crest of LDN-waves increasing its

inertia (TADEPALLI & SYNOLAKIS 1994, p. 109). In contrast, LIANG *et al.* (2013b) showed LEN-waves having the slightest onshore response compared to LDN- and solitary waves, although their wave crest has the same steepness as solitary waves. This arises from the subsequent wave trough of LEN-waves, which reduces the length of the preceding wave crest and therefore shortens the time period of a high water level during the run-up compared to solitary waves (LIANG *et al.* 2013b, pp. 150, 151). Simulation results revealed, that the *N*-wave effect becomes stronger with increasing wave amplitude as the wave steepness increases disproportionately with the amplitude (Chap. 3.4).

According to SYNOLAKIS *et al.* (2008, p. 2198) and RABINOVICH (2014, p. 452), the mathematical description and numerical modelling of tsunami waves considerably improved since the implementation of DART-buoys (NOAA 2015b) and familiar monitoring stations. Such *tsunameters* or *tsunamographs* build a real-time tsunami monitoring system in the world's oceans and allow recording the original tsunami signal or waveform in the deep-ocean, unaffected by coastal resonance effects and in the absence of non-linear effects resulting from changing bathymetry, bottom friction etc. (RABINOVICH 2014; RABINOVICH & EBLÉ 2015; RABINOVICH *et al.* 2011a). Where available, data on original tsunami waveforms has become an essential initial input in numerical tsunami models and has significantly narrowed the gap between simulations results and actual field observations and measurements (e.g. TITOV *et al.* 2005; RABINOVICH *et al.* 2011b; SAITO *et al.* 2011; THOMSON *et al.* 2011).

2.6 Summary

Tsunamis (from Japanese *tsu* = harbour, *nami* = wave) are single or a series of long water waves in oceans or large lakes that are triggered by a sudden vertical displacement of the water column. Due to their long wave length (100 km to 500 km in the open sea), tsunamis are shallow water waves that can travel across entire oceans without major energy loss. Approaching the coast, tsunamis experience strong wave shoaling, which together with funnel and internal resonance effects may increase the tsunami wave height from few metres in the deep ocean to tens of metres at the shoreline. At the coast, tsunamis usually form spilling breakers but do not break in a plunging way. During the wave run-up, flow velocities may reach up to 20 m s^{-1} and inundation depths of more than 20 m resulting in considerable run-up heights and a large inland penetration. After the run-up, tsunami waters form a backflow mostly following topographic depressions in the direction of the sea.

Almost 82.3 % of all tsunamis recorded worldwide are associated with seismic activity. Most seismogenic tsunamis are caused by the vertical shift of the seafloor during the focal mechanism. Shallow focus earthquakes along dip-slip or oblique-slip faults with a rupture velocity ranging from 0.1 km s^{-1} to 1 km s^{-1} are most efficient in generating tsunamis. Gravitational mass wastings are the second major cause of tsunamis. Especially submarine landslides are known to have triggered severe tsunamis due to their large sediment volume (up to $20\,000 \text{ km}^3$). The strongest tsunamis are expected, if the slide velocity approximates the local tsunami wave celerity. In this case, most of the wave energy is focused in two opposite directions along the axis of the slide, which is in contrast to the elongated beaming usually found for seismogenic tsunamis. Volcanic eruptions can cause devastating tsunamis in the near-field, but usually their wave energy rapidly diminishes due to the limited point source. There are various mechanisms generating volcanogenic tsunamis ranging from volcanic earthquakes to pyroclastic flows, submarine volcanic explosions, caldera formation and atmospheric pressure waves due to volcanic explosions. So-called meteotsunamis are associated with atmospheric disturbances of only few hectopascals traveling across the sea or lakes, as for instance meteorological fronts or gravity waves.

Such atmospheric disturbances are accompanied by a corresponding fall or rise in water level by several centimetres forming a very shallow and long water wave. Due to external resonance effects in shallow shelf waters and internal resonance effects in marine inlets, these initial waves can strongly be amplified resulting in tsunamis more than 5 m high at the shore. Tsunamis may also be cosmically induced. Meteorite impacts generate tsunamis in a similar manner as is known from subaerial mass wastings. However, due to the considerable impact velocity of between 10 km s^{-1} and 45 km s^{-1} , meteorite impacts can cause gigantic tsunamis with initial wave heights in the order of the local water depth in the deep ocean.

Causing strong currents and large flow depths in the shelf and coastal area, tsunamis can have a significant sedimentary on- and offshore response on coasts. Whereas little is known about offshore sedimentary signals of tsunamis (especially turbidities), there is much information on onshore tsunamites, i.e. sedimentary tsunami layers in the geological record and dislocated boulders. Tsunamites have been identified along coasts worldwide showing a number of typical sedimentary features that are directly ascribed to the hydrodynamic characteristics of tsunami run-up and backflow. Such features are especially basal erosional unconformities, fining-upward and -landward, a high content of marine fossils, rip-up clasts and mud caps as well as a general high-energetic allochthonous character of the sediment. Due to the sediment transport by sea water, tsunami deposits can further be detected using palynological analyses as well as several geochemical parameters, particularly the (relative) concentration of various chemical elements. The differentiation between tsunamites and tempestites, i.e. storm-related deposits, can be particularly challenging, especially, if event deposits are located in the sphere of influence of both, tsunamis and storm surges. Although, a number of physical criteria exist to distinguish both types of event deposits, the actual appearance of tsunami and storm deposits ultimately depends on the physiogeographical setting, including potential tsunami trigger mechanisms, climate, off- and onshore topography, fetch length and sediment supply.

In terms of modern tsunami risk assessment, numerical modelling and simulation of tsunamis increasingly gain in importance. The essential first step in tsunami modelling is discretisation, since numerical models can calculate natural flow phenomena only in limited resolution. In order to determine a reasonable spatial and temporal resolution according to the physical features of the study area and flow phenomenon, the Courant-Friedrichs-Lewy condition is used. High spatial resolution in the area of interest is often achieved using the method of nested calculation grids, which significantly reduces the computational cost.

Whereas tsunami propagation in the deep ocean is adequately reflected by shallow water equations in their linear form, non-linear and dispersive effects gain in importance with decreasing water depth in the area of the continental slope and shelf. This is demonstrated by the Ursell number indicating the degree of non-linearity of longwave gravity waves. The basic theory describing the non-linear, dispersive wave behaviour of tsunamis is the Boussinesq approximation, which is a simplification of the three-dimensional Navier-Stokes equations. This approximation balances weak frequency dispersion against weak wave non-linearity. The Boussinesq approximation has been varied several times, yielding, among others, the Korteweg-de-Vries equations and the fully non-linear Boussinesq equations. Although various wave theories exist that may adequately reflect tsunami wave behaviour on certain conditions, the single use of the fully non-linear Boussinesq equations in numerical models is advantageous because it adapts automatically to the particular conditions of each range.

Depending on the progress in wave propagation and the bathymetric conditions, tsunamis can be mathematically described by various waveforms in numerical models. Tsunamis in the open ocean are basically represented by sinusoidal waves according to the linear shallow water wave theory. In contrast, non-linear and dispersive effects become more influential near the

2 The tsunami phenomenon

continental shelf, which may result in the formation of solitary waves. Although the suitability of using solitary waves to represent tsunamis has recently been questioned, this wave form has widely been used for tsunami simulations because it describes the ideal waveform to minimise energy loss during propagation and reflects adequately the compensation of frequency dispersion by non-linearity. Solitary waves only comprise a positive wave component but no wave trough. However, especially tsunamis generated nearshore have often been observed to cause a significant withdrawal of the shoreline before the first wave crest arrives. Based on this observation, the *N*-wave theory has been developed, which comprises leading-depression and leading-elevation *N*-waves.

3 Effects of different boundary conditions and palaeo-/topographies on the onshore response of tsunamis in a numerical model — a case study from western Greece¹

Abstract Hydrodynamic numerical models are essential in modern tsunami risk assessment. They allow the economical simulation of possible tsunami scenarios for areas at risk and provide reliable and detailed insights into local onshore dynamics. This is especially true, when simulations are calibrated with field traces of past tsunami inundation events. Following this approach, the current study focuses on palaeotsunami events indicated by sedimentary and geomorphological field traces in the northern Gulf of Kyparissia (NW Greece). Based on three different digital elevation models (DEM)—reflecting the recent and two palaeotopographies—various tsunami wave constellations according to the solitary and *N*-wave theory are numerically simulated. The main objective is to investigate the effects of both, different palaeo-/topographies and boundary conditions on the tsunami onshore response in the numerical model.

Tsunami landfall related to *N*-waves is found to be considerably stronger compared to solitary waves. This phenomenon, known as the *N*-wave effect, is demonstrated for the first time in a specific study area. Inundation dynamics are even stronger affected by the different palaeo-/topographies, which is due to substantial vertical crust movements in the northern Gulf of Kyparissia considered in the palaeo-DEMs. By applying different waveforms and palaeo-/topographies, the model achieves close agreement with field observations, altogether revealing a high tsunami risk for the Gulf of Kyparissia, which is in contrast to conventional numerical studies of the area. The marked differences between the presented scenarios emphasise the need to consider a wide variety of possible hydrodynamic boundary conditions and probable topographical conditions in order to find scenarios in plausible accordance with palaeotsunami field traces. Once a plausible scenario is found, it can be applied to the recent topography in view of a reliable modern risk assessment.

3.1 Introduction

Coasts have become increasingly popular, both as places of residence and centres of tourism. Accordingly, coastal risk assessment serves a key role in today's society. Besides storm surges, tsunamis are the greatest threat to those living by the coast, having resulted in huge human and economical losses worldwide, recently in Papua New Guinea (1998), south-east Asia (2004) and Japan (2011). In terms of risk assessment and mitigation plans, hydrodynamic numerical models play an important part. They are a powerful tool to economically simulate possible inundation scenarios for coasts known to be historically vulnerable to tsunamis. Particularly, the synthesis of numerical simulations and tsunami field traces has proven fruitful as it supports a detailed reconstruction of past tsunami events — vital for reliable future risk assessment (RÖBKE *et al.* 2012, 2013). Field observations allow numerical models to be calibrated and verified, so

¹This chapter is based on the paper by RÖBKE *et al.* (2016a) submitted to *Continental Shelf Research*.

3 Effects of different boundary conditions and palaeo-/topographies

that calculation results are improved. Furthermore, valuable information about the trigger mechanism of tsunamis can be derived from numerical models when calibrated with field data.

Several studies exist pursuing this promising approach in view of recent tsunami events. A common strategy is to work out the best fit between simulations and field traces by adjusting appropriate parameters in the numerical model. Frequently, different fault-plane solutions of tsunamigenic earthquakes are considered in order to determine the focal mechanism and tsunami which compare favourably with field observations. For example, BORRERO *et al.* (2009) computed four different slip distributions and associated tsunami waves of the 12 September 2000 Bengkulu earthquake (Sumatra, Indonesia) to reproduce observations from DART-buoys, eyewitnesses and the field. Also BOURGEOIS *et al.* (1999) and LE FRIANT *et al.* (2008) made use of different fault-plane solutions to adjust their numerical model to field data they obtained from the 21 February 1996 Chimbote tsunami in northern Peru and the 21 November 2004 Les Saintes tsunami (Lesser Antilles) respectively. SATAKE *et al.* (2013) took this approach one step further by considering detailed tsunami tide gauge, GPS- and DART-buoy data as well as field traces in order to accurately estimate the slip distribution of the associated 25 October 2010 Mentawai earthquake (Indonesia) using a numerical model. In the case of tsunamis generated by submarine landslides, varying essential slide properties or the initial free water surface conditions characteristic of different slide scenarios allow numerical models to be adjusted with regard to survey data. This approach was taken by LYNETT *et al.* (2003), who compared field traces of the 17 July 1998 Papua New Guinea tsunami with simulations that considered three realistic initial water surface disturbances. Finally, independent from the trigger mechanism, numerical tsunami models are calibrated with field observations by using different friction maps, as performed, e.g. by GOPINATH *et al.* (2014), who focused on the 2004 Indian Ocean tsunami along the Tamil Nadu coastline.

The calibration of numerical models by means of field data is a particularly demanding task with regard to pre-/historical tsunami events. This is mainly due to (i) less accurate or even missing data of the tsunami waves and trigger mechanisms, as for instance seismological and gauge/buoy data or historical accounts, (ii) lower spatial resolution and accuracy of field data owing to a general low preservation level of sedimentary tsunami traces (e.g. WHEATCROFT 1990; WHEATCROFT & DRAKE 2003; SZCZUCIŃSKI 2012; SPISKE *et al.* 2013), (iii) the fact that tsunami deposits do not allow a precise deduction of the maximum inland penetration of tsunami waters but only of the maximum sediment transport and (iv) uncertainties about the contemporary landscape conditions concerning coastline configuration (highly variable, even in short timescales), topography or sea level. Consequently, there are only a few studies dealing with this issue, and often a simplistic approach is used.

For example, MARTIN *et al.* (2008) compared numerical simulation results of the 1969 Ozernoi and 1971 Kamchatskii tsunamis with sedimentary deposits found along the Pacific coast of Kamchatka. Despite using the recent coastline configuration and coastal topography, MARTIN *et al.* (2008) were able to untangle the identified deposits in relation to both tsunamis by investigating five different focal mechanisms and associated tsunamis for each event. MACINNES *et al.* (2010) simulated and compared 21 variations of the great 1952 Kamchatka earthquake and tsunami with eyewitness accounts and the alongshore distribution of tsunami deposits but again did not consider the former landscape conditions.

BONDEVIK *et al.* (2005a) focused on the Storegga Slide tsunami in the Norwegian Sea (almost 8000 years ago) and associated deposits along adjacent coasts. The authors numerically simulated different slide velocities and geometries to determine a slide mechanism and tsunami which can explain the detected deposits. However, hydrodynamic calculations were based on the present sea level and coastline configuration, hindering an accurate comparison and calibration

with event deposits. Furthermore, their numerical model did not include the onshore topography, which need the run-up heights to be estimated for comparison with tsunami sediments. HILL *et al.* (2014) expanded the simulations of the Storegga Slide tsunami by considering the regional palaeobathymetry but also refrained from calculating onshore hydrodynamics.

Synthesising field data and numerical simulations, SHISHIKURA *et al.* (2011), SAWAI *et al.* (2012), SUGAWARA *et al.* (2013) and NAMEGAYA & SATAKE (2014) investigated the impact of the seismogenic 869 AD Jōgan tsunami on the Sendai Bay in north-east Japan. The authors derived the contemporary coastline from an ash layer and reconstructed the local palaeotopography by subtracting tsunami deposits and overlying sediments from the present ground surface in their DEM. Based on this and on numerous possible focal mechanisms, the authors achieved close agreement between tsunami computations and field traces, which showed that the 869 Jōgan tsunami was an unambiguous forewarning of the 2011 Tōhoku tsunami killing an estimated 17 000 people.

As demonstrated by these studies, the calibration of numerical models by means of palaeotsunami field traces requires the simulation of a variety of possible tsunami waves, based on one or even several DEMs reproducing the probable former landscape conditions. This is confirmed by the fact that each tsunami shows an individual initial waveform, amplitude, wavelength, period and spatial extent and thus can mathematically be described by several wave theories (Chap. 2.5; TADEPALLI & SYNOLAKIS 1994, 1996; CARRIER *et al.* 2003; SYNOLAKIS *et al.* 2008; MADSEN & SCHÄFFER 2010; MADSEN *et al.* 2008; CONSTANTIN 2009; VORONINA 2011). Furthermore, coasts are one of the most dynamic landscapes on earth, frequently changing their shape over short periods of time (e.g. BIRD 2005; SCOTT 2005; THOM & COWELL 2005). This together with high tectonic activity, often observed in tsunami prone areas (cf. NOAA 2015a), emphasises the need to reconstruct and consider possible palaeolandscape conditions.

Based on sedimentary field traces and numerical simulations, this study investigates palaeotsunami events in the northern Gulf of Kyparissia (western Peloponnese, Greece; Figs. 1.1, 3.1a). Due to its immediate vicinity to the seismogenic subduction zone of the Hellenic Trench, the Gulf of Kyparissia is located within the most tsunamigenic region in the whole Mediterranean (Chap. 1.2). Accordingly, sedimentary and geomorphological traces of multiple tsunami impacts were detected within fieldwork along the northern Gulf of Kyparissia during recent years. This study takes account of tsunami field findings published for three key locations in the northern gulf coast: the Katakolo Pass (Figs. 3.2a, 4.3; VÖTT *et al.* 2010a, pp. 32–39; VÖTT *et al.* 2011b), the Epitalio Pass (Figs. 3.2b, 4.3; VÖTT *et al.* 2015) and the small valley of Kato Samiko (Figs. 3.2c, 4.3; WILLERSHÄUSER *et al.* 2012, 2015a). All three sites repeatedly fell victim to severe tsunami inundation during the last 8000 years, as is implied by vibracore data, geoelectrical measurements and geomorphological observations obtained in each key location (Chap. 4.3).

Taking account of the recent topography and two different palaeotopographies derived from neotectonic field data and literature and by applying the solitary and the *N*-wave theory, different tsunami scenarios are numerically simulated that reflect inundation as indicated by sedimentary and geomorphological tsunami field traces for the three key locations in the study area. In a further step, the appropriate scenarios are simulated with regard to the present topography and coastline configuration in view of a modern risk assessment. In this process, the current study methodologically analyses the implications of (i) different hydrodynamic boundary conditions and tsunami wave theories (*N*-wave effect) and (ii) different palaeo-/topographies for numerical simulation results and discusses the possibility and difficulty of numerical models to be calibrated and compared with palaeotsunami field traces.

3.2 Methodology

3.2.1 Digital elevation models

The DEMs employed in the numerical simulations were compiled from several different sources of data. Bathymetric data covering the deep sea to the coastline was taken from the

- Soviet Military Map on scales of 1:50 000, 1:200 000 and 1:500 000 (SCGC USSR 1983, 1986, 1988),
- British Admiralty Chart on the scale of 1:300 000 (UKHO 2010),
- Imray Chart on the scale of 1:190 000 (Imray Laurie Norie & Wilson Ltd 2008) and
- Eagle Ray Pilot Chart on the scale of 1:218 536 (Eagle Ray Publications 2010a).

The onshore topography is based on

- 90 m SRTM elevation data version 4 (JARVIS *et al.* 2008) in general and
- 30 m Google Earth elevation data (Google Inc. 2013; generated using the eSurvey Earth website <http://www.esurveyearth.com/>) in the area of interest.

Both, SRTM and Google Earth elevation data compare favourably with DGPS-heights measured in the field and altitude markings in topographical maps and therefore — among all freely available data — are most appropriate for hydrodynamic simulations in the northern Gulf of Kyparissia.

All elevation data was edited and compiled using ArcGIS Desktop 10.2 by Esri. Nautical charts were scanned and geocoded in order to digitise inherent depth contours and depth points. Bathymetric data was then triangulated and combined with SRTM and Google Earth onshore elevation data to create DEM_{rec}, reflecting the recent off- and onshore topography of the study area (Fig. 3.1a). The model has a maximum resolution of 30 m in the area of interest and covers the domain from 36.67° N to 38.00° N and from 20.11° E to 22.00° E, resulting in a total area of almost 148 km by 166 km.

Based on DEM_{rec}, two further elevation models were created in order to represent two possible palaeotopographies which might be associated with prehistorical tsunami impacts derived from field observations by VÖTT *et al.* (2010a, pp. 32–39), VÖTT *et al.* (2011b, 2015) and WILERSHÄUSER *et al.* (2012, 2015a) (Chap. 4.3). Due to great tectonic stress related to the nearby subduction zone of the Hellenic Trench, the study area experiences substantial vertical crust movements (PAPANIKOLAOU *et al.* 2007). This is most obvious from the Epitalio fault zone (EFZ) and adjacent fault zones running — clearly visible in the landscape — all along the northern Gulf of Kyparissia, almost parallel to the coastline (see white dashed-dotted lines in Fig. 3.1a). These faults separate “the uplifted blocks of the continent in the east from the subsided blocks of the ocean in the west” (PAPANIKOLAOU *et al.* 2007, p. 24), reflecting a complex interaction between ENE-WSW-oriented compression in the lower crust and ENE-WSW-oriented extension in the upper crust (PAPANIKOLAOU *et al.* 2007, p. 28; FOUNTOULIS & MARIOLAKOS 2008, p. 485).

Judging by correlating sea level indicators detected in vibracores from both sides of the EFZ and adjacent fault zones, VÖTT *et al.* (2015) deduce a minimum tectonic throw (vertical component of the dip separation) of 13 m for the last 6000 years or so. Based on these findings, the first palaeo-DEM, DEM₁₃, was created (Fig. 3.1b). Unaware of the continuity of the crust movements, DEM₁₃ reflects the palaeotopography at around 4000 BC at the earliest but possibly

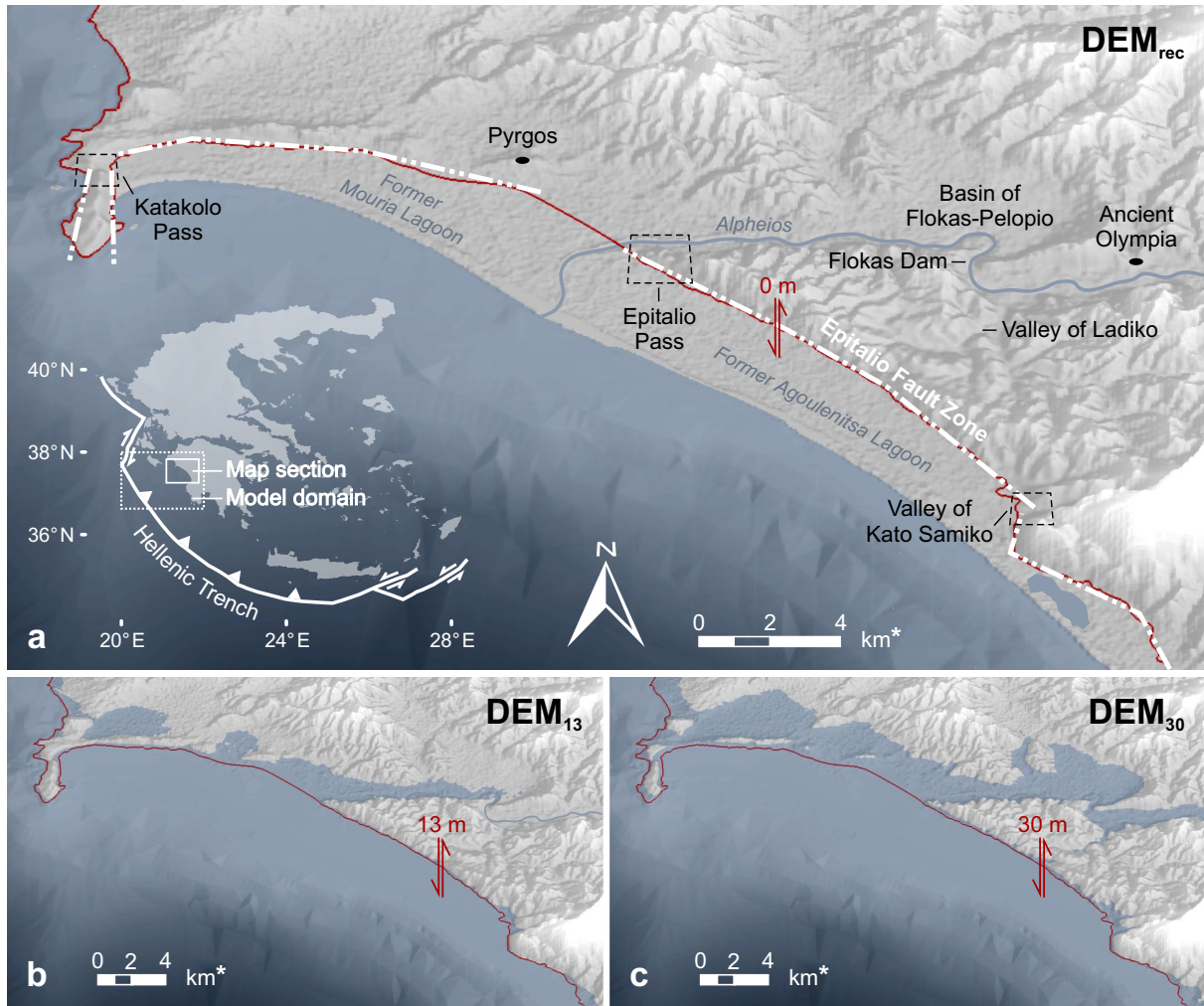


Figure 3.1: 3-D excerpts of three different digital elevation models of the northern Gulf of Kyparissia employed in the numerical simulations of this study. The geographical location of the gulf, the simplified regional tectonic regime and the numerical model domain is illustrated in the overview map in (a). DEM_{rec} (a) shows the recent topography of the northern Gulf of Kyparissia. DEM_{13} (b) reflects topographical conditions considering a 13 m throw along the Epitalio fault zone and adjacent fault zones (white dashed-dotted lines in (a)), which was derived by VÖTT *et al.* (2015) from the analysis of correlating sea level indicators detected in vibracores from either side of the faults for the last 6000 years or so. Based on a 30 m throw, which is suggested by PAPANIKOLAOU *et al.* (2007, p.24) for the same period, DEM_{30} (c) was created. The red line marks the exact position, where the throw of 13 m and 30 m respectively was considered in the palaeo-DEMs and derives from the present boundary between the coastal lowlands and adjacent Neogene hills (about 23 to 2.5 million years before present), which is associated with the approximate course of the faults (IGME 1982a,b; LEKKAS *et al.* 1992; PAPANIKOLAOU *et al.* 2007, pp. 26, 27). The blue surface indicates the sea level, which is the present sea level in all DEMs. Dashed boxes in (a) mark the map sections of the three key locations shown in Fig. 3.2 and Fig. 3.5. *Due to the 3-D view, the scale bar gives the true distance only in the centre of the figure.

also in younger times. The second palaeo-DEM refers to data published by PAPANIKOLAOU *et al.* (2007, p.24) suggesting a Holocene throw (covering the last 10 000 years) of up to 60 m and more along the fault zones in the northern Gulf of Kyparissia. According to this and assuming continuous crust movement without major coseismic uplift events, a 30 m-throw for the last 6000 years was chosen in DEM_{30} (Fig. 3.1c) to also examine the effects of a throw of this order on

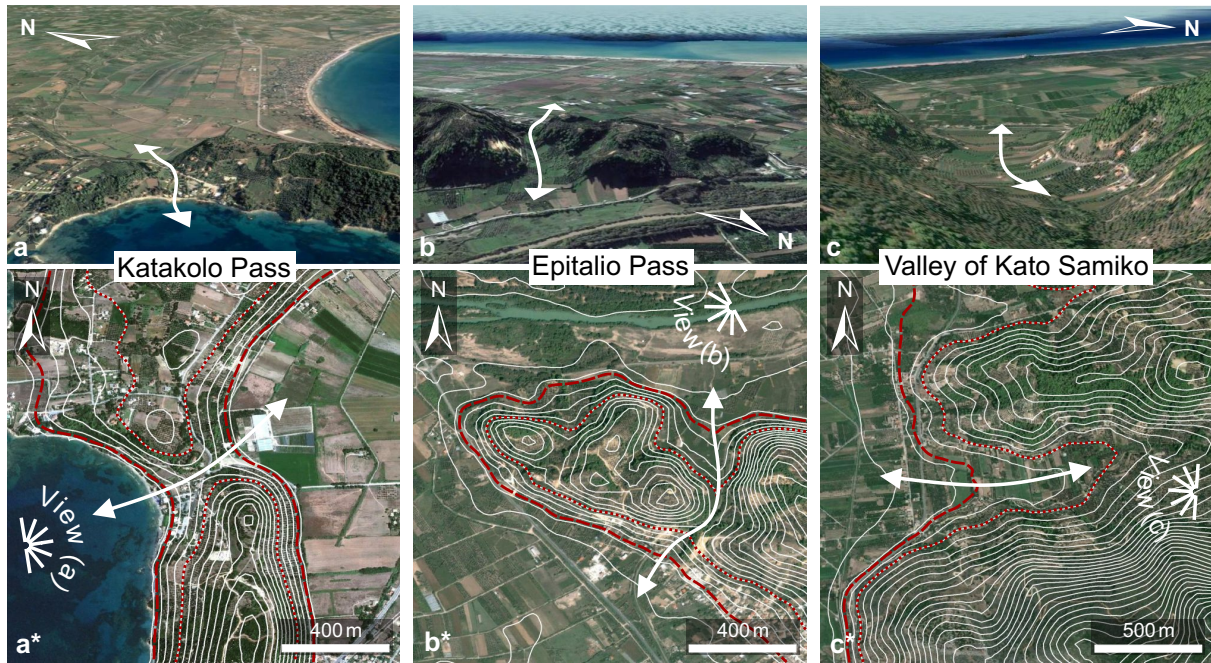


Figure 3.2: Oblique view (a, b, c) and plan view (a*, b*, c*) of the three key locations in the northern Gulf of Kyparissia (Google Inc. 2013). White arrows indicate tsunami inundation derived from sedimentary and geomorphological field traces found across the Katakolo Pass at the northernmost tip of the gulf (left; VÖTT *et al.* 2010a, pp. 32–39; VÖTT *et al.* 2011b), the Epitalio Pass at the entrance of the Alpheios River valley (centre; VÖTT *et al.* 2015) and in the valley of Kato Samiko to the north of Mount Lapithas (right; WILLERSHÄUSER *et al.* 2012, 2015a). For a summary of the field traces see Fig. 4.3 and Chap. 4.3. White contour lines (5 m steps in (a*) and (b*), 10 m steps in (c*)) represent the local topography. Contour lines have not been labelled due to varying absolute elevation in the three DEMs employed in the numerical model. The red dashed and red dotted line mark the coastline in DEM₁₃ and DEM₃₀ respectively. For the location of the three sites see the dashed boxes in Fig. 3.1a.

the onshore tsunami response (i.e. the intensity of the tsunami overland flow with regard to the inundation depth and flow velocity).

KRAFT *et al.* (2005, pp. 10, 11) and VÖTT (2007) suggest a relative sea level 2.5 m to 4.5 m lower than today for the time 4000 BC in the wider study area (cf. VÖTT *et al.* 2011b, pp. 274–276; WILLERSHÄUSER *et al.* 2015a). Recent results by VÖTT *et al.* (2015), however, reveal marked differences in local relative sea levels of at least 13 m for the past 6000 years. Consequently, it is not possible to say, to what extent the seaward part of the EFZ and adjacent fault zones was lowered and the landward part was uplifted or to deduce a reliable contemporary sea level. It was therefore decided to reduce present elevation data to the east of the faults by the amount of the neotectonic uplift (13 m and 30 m) and to take the present sea level as a base in all DEMs. This allowed to investigate the maximum impact of vertical crust movements on the onshore tsunami response in the numerical model.

Subtraction of elevation data in the palaeo-DEM_s caused a mismatch along the faults, which was compensated by an artificial ramp mediating between the new 0 m contour, i.e. the new coastline, and the –20 m contour in the sea (cf. Figs. 3.1a–c). This ramp represents a realistic inclination of the foreshore zone in the gulf, comparable to the recent nearshore bathymetry. A second ramp was incorporated to achieve a realistic decrease in water depth in the lower Alpheios River valley, that area becoming a marine inlet in both palaeo-DEM_s. The incorporation of the ramps is associated with the complete disappearance of the prominent dune belt and sandspit

system along the recent coastline of the northern Gulf of Kyparissia (cf. Figs. 3.1a–c). This is in agreement with palaeogeographical studies by KRAFT *et al.* (2005), which show that the mid-Holocene (circa 5000 years ago) coastline was located at the present boundary between the coastal lowlands and adjacent Neogene hills (see red line in Fig. 3.1a) and no major sandspit system existed.

3.2.2 Hydrodynamic numerical model

Tsunami simulations in this study were performed using Delft3D-FLOW, a hydrodynamic program module of the Delft3D suite (version 4.01.00.rc.12) developed by the Dutch Institute Deltares. Delft3D has been validated as an appropriate software tool in respect to tsunami propagation and run-up simulations (APOTSOS *et al.* 2011a) and was successfully employed for 1-D (APOTSOS *et al.* 2011d, 2012), 2-D (RÖBKE *et al.* 2012, 2013; CHACÓN-BARRANTES *et al.* 2013; PERÍÁÑEZ & ABRIL 2014a,b; Chaps. 4, 5) and 3-D tsunami simulations (GELFENBAUM *et al.* 2007; APOTSOS *et al.* 2009, 2011a,b,c).

Delft3D-FLOW solves the non-linear shallow water equations of unsteady flow and transport phenomena in two, meaning depth-averaged, or three dimensions (for this and the following see Deltares 2014, pp. 177–180). The module requires the flows' horizontal length and time scales to be significantly larger than the vertical scales, as is the case for tsunami waves. The non-linear shallow water equations are derived from the full Navier Stokes equations for incompressible fluids, as approximately represented by water. As the water column of the ocean can approximately be regarded as homogeneous concerning density, the two-dimensional, depth-averaged flow equations can be applied for tsunami simulations with Delft3D-FLOW — as was performed here.

Delft3D-FLOW takes account of wave-induced turbulence and through this of wave energy decay. In this context, the two most important processes are wave breaking and bottom roughness related to the oscillatory water motion at the sea bottom. Both processes are considered by introducing appropriate source terms in the turbulent kinetic energy and dissipation equations of the basic turbulence models behind Delft3D-FLOW (for further methodological details see Deltares 2014, pp. 236, 237; cf. APOTSOS *et al.* 2011a, pp. 2099, 2116, 2117). Owing to strong wave shoaling and the calculation of wave propagation in shallow water and of overland flow, in this study, the incorporation of energy dissipation due to wave breaking and bottom roughness played an important part.

The numerical model employed here solely considered the hydrodynamics of tsunami waves but did not simulate trigger mechanisms and the process of wave generation. Instead, various hydrodynamic boundary conditions were determined, which describe tsunami waves with different waveforms and wave amplitudes in order to examine differences in the hydrodynamic onshore response and to derive tsunami scenarios which agree with published field data (cf. PRITCHARD & DICKINSON 2008; GELFENBAUM *et al.* 2007; APOTSOS *et al.* 2009, 2011a,b,c,d, 2012). This approach was chosen due to the fact that there are no written accounts on the palaeotsunami events indicated by the field traces and, therefore, information on related trigger mechanisms is missing. Moreover, there is a plethora of possible tsunami trigger mechanisms ranging from nearby earthquakes and submarine landslides in the Hellenic Trench to remote earthquakes, submarine landslides and volcanic eruptions in the central and eastern Mediterranean (RÖBKE *et al.* 2013, pp. 74–77), which therefore could not be considered in this study.

The solitary and the *N*-wave theory were applied to determine the boundary conditions of the simulated tsunami waves. The suitability of using solitary waves to represent tsunamis has recently been questioned, since their time- and length-scales have been shown to differ from those of real tsunami waves (Chap. 2.5; MADSEN *et al.* 2008; MADSEN & SCHÄFFER 2010; CONSTANTIN

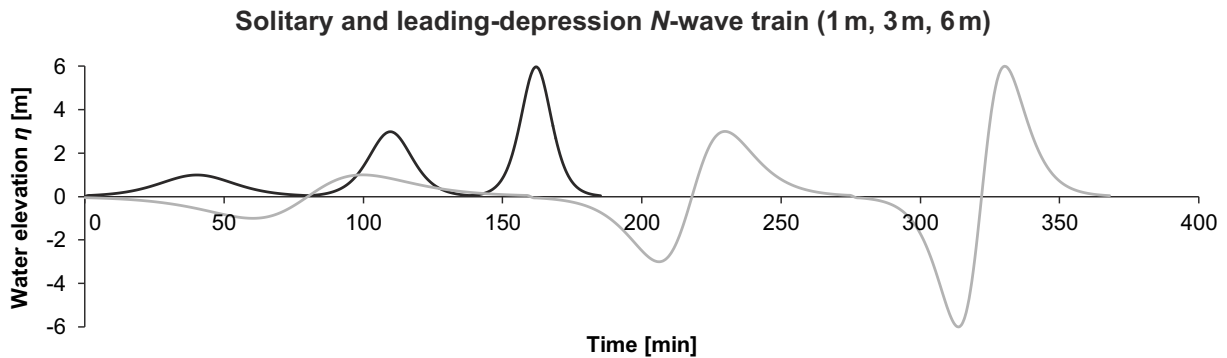


Figure 3.3: Tripartite solitary (black) and LDN-tsunami-wave train (grey) with increasing wave amplitudes from 1 m over 3 m to 6 m. Leaps between the separate LDN-waves around $\eta = 0$ m are due to the calculation of the wavelength according to the 1 %-definition by SYNOLAKIS *et al.* (2008, p. 2211). See text for further explanations.

2009; CONSTANTIN & JOHNSON 2008; APOTSOS *et al.* 2011a, p. 2114). Nevertheless, the solitary wave theory has often been employed in numerical and experimental tsunami simulations since it describes the ideal waveform to minimise energy loss during propagation and because tsunamis often appear as leading wave crests after long propagation distances (teletsunamis), as can be approximately represented by solitary waves (Chap. 2.5; DAILY & STEPHAN JR. 1952; SYNOLAKIS 1987; WIEGEL 1992, pp. 65–108; YEH *et al.* 1994; BRIGGS *et al.* 1995; GEIST 1997; WEISS 2004, p. 26; LEHFELDT *et al.* 2007; SYNOLAKIS *et al.* 2008; BALDOCK *et al.* 2009; APOTSOS *et al.* 2011a; TKALICH & DAO 2011, pp. 275–279; RÖBKE *et al.* 2012, 2013; LIANG *et al.* 2013b). Despite the ongoing debate on its validity for tsunami simulations, the solitary wave theory was considered in this study for the following reasons: (i) solitary waves still cannot be ruled out as a possible representation of real tsunami waves in certain conditions (see references above), (ii) in order to demonstrate the *N*-wave effect for a specific study area as this is a major objective of this study (Chap. 3.1) and (iii) to tie on former tsunami simulations for the northern Gulf of Kyparissia by RÖBKE *et al.* (2012, 2013), which are based on the solitary wave theory.

Based on the shallow water wave theory by KORTEWEG & DE VRIES (1895), the solitary wave of BOUSSINESQ (1872) was calculated according to the formulation by DEAN & DALRYMPLE (1991, pp. 314–317) (see Eq. 2.13; cf. LEHFELDT *et al.* 2007, pp. 112–114; RÖBKE *et al.* 2012, p. 562; RÖBKE *et al.* 2013, pp. 77–79). Solitary waves are only defined for positive values of the water elevation η , meaning, there is no negative wave component below the water surface. Thus, wave amplitude and wave height are the same for this waveform (Fig. 3.3). However, tsunamis have often been observed to cause coastal waters initially to recede before the first wave crest arrives. This means, tsunamis can also comprise a negative wave component — a wave trough. This is especially the case, when tsunamis are generated close to the coast and the propagation distance is too short for the waves to evolve into leading solitary waves (TADEPALLI & SYNOLAKIS 1994, p. 100; TADEPALLI & SYNOLAKIS 1996, p. 2143; MADSEN *et al.* 2008, pp. 2, 3; SYNOLAKIS *et al.* 2008, pp. 2202, 2203; LIANG *et al.* 2013b, pp. 144, 145), which may be the case in the Gulf of Kyparissia.

Based on these observations, TADEPALLI & SYNOLAKIS (1994, 1996) developed the *N*-wave theory, which describes dipolar tsunami waves showing both a wave trough and a wave crest (Chap. 2.5). The *N*-wave theory comprises leading-depression (LDN) and leading-elevation (LEN) *N*-waves, both isosceles or asymmetrical in shape. The hydrodynamic model of the current study was based on isosceles LDN-waves. The respective boundary conditions were calculated using the formulation by TADEPALLI & SYNOLAKIS (1994, p. 105) (see Eq. 2.14; cf.

CHACÓN-BARRANTES *et al.* 2013, pp. 15, 16).

In Eqs. 2.13 and 2.14, η -values were calculated based on the average water depth at the western model boundary (-2925 m) for every 60 s and then interpolated to define the respective waveform. In order to get the same length and time period for the wave crests for both waveforms, a corresponding α -value (0.0285 to 0.043) for each LDN-wave was determined.

As both, solitary and N -waves are non-periodic waves and therefore theoretically have an infinite wavelength, the starting and end point of each wave had to be set manually. According to SYNOLAKIS *et al.* (2008, p. 2211), the wavelength of solitary and N -waves can be defined as the distance between the points at the wave's front and tail where the water level is 1 % of the wave amplitude (cf. CHACÓN-BARRANTES *et al.* 2013, p. 16). According to this definition, the wave period in the simulations ranged from 46 min to 160 min (Fig. 3.3), which is representative of seismogenic or landslide generated tsunamis (BOLT *et al.* 1975, p. 134; HELAL & MEHANNA 2008, p. 788). For example, the wave energy of the Indian Ocean tsunami 2004 was concentrated in a broad frequency range with peaks at wave periods of between 30 min and 80 min and even longer wave periods for individual waves (RABINOVICH *et al.* 2011a, p. 4; RABINOVICH & EBLÉ 2015, p. 23). Besides the wave height, it is mostly the wavelength or wave period respectively that determine the onshore response of tsunami waves. Nevertheless, in order to produce a reasonable number of scenarios, for each wave, a fixed wavelength/wave period, predetermined by the solitary wave of BOUSSINESQ (1872) for a specific wave height (Eq. 2.13) was applied and only the wave height itself was varied (see below).

As tsunamis commonly consist of a series of waves, both wave types were simulated as tripartite tsunami wave trains (Fig. 3.3). Rarely is the first wave the highest of the wave train, it is usually one of the following waves (MURTY 1977, p. 2; BRYANT 2008, pp. 27, 28; BORMANN 2012, pp. 5, 6). Thus, wave trains of increasing wave amplitude were created, with the third and last wave being the highest one. This further allowed to consider the effects of preceding waves on the inundation extent by the following higher waves (cf. APOTSOS *et al.* 2009; APOTSOS *et al.* 2011b, p. 14; APOTSOS *et al.* 2012) — a feature important for inlet-shaped coasts or in coastal river valleys (see Alpheios River valley in the Fig. 3.1), where tsunami waters may gather during tsunami landfall and increase the total onshore response. In all scenarios, wave trains started from the western model boundary (36.67° N to 38.00° N and 20.11° E; see overview map in Fig. 3.1a) because tsunamis from a western direction caused the strongest onshore response in previous studies for the northern Gulf of Kyparissia based on solitary waves (RÖBKE *et al.* 2012, 2013).

Various tsunami wave trains with different increasing wave amplitudes were simulated for both waveforms in order to find waves which cause inundation of all three key locations at once and agree with field traces described by VÖTT *et al.* (2010a, pp. 32–39), VÖTT *et al.* (2011b, 2015) and WILLERSHÄUSER *et al.* (2012, 2015a) (Chap. 4.3). This was done on the basis of DEM₁₃, reflecting the most probable landscape conditions for the time of palaeotsunami inundation (Chap. 3.2.1). In the case of the solitary wave, a set of 1 m, 3 m and 6 m waves (Fig. 3.3) produced the intended onshore hydrodynamics. The same wave amplitudes were applied to the LDN-wave train to also investigate the onshore response of this waveform. Both wave trains were then also simulated on the basis of DEM₃₀ and DEM_{rec} in order to discover the local effects of different palaeo-/landscape conditions on the onshore tsunami response and to examine the present day tsunami risk. Combining two different waveforms and three different DEMs, a total of six tsunami scenarios are presented in this chapter (Tab. 3.1).

Regarding the findings by TSELENTIS *et al.* (2010) and YALÇINER *et al.* (2014), who both simulated tsunamis in the Gulf of Kyparissia originating from local fault zones, tsunami wave trains of the order chosen in this investigation must be regarded as extreme (Imamura-Iida

3 Effects of different boundary conditions and palaeo-/topographies

Table 3.1: Overview of the six presented tsunami scenarios, resulting from three different DEM-configurations created for the Gulf of Kyparissia and two different wave types.

Scenario	DEM employed	Wave type	Amplitudes of separate waves	Wave train from
Scenario _{13S}	DEM ₁₃	Solitary wave	1 m, 3 m, 6 m	West
Scenario _{13N}	DEM ₁₃	LDN-wave	1 m, 3 m, 6 m	West
Scenario _{30S}	DEM ₃₀	Solitary wave	1 m, 3 m, 6 m	West
Scenario _{30N}	DEM ₃₀	LDN-wave	1 m, 3 m, 6 m	West
Scenario _{recS}	DEM _{rec}	Solitary wave	1 m, 3 m, 6 m	West
Scenario _{recN}	DEM _{rec}	LDN-wave	1 m, 3 m, 6 m	West

magnitude < 1 ; IIDA 1963; cf. Tab.2.1) rather than moderate or weak. However, the physiogeographical background—especially the tectonic activity and distinct submarine relief—of the Gulf of Kyparissia and the central and eastern Mediterranean in general as well as historical accounts (Chap.1.2; RÖBKE *et al.* 2013, pp.74–77) suggest these scenarios to be realistic in the case of huge local submarine landslides, nearby or remote extraordinary strong earthquakes or even remote volcanic eruptions.

In all simulations, the computational grid resolution was 900 m in the open sea, 60 m in the nearshore zone and 30 m in the main study area. The hydrodynamic boundary conditions of the model were based on the Riemann invariant in order to minimise unintended wave reflections along the open boundary (for further methodological details see VERBOOM & SLOB 1984; Deltares 2014, pp.205–209; cf. APOTSOS *et al.* 2011b, p.16). According to the CFL stability condition, the calculation step was set to 0.3 s (cf. Deltares 2014, pp.39, 40, 279, 280). Depending on the wave period of the separate waves, the total simulation time was 6 h in the solitary wave scenarios and 9 h in the LDN-wave scenarios.

Finally, the terrain roughness of the study area was considered using appropriate values of Manning’s n coefficient. Judging by field surveys, photographs and Google Earth satellite images (Google Inc. 2013), n -values from minimum 0.015 (sea floor, pastures) over 0.03 (moderate vegetated or developed land) to maximum 0.07 (densely forested landscapes) were attributed to the respective areas (cf. HILLS & MADER 1997, p.388; BRYANT 2008, pp.45, 46).

3.3 Results

This section presents the inundation of the northern Gulf of Kyparissia caused by each wave of the tripartite solitary and LDN-tsunami-wave trains in all six scenarios (Tab.3.1). Snapshots show water levels in m a.s.l. at the time of the maximum onshore inundation extent of the individual waves and also during the run-up of the third wave of each wave train, which are the maximum water levels in each scenario. Furthermore, water levels and flow vectors associated with the inundation of the Katakolo Pass, the Epitalio Pass and the valley of Kato Samiko are illustrated in detail for Scenario_{13S} and Scenario_{13N}. Due to the vast area, a single snapshot cannot reflect the maximum inundation extent or the maximum water levels recorded for the complete study area. Consequently, the snapshots chosen were those that displayed the discrepancies between the onshore response of each wave most clearly.

3.3.1 Solitary and LDN-wave train based on DEM₁₃

Overview and general characteristics

The 1 m solitary wave in Scenario_{13S} causes a rise in water level of up to 3.5 m a.s.l. in the Alpheios River valley. Here and in the Pyrgos area, tsunami waters reach their maximum on-shore extent (Fig. 3.4a; see Fig. 3.1a for the labelling of places). In Scenario_{13N}, the maximum water level during the run-up of the first wave is nearly twice as large (6.5 m) as in Scenario_{13S}. However, Fig. 3.4b shows the maximum inundation extent of the first LDN-wave to be insignificantly larger. This is due to the hilly landscape along the coastline in DEM₁₃, which prevents water masses from spreading easily and therefore results in higher water levels rather than extensive flooding for stronger wave impulses.

The second solitary wave strikes the study area about 1 h after the first wave arrives. Due to the steep continental margin in the gulf, strong wave shoaling is observed before landfall, increasing the wave amplitude from 3 m at the model's western boundary to almost 9 m close to the coastline. The maximum run-up heights are of the order of 12 m to the north of Pyrgos and in the basin of Flokas-Pelopio (Fig. 3.4c). Inundation by the second LDN-wave in Scenario_{13N} is even more extensive (here and in the following in view of the inundation depth and flow velocity), which is most obvious to the north of the former Mouria Lagoon and in the basin of Flokas-Pelopio (Fig. 3.4d). The maximum run-up height — almost 27 m to the north of Pyrgos — is more than doubled compared to Scenario_{13S}.

Differences between the onshore response of both wave types are most evident in view of the third, 6 m waves. Both waves are characterised by strong wave shoaling in the gulf and generate distinct reflection waves along the hilly coastlands landwards from the former lagoons (Figs. 3.4e,f). However, water levels during the run-up of the third wave are much higher in Scenario_{13N} (40 m) compared to Scenario_{13S} (27 m). This further results in deeper inland penetration in Scenario_{13N} all along the coast and especially in the Alpheios River valley, where waters even reach the ancient site of Olympia (cf. Figs. 3.4g,h).

Key locations

Fig. 3.5a shows the maximum inundation dynamics of the third wave in Scenario_{13S} located across the Katakolo Pass lying almost 7 m a.s.l. in DEM₁₃ (the first and second wave do not cause the pass to flood). Although the third solitary wave originally approaches from the west, waters stream across the pass from a north-eastern direction. This is due to (i) strong diffraction and refraction of the wave around Cape Katakolo located 4 km to the south, (ii) funnelling of tsunami waters in the Bay of Katakolo and (iii) the impact of reflection waves generated along the hilly coastlands between Epitalio and Kato Samiko. Owing to the steep gradient of water level between both sides of the pass, the tsunamigenic overflow shows a high depth-averaged flow velocity, ranging from 6.3 m s^{-1} at the top of the pass to 17.8 m s^{-1} further west. The maximum inundation depth across the pass during overflow is 9.5 m. As another important feature of the local hydrodynamics, a persistent lee eddy emerges on the western side of the pass during overflow. In Scenario_{13N}, the 3 m LDN-wave causes similar hydrodynamics in the area of the pass as described for the 6 m solitary wave. The 6 m LDN-wave has an even stronger impact on the site, as shown by Fig. 3.5b: Inundation is no longer restricted to the narrow pass, but the entire area to the north is also affected. The maximum inundation depth across the pass is about 21 m; the maximum depth-averaged flow velocity increased from 17.8 m s^{-1} to 23 m s^{-1} compared to Scenario_{13S}.

There are several parallels between the inundation dynamics observed for the Katakolo Pass and the Epitalio Pass, circa 17 m a.s.l. in DEM₁₃: Only the third wave affects the Epitalio Pass in

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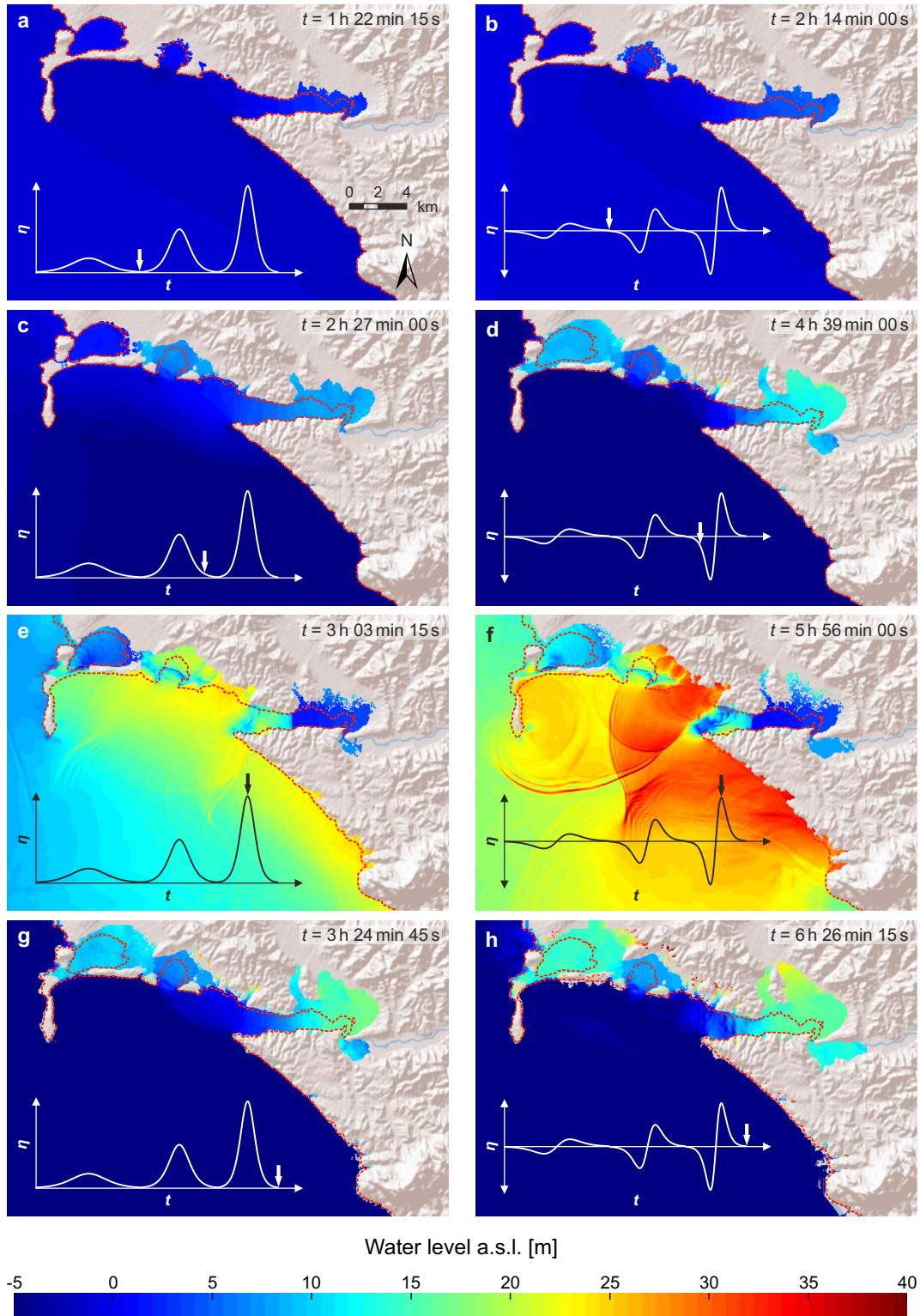


Figure 3.4: Snapshots documenting the inundation of the northern Gulf of Kyparissia caused by a solitary (Scenario_{13S}, left column) and LDN-tsunami-wave train (Scenario_{13N}, right column) based on DEM₁₃ considering a 13 m throw along the EFZ and adjacent faults since the mid-Holocene (Fig. 3.1b). (a) to (d), (g) and (h) show the inundation extent and associated water levels after the landfall of each separate wave. The maximum water levels during the run-up of the third, 6 m wave of each wave train are illustrated in (e) and (f). The red dashed line presents the coastline in DEM₁₃.

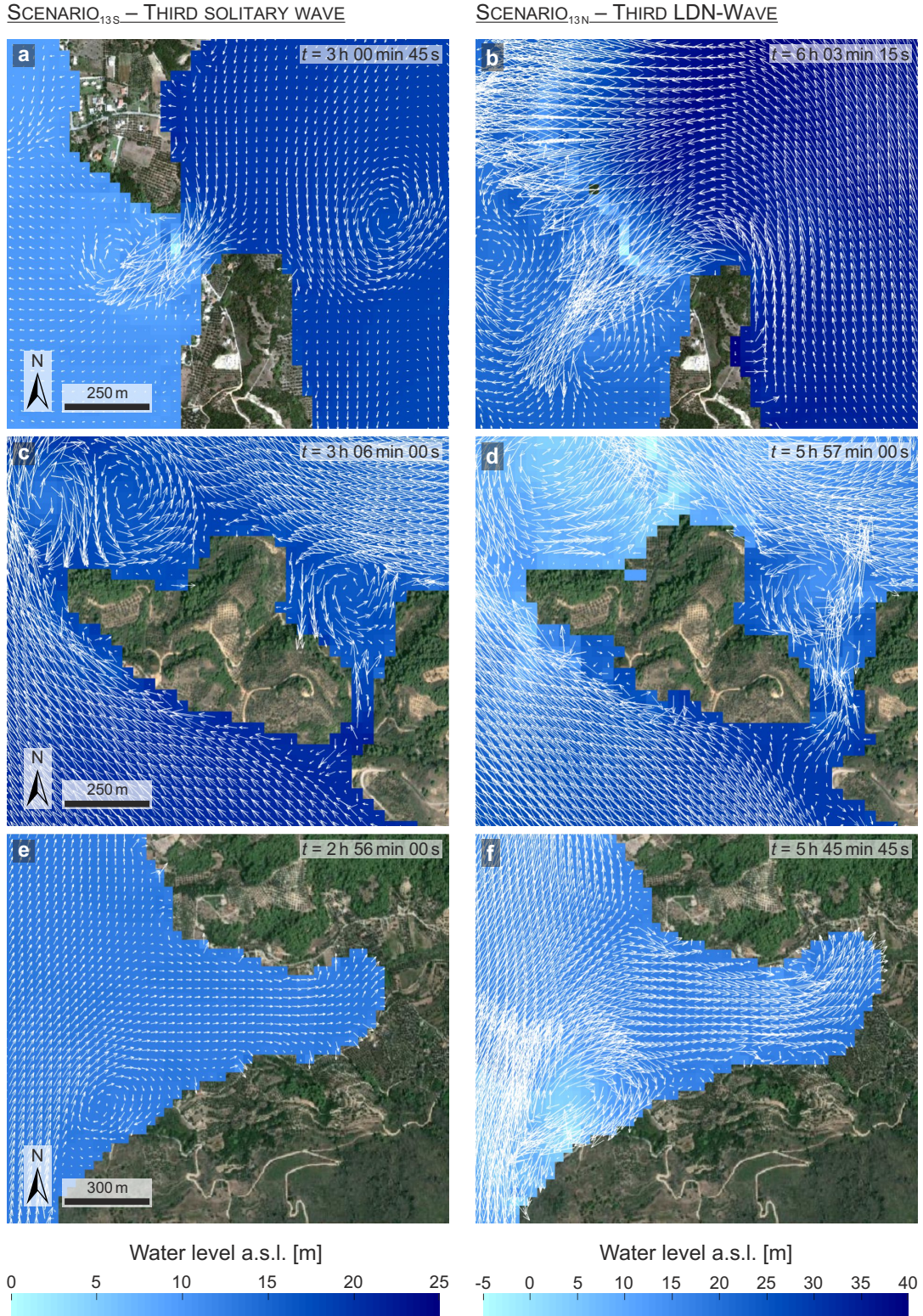


Figure 3.5: Snapshots documenting the inundation of the Katakolo Pass (a, b) (cf. Fig. 4.6), Epitalio Pass (c, d) (cf. Fig. 4.9) and valley of Kato Samiko (e, f) (cf. Fig. 4.12) during the run-up of the third, 6 m wave of a solitary (Scenario_{13S}, left column) and LDN-tsunami-wave train (Scenario_{13N}, right column) based on DEM₁₃ (Fig. 3.1b). Besides the water level a.s.l. (blue scale bar), the depth-averaged flow direction and velocity is indicated by white vectors. For the location of the three map sections see dashed boxes in Fig. 3.1a, for aerial views of each site Fig. 3.2.

Scenario_{13S} (Fig. 3.5c). Inundation is not induced by the main wave crest, but reflection waves from the Bay of Katakolo (Fig. 3.4e) cause waters to overflow the pass and generate a large eddy in the lee. The maximum inundation depth and the maximum depth-averaged flow velocity associated with the overflow of the pass are 4.2 m and 6.2 m s^{-1} respectively. Overflow dynamics across the Epitalio Pass are even stronger for the 6 m LDN-wave (Fig. 3.5d), which directly inundates the pass intensified by reflection waves, which are clearly visible in Fig. 3.4f. Both, the maximum inundation depth (10.5 m) and maximum depth-averaged flow velocity (13 m s^{-1}) more than doubled compared to Scenario_{13S}. Due to high water levels in the Alpheios River valley, the flow direction across the pass reverses with time and the pass is also overflowed from north to south-west.

The inundation dynamics observed in the valley of Kato Samiko, roughly 2 m b.s.l. (below sea level) in DEM₁₃, are less complex. In both, Scenario_{13S} and Scenario_{13N}, each wave generates an almost linear run-up and backflow in the valley, accompanied by an eddy at the southern entrance of the valley during the landfall of the third wave (Figs. 3.5e,f). As before, inundation by the LDN-waves is stronger, especially in the case of the 6 m wave. The maximum inundation depth is recorded during the run-up of the third wave, which is 28 m in Scenario_{13S} but 42 m in Scenario_{13N} in the central valley. However, the maximum depth-averaged flow velocity at coring site SAM 1 relates to the backflow of the third wave and is about half the value for the solitary wave (6 m s^{-1}) than for the LDN-wave (12 m s^{-1}).

3.3.2 Solitary and LDN-wave train based on DEM₃₀

Overview and general characteristics

In Scenario_{30S} and Scenario_{30N}, the 1 m waves mainly inundate the area to the north of the basin of Flokas Pelopio (Figs. 3.6a,b). Although the maximum water level observed for the solitary wave in the Alpheios River valley (3 m) is only half the level recorded for the LDN-wave (6 m), the area of inundation is nearly the same in both scenarios as tsunami waters are enclosed by the hilly coastlands. The absence of wide coastal flats hindering a vast inundation extent is a feature even more evident for DEM₃₀ than for DEM₁₃ (cf. Chap. 3.3.1).

Strong wave shoaling causes the second solitary wave to inundate the area to the north of the former Mouria Lagoon, around Pyrgos and the basin of Flokas-Pelopio (Fig. 3.6c). Tsunami waters further penetrate the Alpheios River valley passing the ancient site of Olympia. The same pattern is found for the second LDN-wave, although the inundation is clearly more dynamic (Fig. 3.6d). Whereas the onshore response of the third solitary wave is slightly stronger compared to the second LDN-wave (cf. Figs. 3.6d,e,g), inundation dynamics on an unprecedented scale are observed for the third wave in Scenario_{30N} (Figs. 3.6f,h). Due to strong wave shoaling, run-up-heights in excess of 40 m are recorded all along the coast with the exception of the Alpheios River valley. Halfway between Epitalio and Kato Samiko, tsunami waters even climb up the hilly coastlands, pass the valley of Ladiko and finally enter the Alpheios River valley to the south of the Flokas Dam. In contrast to the DEM₁₃-Scenarios, for both the third solitary and LDN-wave in the DEM₃₀-Scenarios, reflection waves are less intense due to the irregular coastline in DEM₃₀ (cf. Figs. 3.4e,f, 3.6e,f).

Key locations

In the DEM₃₀-Scenarios, the Katakolo Pass lies about 10 m b.s.l. and therefore is covered by water during the complete simulation. As a result of this, strong currents through the pass are associated with the passage of each wave of the solitary and the LDN-wave train. Contrary to

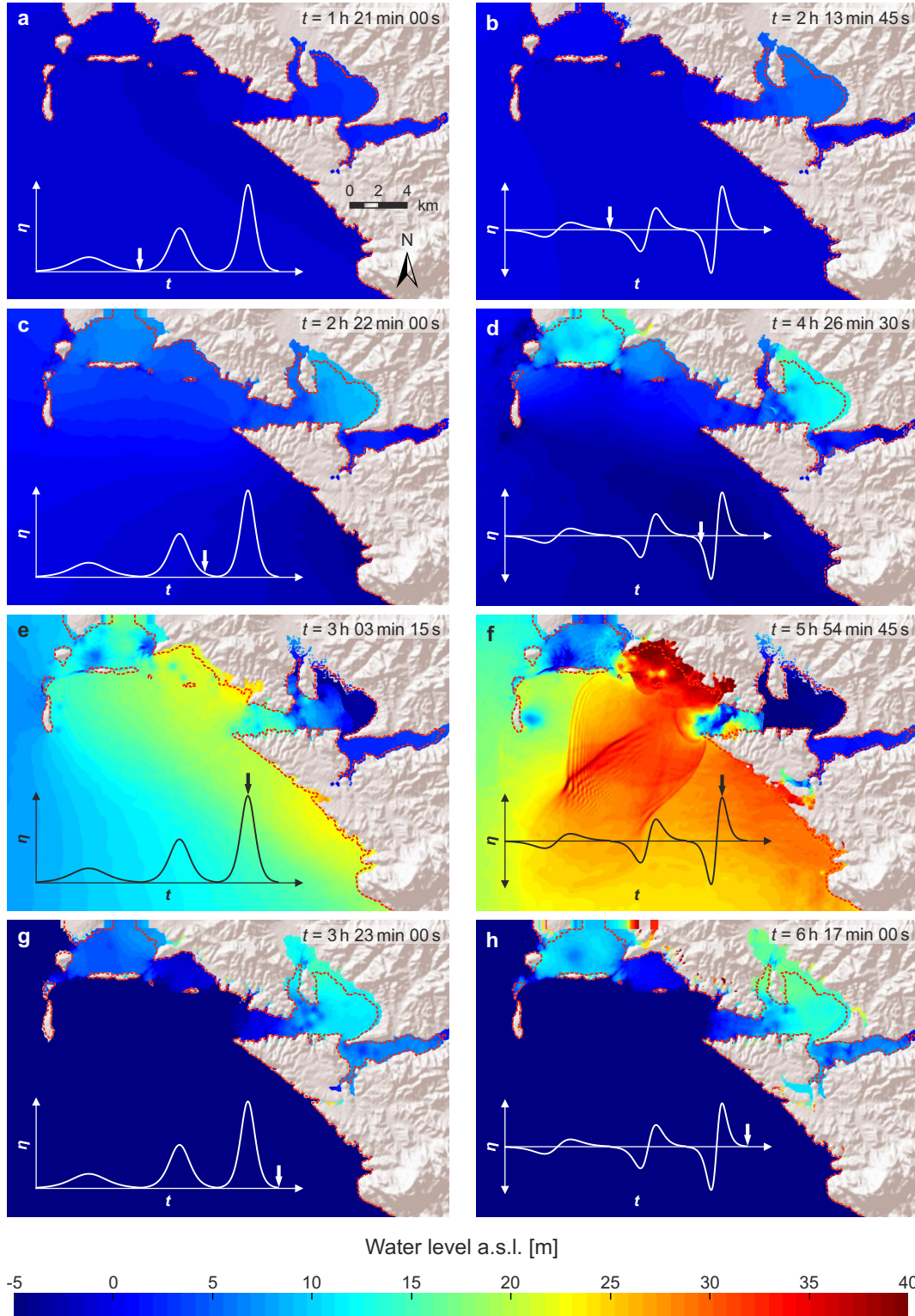


Figure 3.6: Snapshots documenting the inundation of the northern Gulf of Kyparissia caused by a solitary (Scenario_{30S}, left column) and LDN-tsunami-wave train (Scenario_{30N}, right column) based on DEM₃₀ considering a 30 m throw along the EFZ and adjacent faults since the mid-Holocene (Fig. 3.1c). (a) to (d), (g) and (h) show the inundation extent and associated water levels after the landfall of each separate wave. The maximum water levels during the run-up of the third, 6 m wave of each wave train are illustrated in (e) and (f). The red dashed line presents the coastline in DEM₃₀.

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the DEM₁₃-Scenarios, here, the flow direction of the currents changes from E-W to W-E several times. As before, eddies emerge on both sides of the pass. In both DEM₃₀-Scenarios, the second and third wave further inundate the entire area to the north of the pass, comparable to the situation described for the third LDN-wave in Scenario_{13N} (cf. Fig. 3.5b). In spite of the same flow characteristics for Scenario_{30S} and Scenario_{30N}, inundation arising from the LDN-wave train is once again stronger.

The Epitalio Pass lies just above sea level in DEM₃₀ and is flooded by the second and third solitary and LDN-waves. Inundation by the 3 m solitary wave is similar to the hydrodynamics observed for the 6 m LDN-wave in Scenario_{13N} (cf. Fig. 3.5d), meaning that the pass is firstly overflowed from south-west to north and subsequently in the opposite direction. The same pattern is found for the second LDN- and the third solitary wave in the DEM₃₀-Scenarios but with more intense dynamics. During the landfall of the third LDN-wave, the complete area to the north-west of the Epitalio Pass is in deep flood. Here again, strong reflection waves play an important part.

The inundation of the valley of Kato Samiko — circa 19 m b.s.l. in DEM₃₀ — follows the same pattern as described for DEM₁₃ (cf. Figs. 3.5e,f). Each wave of both wave trains causes an almost linear run-up and subsequent backflow in the valley. As before, the impact of the LDN-wave train is stronger compared to the solitary wave train.

3.3.3 Solitary and LDN-wave train based on DEM_{rec}

Overview and general characteristics

Due to the extensive dune belt and sandspit system (up to 600 m wide and 15 m high) along the northern gulf, the recent coastlands are well protected from the first wave in Scenario_{recS} with the exception of the former Mouria Lagoon (Fig. 3.7a). In contrast, the first LDN-wave causes vast inundation of the former Mouria and Agoulenitsa Lagoon (Fig. 3.7b). Significant differences are also found between the 3 m solitary and LDN-wave. Whereas the solitary wave inundates a 4 km to 6 km wide belt parallel to the coastline, the LDN-wave penetrates deeper inland, especially around Pyrgos and in the Alpheios River valley (cf. Figs. 3.7c,d). The onshore response of the third wave in Scenario_{recS} is slightly stronger compared to the second LDN-wave (cf. Figs. 3.7d,g). This is most evident from the inundation extent around Pyrgos and the greater inland penetration in the Alpheios River valley. The maximum run-up of the wave (25 m) is recorded in the valley of Kato Samiko (Fig. 3.7e).

As found in all scenarios, in the DEM_{rec}-Scenarios, the 6 m LDN-wave has the strongest impact on the study area, causing maximum run-up heights of 35 m in several small valleys between Epitalio and Kato Samiko and to the north-east of Pyrgos (Fig. 3.7f). Once again, strong reflection waves are generated. The lowlands to the north of the former Mouria Lagoon and northwards of Pyrgos receive strong flooding (Fig. 3.7h). Finally, tsunami waters penetrate into the Alpheios River valley and inundate large parts of the basin of Flokas-Pelopio and even the area beyond the Flokas Dam.

Key locations

With an elevation of roughly 20 m a.s.l. in DEM_{rec}, the Katakolo Pass is not inundated by the third wave in Scenario_{recS}, although a large eddy emerges at the eastern edge of the pass during the run-up. The third LDN-wave, however, produces similar hydrodynamics across the pass as described for the third wave in Scenario_{13S} (cf. Fig. 3.5a) but less intense. The Epitalio Pass is not affected by any of the wave trains in the DEM_{rec}-Scenarios due to its high present altitude

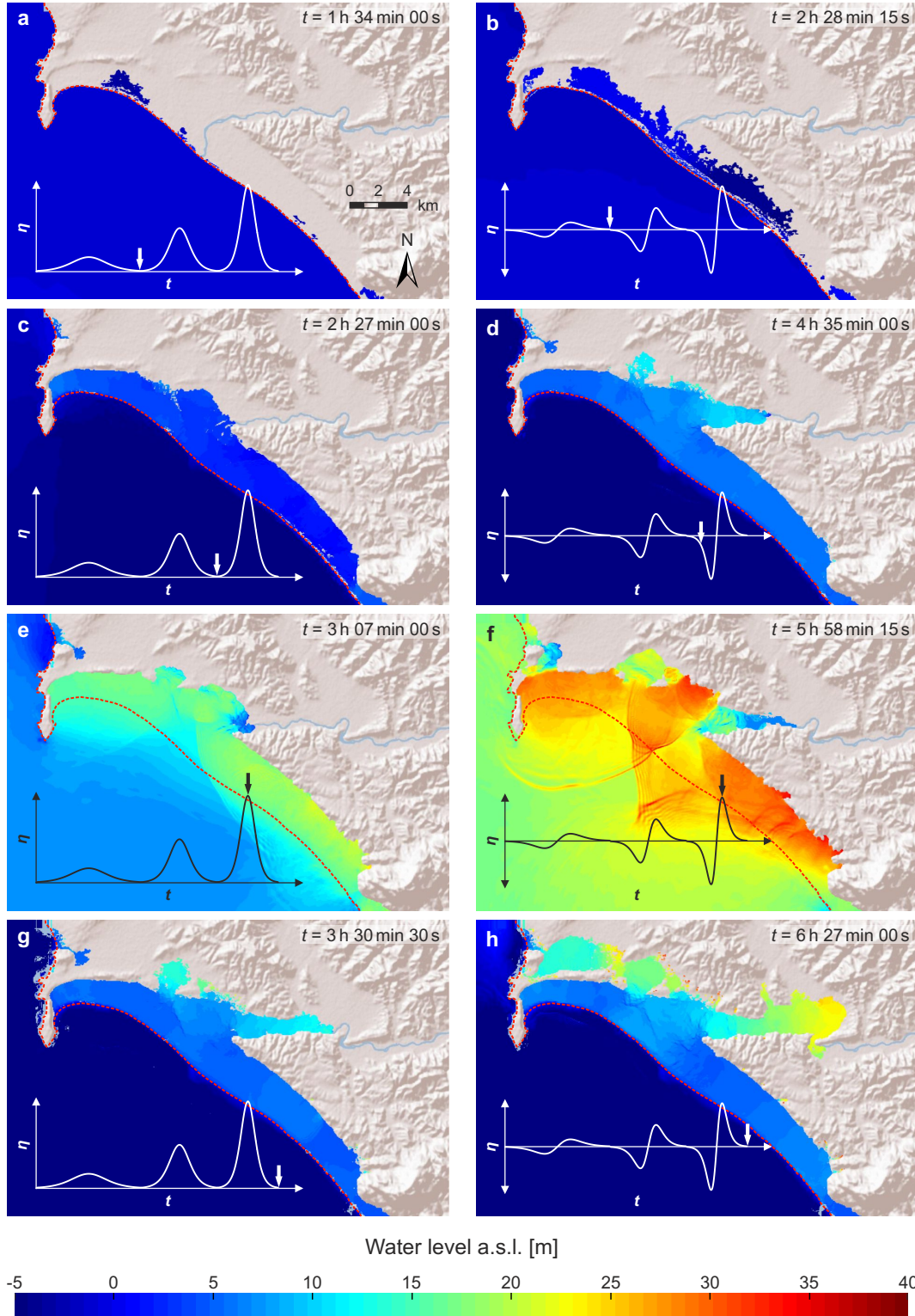


Figure 3.7: Snapshots documenting the inundation of the northern Gulf of Kyparissia caused by a solitary (Scenario_{rec S}, left column) and LDN-tsunami-wave train (Scenario_{rec N}, right column) based on DEM_{rec} reflecting the present topography of the study area (cf. Fig. 3.1a). (a) to (d), (g) and (h) show the inundation extent and associated water levels after the landfall of each separate wave. The maximum water levels during the run-up of the third, 6m wave of each wave train are illustrated in (e) and (f). The red dashed line presents the coastline in DEM_{rec}.

of about 30 m a.s.l. Nevertheless, significant eddies emerge to the north of the pass with the third solitary and LDN-wave forcing their way through the Alpheios River valley (Figs. 3.7e–h). Finally, tsunami waters of the third solitary and LDN-wave deeply funnel into the valley of Kato Samiko, which is almost 11 m a.s.l. in DEM_{rec}. As before, a linear run-up and backflow is observed for both waves with persistent eddies at the valley’s entrance (cf. Figs. 3.5e,f). These flow phenomena are once again stronger for the LDN-wave.

3.4 Discussion

The six tsunami scenarios show several common hydrodynamic features that increase the onshore response in the study area, or cause inundation dynamics not necessarily expected from the regional setting:

- In all scenarios, strong wave shoaling appears owing to the deep gulf waters (up to 1850 m) and the steep continental margin inherent in all three DEMs employed. As a result, wave amplitudes frequently more than triple from the model’s boundary to the shelf area.
- Strong diffraction and refraction of the waves around Cape Katakolo is observed, independent from the waveform and topography, forcing waters to strike the sheltered Bay of Katakolo and to inundate the Katakolo Pass from an eastern direction.
- Significant funnelling of tsunami waters in the Alpheios River valley and subsequent divergence in the basin of Flokas-Pelopio are features all scenarios have in common, resulting in deep inland penetration and vast inundation in the river valley.
- Reflection waves are generated along the steep slopes of the hilly coastlands in all three DEMs (e.g. Fig. 3.7f), which intensify or even cause the inundation of the Katakolo and Epitalio Pass.

Certainly, most of these hydrodynamic features depend on the western propagation direction of tsunami waves chosen in the numerical model. As was indicated in Chap. 3.2.2, RÖBKE *et al.* (2012, 2013) showed that—based on the recent topography—solitary tsunami waves from a western direction cause the strongest onshore response in the northern Gulf of Kyparissia, whereas inundation by waves from the south-west and south is less intense. Furthermore, for both these directions (SW and S), wave shoaling, diffraction and refraction around Cape Katakolo and the generation of reflection waves are less distinct due to the combined effects of the propagation direction and the local on- and offshore topography.

Despite the hydrodynamic similarities mentioned above, there are also marked differences between the six tsunami scenarios, related to the different waveforms and topographies. The LDN-waves consistently have a stronger onshore response in the study area compared to the solitary waves. This is most evident from the comparison of the 3 m LDN-waves and the 6 m solitary waves, both having a similar impact for all three DEMs. This phenomenon is known as the *N*-wave effect, which is mainly due to the steeper wave crest of LDN-waves increasing the inertia of the wave (TADEPALLI & SYNOLAKIS 1994, p. 109). In contrast, theoretical studies by LIANG *et al.* (2013b) showed leading-elevation (LEN) *N*-waves having a smaller onshore response compared to LDN- and solitary waves, although their wave crests have the same steepness as solitary waves. This arises from the subsequent wave trough of LEN-waves, which reduces the length of the preceding wave crest and therefore shortens the time period of high water level during the run-up compared to solitary waves (LIANG *et al.* 2013b, pp. 150, 151). The simulation results show that the *N*-wave effect becomes stronger with increasing wave amplitude since

the wave steepness increases disproportionately with amplitude. The N -wave effect has been investigated in theory by numerical studies (TADEPALLI & SYNOLAKIS 1994; APOTSOS *et al.* 2011a; LIANG *et al.* 2013b) but has never been numerically demonstrated in view of a specific coast before.

Considering the short distance between the Gulf of Kyparissia and the major tsunamigenic zone of the Hellenic Trench (< 100 km) and the fact that LDN-waves are primarily caused by tsunamigenic earthquakes close to the coastline, the appearance of LDN-waves in the gulf is more likely compared to the weaker solitary waves. In the case of teletsunamis however, solitary or even LEN-waves might be the more realistic waveform due to the long propagation distance of such tsunamis (Chaps. 2.5, 3.2.2).

The stronger onshore response of the LDN-waves compared to the solitary waves is clearest for DEM_{rec} (Fig. 3.7) but less distinct in DEM₁₃ (Fig. 3.4) and even lesser distinct in DEM₃₀ (Fig. 3.6). This is due to the vast coastal flats of the recent topography, emphasising the differences in the onshore inundation extent more clearly. Thus, the N -wave effect for the palaeo-DEMs, where large lowlands are missing, is reflected by high water levels rather than by an increased area of inundation.

The different palaeo-/topographies have an even stronger affect on the onshore tsunami response than the different hydrodynamic boundary conditions. This is related to the vast lowlands and the prominent dune belt and sandspit system all along the recent coastline, which result in significant topographical changes when vertical crust movements are considered (Chap. 3.2.1). In particular, the dune belt and sandspit system completely disappear in the palaeo-DEMs and this removes an important line of protection against tsunami waves (cf. Figs. 3.1a–c, 3.4a,b, 3.6a,b, 3.7a,b). Furthermore, the coastline is quite irregular in the palaeo-DEMs and there are several small coves, especially in DEM₃₀. This severely affects the onshore response, in that an irregular coastline breaks tsunami waves more effectively and therefore reduces the power of reflection waves. The latter are of great importance for the local hydrodynamics as observed in Chap. 3.3.

Given the great age of tsunami deposits in the northern Gulf of Kyparissia (Chap. 4.3), the created palaeo-DEMs are only an approximate representation of the contemporary landscape conditions and do not reflect realistic small-scale palaeotopography. Nevertheless, the major topographical changes relevant to the onshore tsunami response in the study area, i.e. vertical tectonic movements, coastline moving and disappearance of the prominent dune belt and sandspit system, have been considered in the palaeo-DEMs based on reliable palaeogeographical studies by KRAFT *et al.* (2005), PAPANIKOLAOU *et al.* (2007) and VÖTT *et al.* (2015, pp. 23–28) (Chap. 3.2.1).

The DEM₁₃-Scenarios show the best agreement with the field traces found at the Katakolo and Epitalio Pass and in the valley of Kato Samiko (VÖTT *et al.* 2010a, pp. 32–39; VÖTT *et al.* 2011b, 2015; WILLERSHÄUSER *et al.* 2012, 2015a; Chap. 4.3). At least one of the waves in each wave train resulted in significant inundation of the sites. Judging by the recorded inundation depths and flow velocities (Chap. 3.3.1), the 3 m and 6 m solitary and LDN-waves are the most appropriate candidates to have a clear sedimentary and geomorphological impact on the three key locations. Also the simulation results based on DEM₃₀ are consistent with field observations. For all three locations, a maximum wave amplitude of 3 m, both of the solitary and the LDN-wave, is sufficient to produce significant inundation. However, the Katakolo Pass and the valley of Kato Samiko lie below sea level in DEM₃₀, which makes it more difficult to estimate the local sedimentary and geomorphological impact of the waves.

Finally, the DEM_{rec}-Scenarios are less appropriate with regard to tsunami field traces. Here, larger wave amplitudes, especially of the solitary waves, are needed to cause inundation of

all three sites. Tsunami waves with wave amplitudes of this order would be more extreme and therefore unlikely to occur. However, by applying the critical amplitudes of 3 m to 6 m derived from the palaeoscenarios to the recent landscape conditions, a high tsunami risk can be concluded for the northern Gulf of Kyparissia (Chap. 3.3.3; Fig. 3.7). This risk is considerably higher than reported by TSELENTIS *et al.* (2010) and YALÇINER *et al.* (2014) for the same region. In both studies, different tsunami scenarios were calculated on the basis of various possible focal mechanisms in the area. However, the authors did not take account of field data and possible tsunamis generated in further distance to the coast and also refrained from calculating onshore inundation dynamics.

3.5 Conclusions

This study systematically investigated the effects of different boundary conditions (solitary and LDN-waves) and various coastal palaeo-/topographies on the onshore response of tsunamis in a hydrodynamic model for the Gulf of Kyparissia. Both factors are found to strongly influence not only the intensity of flooding, meaning the inundation depth and flow velocity but also the inundation pattern. In all scenarios, LDN-waves have a significantly stronger onshore response compared to solitary waves. This phenomenon is known from theoretical studies as the *N*-wave effect and is ascribed to the steeper wave crest of LDN-waves. Here, for the first time, the *N*-wave effect has been numerically demonstrated in view of a specific study area, for which, in addition, detailed field evidence of multiple tsunami landfall is known from previous studies. The various palaeo-/topographies are found to have an even greater impact on the simulation results than the different hydrodynamic boundary conditions. This is due to the high tectonic activity and the vast coastal lowlands characteristic of the northern Gulf of Kyparissia. Both combined result in significant topographical changes, once vertical crust movements and associated coastline changes are considered in the palaeo-DEMs, which in turn strongly affect the hydrodynamics. This observation is of great general importance, in that tsunami-prone coasts tend to be tectonically active and change their shape over short periods of time.

The marked differences between the six tsunami scenarios presented in this chapter emphasise the need to consider various waveforms and to reconstruct former landscape conditions in numerical simulations, especially when compared with palaeotsunami field traces. Furthermore, numerical models should not be limited to the marine environment; they also need to involve onshore hydrodynamics, so that a direct comparison to field data can be made. As demonstrated here, this can be an appropriate approach to significantly narrow the gap between numerical simulation results and tsunami hydrodynamics derived from field observations. Once a plausible tsunami scenario is found in accordance with tsunami field evidence, it can be applied to the recent landscape conditions to derive a realistic modern risk assessment.

In the Gulf of Kyparissia, the closest accordance between simulation results and field data is found for the DEM₁₃-Scenarios considering a 13 m tectonic throw along the Epitalio fault zone and adjacent fault zones. With a view to the recent topography, this means that tsunami inundation risks have decreased with time in the uplifted area to the east of the faults. In contrast, simulations reveal a high tsunami risk for the coastal flats, currently found westwards to the fault zones in the northern gulf. This risk is far higher than the risk reported in other studies based on conventional modelling and without calibration using field data.

Despite the close agreement between the numerical model and published tsunami field traces in the Gulf of Kyparissia, future simulations are in need of further adjustment. Although the dominating topographical changes of the study area have been considered, the detected tsunami sediments and overlying layers must be removed from the surface in the palaeo-DEMs

to describe the contemporary relief more realistically in following studies. This, however, requires an even higher density of tsunami field data. Once the palaeotopography is reconstructed more accurately, future simulations should also involve morphodynamic processes (cf. Chap. 5) in order to quantify the actual sedimentary and geomorphological impact of tsunami inundation and, through this, to further bring into line numerical and field findings.

4 Reconstruction of tsunami inundation dynamics in the Gulf of Kyparissia considering palaeogeographical changes and field traces in a numerical model¹

Abstract The west coast of Greece is exposed to high tsunami risks. This is reflected by a large number of historical accounts and geoscientific field traces found all along the coast. Using a hydrodynamic numerical model, in this study various tsunami scenarios are computed and compared with sedimentary and geomorphological field data discovered in palaeotsunami studies in three key locations in the northern Gulf of Kyparissia (western Peloponnese): the Katakolo Pass, the Epitalio Pass and the valley of Kato Samiko. The numerical scenarios take account of three different digital elevation models (DEM), one reflecting the recent topography of the study area and the other two based on possible palaeotopographies derived from neotectonic field data and literature. Moreover, different hydrodynamic boundary conditions are applied considering the solitary and the *N*-wave theory.

Generally, field findings compare favourably with scenarios based on the first palaeo-DEM, which considers a 13 m tectonic throw along the major fault zones of the study area. In these scenarios, 3 m to 6 m solitary or leading-depression *N*-waves cause significant inundation of the three key sites and are assumed to have a sedimentary response which is in accordance with the field traces. However, there is no particular scenario or individual wave that can fully explain tsunami field traces for all key locations at once. This particularly may be due to the fact that (i) tsunami deposits found at the key sites relate to different tsunami impacts with differing inundation dynamics, as is also implied by the field data itself, (ii) that the palaeo-DEMs do not reflect the exact topography at the time of tsunami inundation and/or (iii) that the numerical model is in need of further adjustment and modification, for instance in view of the hydrodynamic boundary conditions. Nevertheless, the current study demonstrates that local inundation dynamics in the northern Gulf of Kyparissia can precisely be derived from palaeotsunami field traces and emphasises the high reliability of the hydrodynamic tsunami model employed.

4.1 Introduction

Worldwide, sedimentary and geomorphological field traces have been identified giving account of recent, historical and prehistorical tsunami events. In view of recent events, field data is generally found well preserved in high density, allowing an accurate reconstruction of the onshore tsunami response, particularly when combined with hydrodynamic numerical simulations (e.g. BOURGEOIS *et al.* 1999; LYNETT *et al.* 2003; LE FRIANT *et al.* 2008; BORRERO *et al.* 2009; WIJETUNGE 2010; FRITZ *et al.* 2011; SATAKE *et al.* 2013; GOPINATH *et al.* 2014). This is a challenge far more demanding in the case of older events. Due to the general low preservation level of tsunami deposits and other field traces (e.g. WHEATCROFT 1990; WHEATCROFT & DRAKE 2003; SZCZUCIŃSKI 2012; SPISKE *et al.* 2013) and the lack of essential information

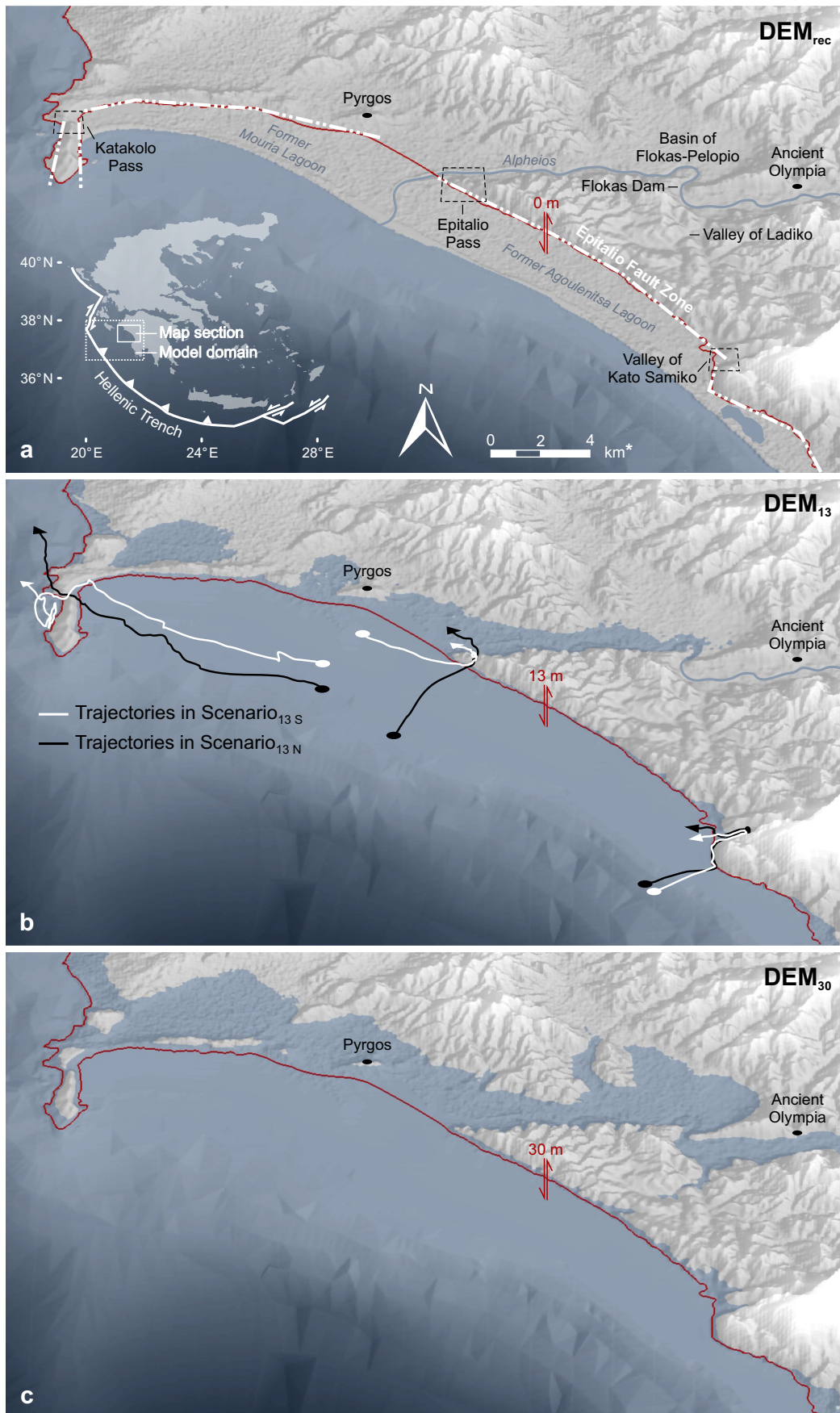
¹This chapter is based on the paper by RÖBKE *et al.* (2015).

on the trigger mechanisms and wave characteristics, pre-/historical tsunami inundation dynamics—in general—can only be derived locally but not covering the complete study area. Here, hydrodynamic numerical models have proven as an appropriate tool as they build the bridge between scattered field findings and allow to reconstruct inundation in the bigger picture (e.g. MINOURA *et al.* 2001; BONDEVİK *et al.* 2005a; MARTIN *et al.* 2008; MACINNES *et al.* 2010). Moreover, essential information about the trigger mechanism can be obtained from numerical tsunami simulations.

With coasts being one of the most dynamic landscapes on earth, they can change their shape over short periods of time (e.g. BIRD 2005; SCOTT 2005; THOM & COWELL 2005). This is particularly true, when the area is under tectonic stress—a feature observed regularly in tsunami-prone areas (cf. NOAA 2015a). Consequently, the interpretation of palaeotsunami field traces and the simulation of those events must take account of contemporary landscape conditions. For example, HILL *et al.* (2014) reconstructed the palaeobathymetry of the Norwegian Sea and the North Sea in their DEM when simulating the Storegga Slide tsunami (almost 8000 years ago). However, simulation results hardly allow a direct comparison with associated tsunami deposits, since no onshore hydrodynamics were calculated, which needs the inundation extent and run-up heights to be estimated. SHISHIKURA *et al.* (2011), SAWAI *et al.* (2012), SUGAWARA *et al.* (2013) and NAMEGAYA & SATAKE (2014) took this approach one step further. Focussing on the seismogenic 869 AD Jōgan tsunami in north-east Japan in the Sendai Bay, the authors considered the former coastline and reconstructed the contemporary topography by subtracting tsunami related deposits and overlying sediments from the present ground in their DEM. In this way, and by testing numerous possible fault-plane solutions, they reproduced a plausible tsunami scenario, which demonstrated that the Jōgan tsunami was the predecessor of the 2011 Tōhoku tsunami.

The current study aims to reconstruct field-implied palaeotsunami inundation dynamics in a hydrodynamic numerical model of the northern Gulf of Kyparissia, a marine inlet of the eastern Ionian Sea in western Greece (Figs. 1.1, 4.1a). Closely related to the subduction of the African Plate along the Hellenic Trench less than a 100 km off the mainland, the west coast of Greece and the eastern Ionian Sea are exposed to high tectonic stress and seismic activity (CGMW 2004; STIROS 2009; SHAW & JACKSON 2010). This makes the area the most tsunamigenic in the entire Mediterranean, as demonstrated by numerous tsunami and earthquake catalogues (Chap. 1.2). Besides seismicity, tsunamis in the Ionian Sea and Gulf of Kyparissia may also arise from submarine landslides or nearby volcanic activity (e.g. by Mount Etna). For a detailed treatise on possible tsunami trigger mechanisms in the Gulf of Kyparissia and its environs see RÖBKE *et al.* (2013, pp. 74–77).

Figure 4.1: 3-D view of digital elevation model DEM_{rec} representing the recent off- and onshore topography of the northern Gulf of Kyparissia (a) and two palaeo-DEMs, which consider a vertical throw along the Epitalio fault zone and adjacent fault zones (white dashed-dotted lines in (a)) of 13 m, DEM₁₃ (b), and 30 m, DEM₃₀ (c). In both palaeo-DEMs, the corresponding throw was levelled out along the red line, which derives from the present boundary between the coastal lowlands and adjacent Neogene hills (about 23 to 2.5 million years before present) associated with the approximate course of the faults (IGME 1982a,b; LEKKAS *et al.* 1992; PAPANIKOLAOU *et al.* 2007, pp. 26, 27). In all DEMs, the present sea level was assumed indicated by the blue surface. The simplified regional tectonic regime and the numerical model domain is displayed in the overview map of Greece in (a). Dashed boxes in (a) show the location of the three key sites. White and black lines in (b) illustrate the trajectories of a hypothetical water particle during the landfall of the 6 m waves for given periods of time in Scenario_{13S} and Scenario_{13N} respectively (Chap. 4.4). *Due to the 3-D view, the scale bar gives the true distance only in the centre of the figure.



Several sedimentary and geomorphological field traces have been found along the entire west coast of Greece indicating tsunami landfall in historical and prehistorical times (Chap. 1.2). This study takes account of field findings of three key locations in the northern Gulf of Kyparissia: the Katakolo Pass, roughly 20 m a.s.l., linking the Katakolo Promontory with the mainland at the northernmost tip of the gulf (Figs. 4.1a, 4.2, 4.3; VÖTT *et al.* 2010a, pp. 32–39; VÖTT *et al.* 2011b), the Epitalio Pass (almost 30 m a.s.l.), leading over from the former Agoulénitsa Lagoon to the entrance of the Alpheios River valley (Figs. 4.1a, 4.3, 4.4; VÖTT *et al.* 2015), and the small valley of Kato Samiko (circa 11 m a.s.l.) to the north of Mount Lapithas (Figs. 4.1a, 4.3, 4.5; WILLERSHÄUSER *et al.* 2012, 2015a). At all three locations, numerous geoelectrical measurements and vibracores were carried out during the last years, which, together with the geomorphological setting, indicate several tsunami inundation events in the last 8000 years. Besides these findings, vibracore stratigraphies further imply fundamental changes in the local topography related to high tectonic and fluvial activity.

Based on the methodology of Chap. 3, this investigation intends to (i) reconstruct the contemporary topographies associated with palaeotsunami inundation and create corresponding DEMs covering the wider study area, (ii) set up a hydrodynamic numerical tsunami model considering different palaeo-/DEMs and various hydrodynamic boundary conditions, (iii) create plausible tsunami wave scenarios in view of the three key locations and the entire study area, (iv) compare the different scenarios with sedimentary and geomorphological tsunami field traces with (v) the aim of reconstructing small-scale palaeotsunami inundation dynamics.

4.2 Methodology

4.2.1 Digital elevation models

The DEMs employed in the simulation calculations of this study were compiled from various sources, including bathymetric data of several topographical and nautical charts on the scale of 1:50 000 to 1:500 000 (SCGC USSR 1983; 1986; 1988; Imray Laurie Norie & Wilson Ltd 2008; Eagle Ray Publications 2010a; UKHO 2010) as well as 90 m SRTM elevation data version 4 (JARVIS *et al.* 2008) and 30 m Google Earth elevation data (Google Inc. 2013) for the onshore topography. Using ArcGIS Desktop 10.2 by Esri, bathymetric and topographical data were edited and assembled to the DEM_{rec}, representing the recent off- and onshore topography of the Gulf of Kyparissia and its environs (Fig. 4.1a). The model covers the domain from 36.67° N to 38.00° N and from 20.11° E to 22.00° E with a resolution of up to 30 m.

As a major goal of this study, two further DEMs were created that represent the possible palaeotopography during prehistorical tsunami impacts detected by VÖTT *et al.* (2010a, pp. 32–39), VÖTT *et al.* (2011b, 2015) and WILLERSHÄUSER *et al.* (2012, 2015a) (Chap. 4.3). Here, particularly vertical crust movements along the major faults in the northern Gulf of Kyparissia—the Epitalio fault zone (EFZ) and adjacent fault zones (see white dashed-dotted lines in Fig. 4.1a)—had to be considered. The analysis of relative sea level indicators in several vibracores from either side of the faults by VÖTT *et al.* (2015) revealed a minimum throw (vertical component of the dip separation) of 13 m for the last 6000 years, which was the basis of the first created palaeo-DEM, DEM₁₃ (Fig. 4.1b). DEM₁₃ shows the on- and offshore topography at around 4000 BC or later in the case the crust movements were discontinuous. A second palaeo-DEM took account of the findings by PAPANIKOLAOU *et al.* (2007, p. 24), who suggest a Holocene (covering the last 10 000 years) throw of almost 60 m along the EFZ and adjacent fault zones in the northern Gulf of Kyparissia. According to this and assuming continuous crust movement, a 30 m-throw was deduced as the maximum vertical offset in the last 6000 years in

palaeo-DEM₃₀ (Fig. 4.1c).

VÖTT *et al.* (2015) report marked differences in relative sea levels of at least 13 m for the past 6000 years in the wider study area. Thus, it cannot be concluded, which side of the faults was subsided or uplifted to what extent or to deduce a reliable contemporary sea level. Therefore, the continental, eastern part of the faults was lowered by 13 m (DEM₁₃) and 30 m (DEM₃₀) and the present sea level was assumed in all DEMs. For further methodological details concerning the created DEMs see Chap. 3.2.1.

4.2.2 Hydrodynamic numerical model

Tsunami simulations were calculated with a two-dimensional, depth-averaged calculation approach using the hydrodynamic program module Delft3D-FLOW of the Delft3D suite (version 4.01.00.rc.12) by Deltares. The program has been successfully employed several times in regard to hydrodynamic off- and onshore tsunami simulations (Chaps. 3, 5; GELFENBAUM *et al.* 2007; APOTSOS *et al.* 2009, 2011a,b,c,d, 2012; RÖBKE *et al.* 2012, 2013; CHACÓN-BARRANTES *et al.* 2013; PERÍÁÑEZ & ABRIL 2014a,b).

The current study solely considers the hydrodynamics of tsunami waves excluding trigger mechanisms as these are generally unknown in respect to the palaeotsunami events indicated by field traces. Consequently, various waveforms and wave amplitudes were applied, intended to find plausible tsunami scenarios. The boundary conditions of the tsunami waves were determined according to the solitary and the *N*-wave theory. The solitary wave, which has widely been used in numerical and experimental tsunami simulations (Chap. 2.5), was calculated pursuant to the formulation by DEAN & DALRYMPLE (1991, pp. 314–317) (see Eq. 2.13; cf. LEHFELDT *et al.* 2007, pp. 112–114; RÖBKE *et al.* 2012, p. 562; RÖBKE *et al.* 2013, pp. 77–79). Furthermore, isosceles leading-depression (LDN) *N*-waves were chosen, which are the typical waveform of nearshore generated tsunamis (Chap. 2.5), as may be the case in the Gulf of Kyparissia due to the nearby seismogenic Hellenic Trench. The boundary conditions of the LDN-waves were calculated according to the formulation by TADEPALLI & SYNOLAKIS (1994, p. 105) (see Eq. 2.14; cf. CHACÓN-BARRANTES *et al.* 2013, pp. 15, 16).

Tripartite tsunami wave trains with increasing wave amplitudes (cf. RÖBKE *et al.* 2013, pp. 77–79) and both waveforms were created (cf. Fig. 3.3). The wavelength of the separate waves ranged from 46 min to 160 min. All wave trains started at the western boundary of the DEMs (−2925 m deep on average) because waves from this direction are associated with the strongest onshore response in the northern Gulf of Kyparissia (RÖBKE *et al.* 2012, 2013).

Based on DEM₁₃, various wave amplitudes for both the solitary and LDN-wave train were simulated in order to generate inundation of the three key locations. In the case of the solitary wave theory, the intended onshore hydrodynamics were found for a set of 1 m, 3 m and 6 m waves (cf. Fig. 3.3). For comparison of both waveforms, the same wave amplitudes were then applied to the LDN-wave train. Besides DEM₁₃, both wave trains were also simulated on the basis

Table 4.1: Outline of the various tsunami scenarios presented in this study. Two different wave types and three different DEM-configurations were combined, resulting in a total of six scenarios.

	Solitary wave train (1 m, 3 m, 6 m)	LDN-wave train (1 m, 3 m, 6 m)
DEM ₁₃	Scenario _{13S}	Scenario _{13N}
DEM ₃₀	Scenario _{30S}	Scenario _{30N}
DEM _{rec}	Scenario _{recS}	Scenario _{recN}

of DEM₃₀ and DEM_{rec}, resulting in a total of six tsunami scenarios presented in this chapter (Tab. 4.1). For further methodological details concerning the numerical model and boundary conditions see Chap. 3.2.2.

4.3 Tsunami field traces

4.3.1 Katakolo Pass

The Katakolo Pass links the mainland of the northern gulf coast with the Katakolo Promontory, which is an almost 4 km long horst structure with Cape Katakolo at its southern tip (LEKKAS *et al.* 2000, pp. 22–25; PAPANIKOLAOU *et al.* 2007, p. 17; Figs. 4.1a, 4.2). During palaeogeographical studies, several vibracores were taken along a W-E running transect across the pass (Figs. 4.2, 4.3), altogether suggesting six distinct inundation events, which are presumably associated with tsunami impacts (VÖTT *et al.* 2011b). Based on radiocarbon ages and diagnostic ceramics, the first event occurred at the beginning of the 6th millennium BC, the second around 4300 ± 200 cal BC and the third in Byzantine (circa 400 AD to 1500 AD) or post-Byzantine times. Sedimentary traces of another three younger events have not yet been dated (VÖTT *et al.* 2011b). Tsunami inundation is indicated by correlating stratigraphic units in the vibracore transect (Fig. 4.3), consisting of coarse-grained allochthonous sediment, mostly sand and gravel of marine origin. The deposits are further characterised by basal erosional contacts, fining-upward sequences and rip-up clasts — features typical of tsunami event layers (cf. Chap. 2.4; e.g. MORTON *et al.* 2007). Besides vibracore data, calcified sandy tsunami deposits in the form of beachrock

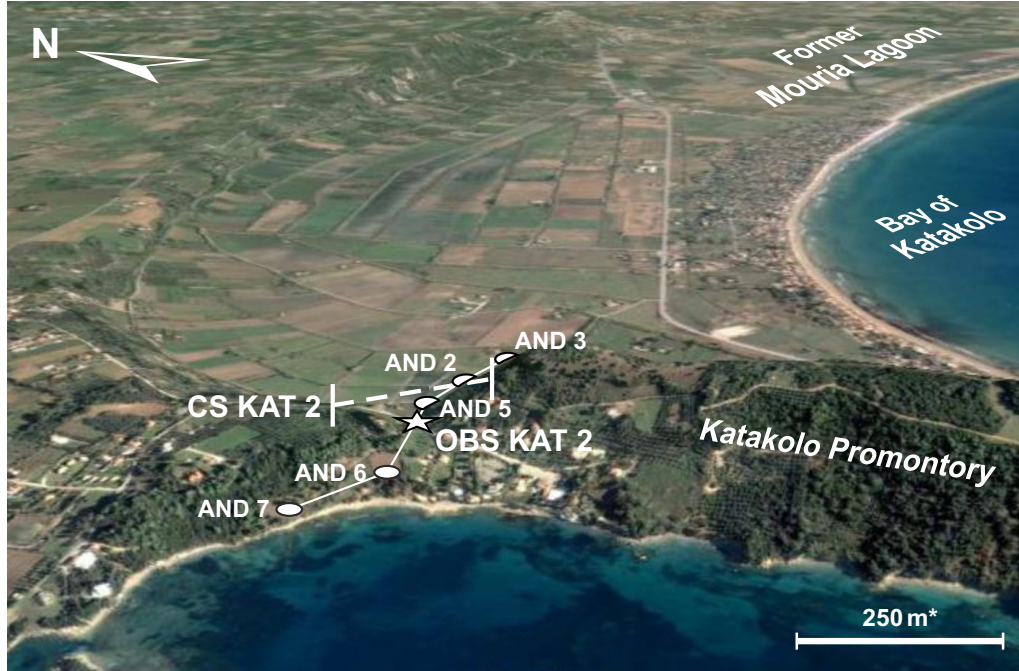


Figure 4.2: Bird’s-eye view of the Katakolo Pass linking the Katakolo Promontory with the mainland of the northern gulf coast (Google Inc. 2013). White dots mark the position of the AND-vibracores (VÖTT *et al.* 2011b). The white star indicates observation point (OBS) KAT 2, the white bars confine cross section (CS) KAT 2, for which the water depth, depth-averaged flow velocity and instantaneous discharge were computed (Fig. 4.7; Chap. 4.4.1). For the location of the Katakolo Pass see Fig. 4.1a. *Due to the 3-D view, the scale bar gives the true distance only at its exact position.

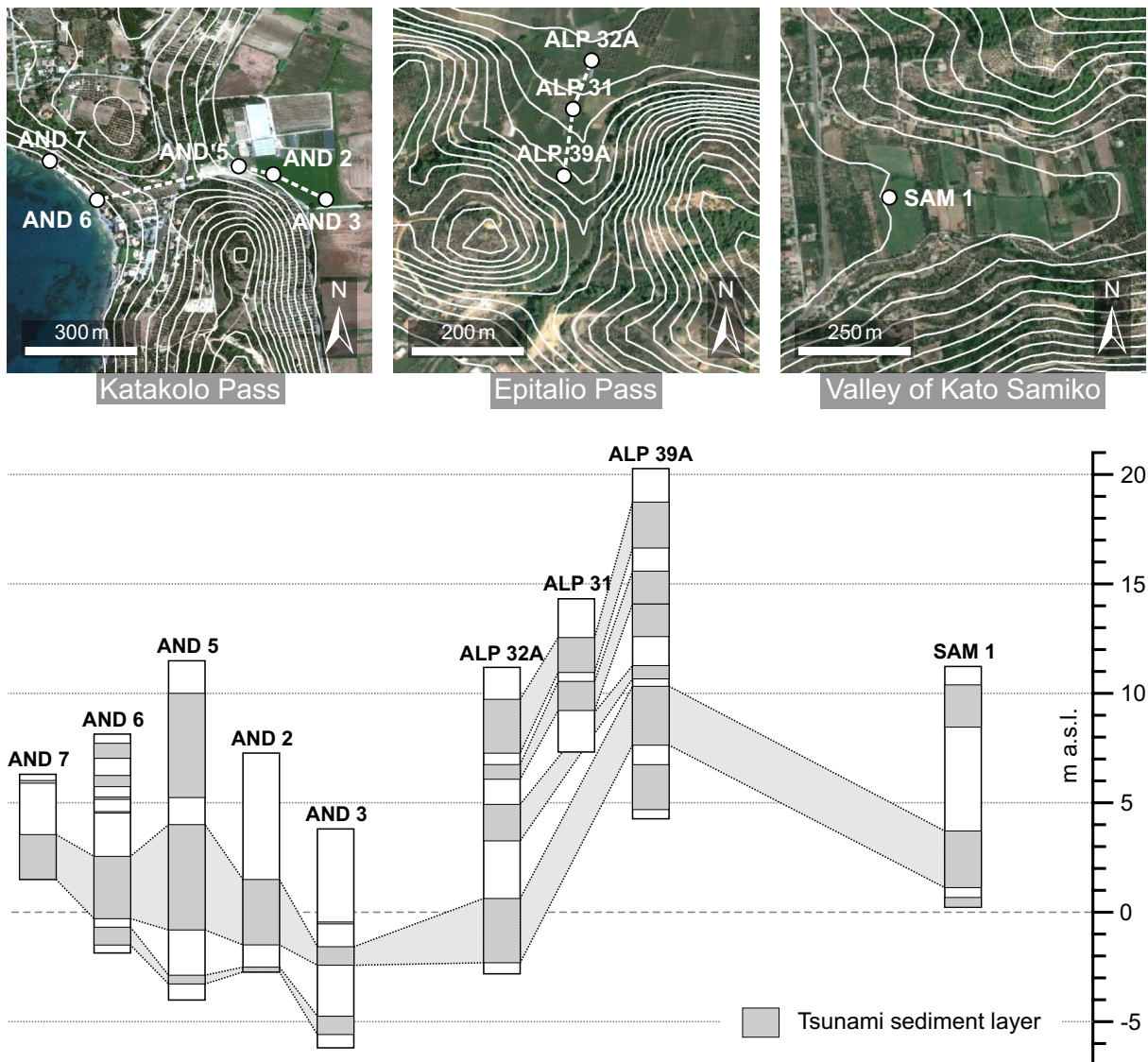


Figure 4.3: Simplified vibracore profiles showing tsunami deposits (grey segments) and correlations to each other as reported by VÖTT *et al.* (2011b) for the Katakolo Pass, by VÖTT *et al.* (2015) for the Epitalio Pass and by WILLERSHÄUSER *et al.* (2012, 2015a) for the valley of Kato Samiko. In all three key locations, tsunami layers typically show coarse-grained allochthonous material of marine origin, basal erosional contacts, fining-upward sequences, rip-up clasts and further characteristic features of tsunamites (cf. Chap.2.4). For the location of the three key sites see Fig. 4.1a.

were found on the western side of the Katakolo Pass (VÖTT *et al.* 2010a, pp. 32–39; VÖTT *et al.* 2011b, pp. 263–267). The interpretation of these deposits being of tsunamigenic nature but not storm-related is discussed in detail by VÖTT *et al.* (2011b, pp. 276, 277) and RÖBKE *et al.* (2013, p. 97).

4.3.2 Epitalio Pass

The Epitalio Pass leads over from the Alpheios River valley to the village of Epitalio and the former Agoulénitsa Lagoon (Figs. 4.1a, 4.4). The Epitalio Valley, located to the north of the pass, is characterised by a huge alluvial fan, which was palaeogeographically investigated by

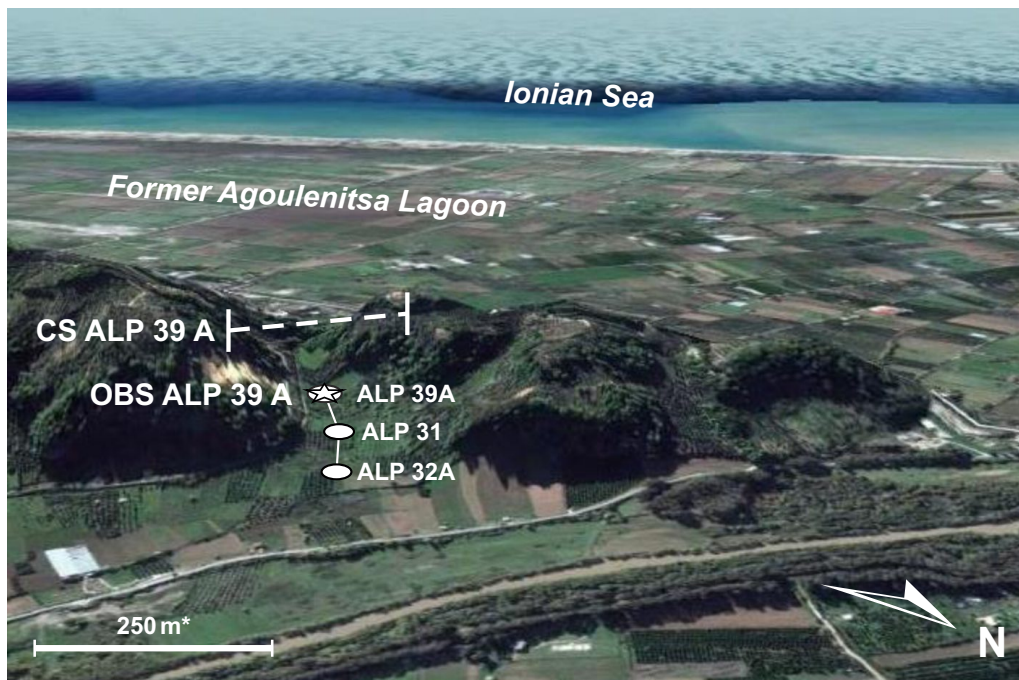


Figure 4.4: Bird's-eye view of the Epitalio Pass leading over from the lower Alpheios River valley to the former Agoulenitsa Lagoon (Google Inc. 2013). White dots mark the position of the ALP-vibracores (VÖTT *et al.* 2015). The white star indicates observation point (OBS) ALP 39A, the white bars confine cross section (CS) ALP 39A, for which the water depth, depth-averaged flow velocity and instantaneous discharge were computed (Fig. 4.10; Chap. 4.4.2). For the location of the Epitalio Pass see Fig. 4.1a. *Due to the 3-D view, the scale bar gives the true distance only at its exact position.

VÖTT *et al.* (2015). Electrical resistivity measurements and vibracores showed the so-called Epitalio Fan to be composed of limnic to fluvial and colluvial deposits with several intersecting coarse-grained event layers characterised by similar tsunamigenic features as observed around the Katakolo Pass (Fig. 4.3; Chap. 4.3.1). Microfaunal analyses of the high-energy deposits document a lagoonal to deep water origin of the transporting waters and therefore substantiate a tsunamigenic origin of the sediments. Furthermore, microfossil signals from the surrounding bedrock were identified in the event layers, indicating strong erosion during inundation. All in all, five inundation events were distinguished in the local stratigraphy, the first event dating back to 5300–5200 cal BC, the second to 4350–4250 cal BC, the third to the 3rd millennium BC and the two most recent events to younger times, not dated yet.

4.3.3 Valley of Kato Samiko

The valley of Kato Samiko is located to the south-east of the former Agoulenitsa Lagoon at the foot of Mount Lapithas (Figs. 4.1a, 4.5), a Cretaceous limestone ridge (IGME 1982b). During field work, vibracore SAM 1 (circa 11 m a.s.l.) and several geoelectrical tomograms were obtained in the valley (Fig. 4.3). The vibracore shows three strata of high-energy deposits that are characterised by typical tsunamigenic features as already described above (RÖBKE *et al.* 2012, pp. 563, 564; RÖBKE *et al.* 2013, pp. 93, 94; WILLERSHÄUSER *et al.* 2012, pp. 29–32; WILLERSHÄUSER *et al.* 2015a). The oldest event occurred prior to 4451–4369 cal BC, the second oldest immediately after this period. The youngest high-energy stratum in SAM 1, found as part of a channel structure, relates to the most distinct inundation event in the valley. Based on

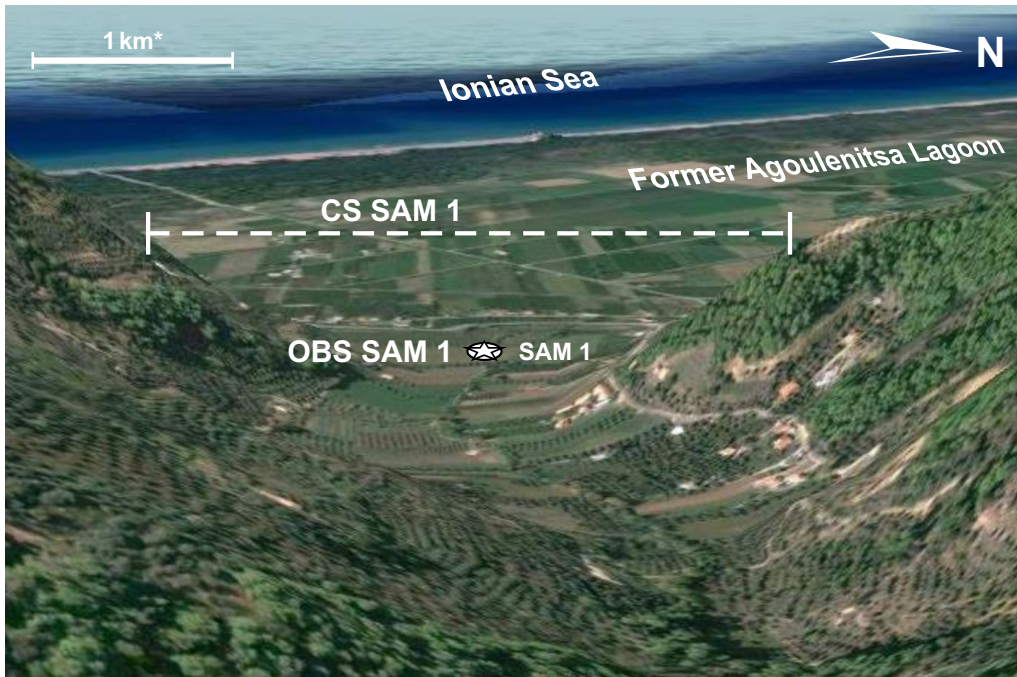


Figure 4.5: Bird's-eye view of the valley of Kato Samiko at the foot of Mount Lapithas to the south-east of the former Agoulenitsa Lagoon (Google Inc. 2013). The white dot marks the position of vibracore SAM 1 (WILLERSHÄUSER *et al.* 2012, 2015a). The white star indicates observation point (OBS) SAM 1, the white bars confine cross section (CS) SAM 1, for which the water depth, depth-averaged flow velocity and instantaneous discharge were computed (Fig. 4.13; Chap. 4.4.3). For the location of the valley of Kato Samiko see Fig. 4.1a. *Due to the 3-D view, the scale bar gives the true distance only at its exact position.

detailed multi-proxy analyses of vibracore and geophysical data, WILLERSHÄUSER *et al.* (2012, pp. 31–33) and WILLERSHÄUSER *et al.* (2015a) associate this channel structure with the backflow of tsunami waters in the valley of Kato Samiko. The backflow channel starts to the east of coring site SAM 1 and continuously widens towards the sea. Judging by radiocarbon ages and diagnostic ceramics, the associated tsunami event took place between the 4th century BC and the 3rd century AD (WILLERSHÄUSER *et al.* 2015a).

4.4 Results and discussion

4.4.1 Katakolo Pass

Simulation results

Regarding DEM₁₃, with an approximate pass elevation of 7 m a.s.l., and the solitary wave train (Scenario_{13S}), the Katakolo Pass receives flooding only by the third, 6 m wave. Inundation is not initiated at the western side of the pass, but instead the wave is strongly refracted and diffracted around Cape Katakolo and therefore approaches the pass from a south-eastern direction (Fig. 4.6a). Subsequently, the wave is funnelled in the Bay of Katakolo and the water level continuously rises on the eastern side of the pass (Fig. 4.6b). Finally, tsunami waters overflow the pass in an east-west direction, increased by reflection waves that originate from the hilly coastlands along the EFZ in the south-east (Fig. 4.6c). Small hydraulic lee eddies emerge to the west of the pass during inundation. Figure 4.7 (left column) shows a maximum inundation

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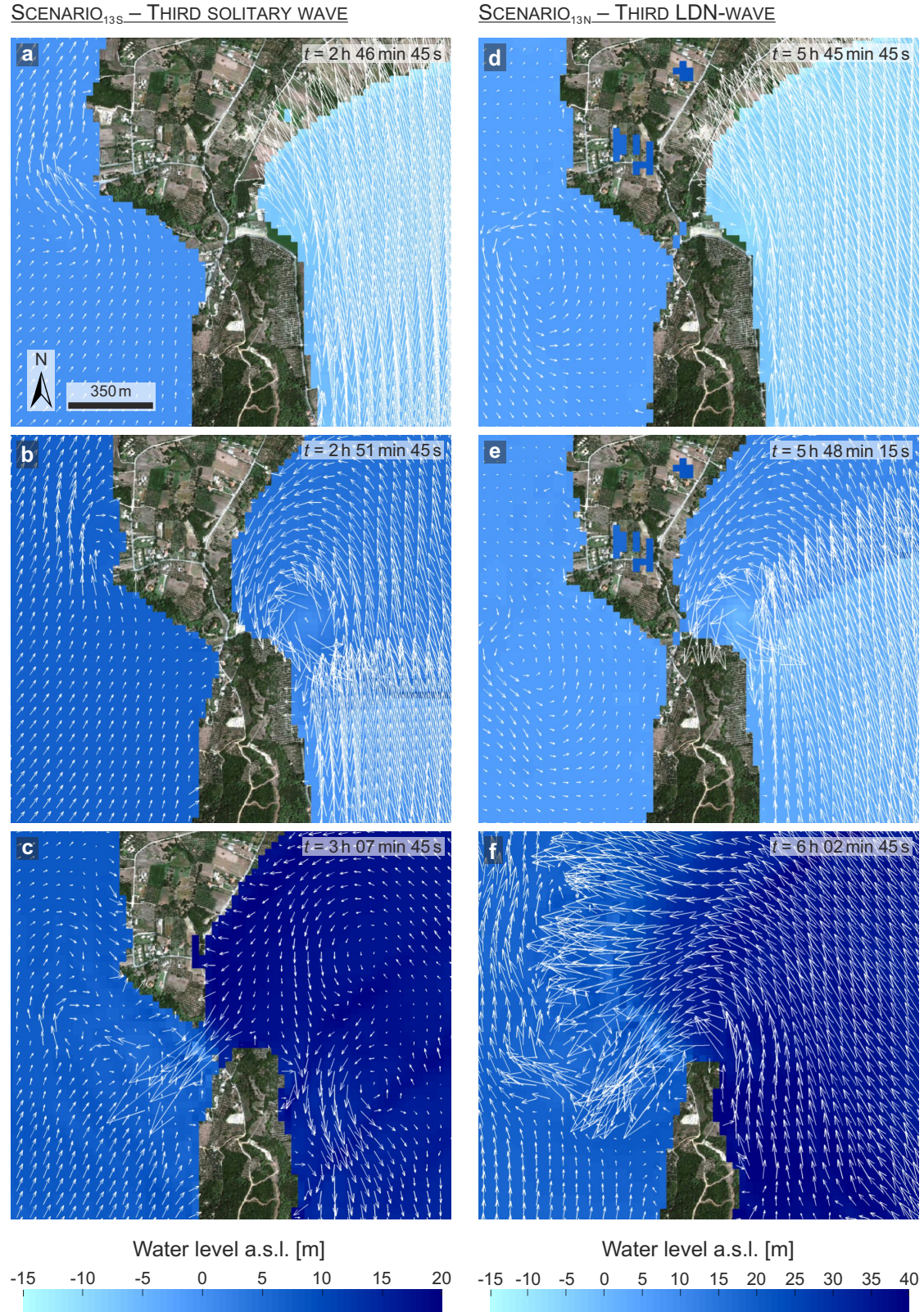


Figure 4.6: Snapshots documenting stepwise the inundation of the Katakolo Pass associated with the third, 6 m wave of the solitary (Scenario_{13S}, left column) and LDN-tsunami-wave train (Scenario_{13N}, right column) based on DEM₁₃ (Fig. 4.1b). White vectors indicate the depth-averaged flow direction and velocity. See Fig. 4.1a and Fig. 4.2 for the location and bird's-eye view of the site respectively.

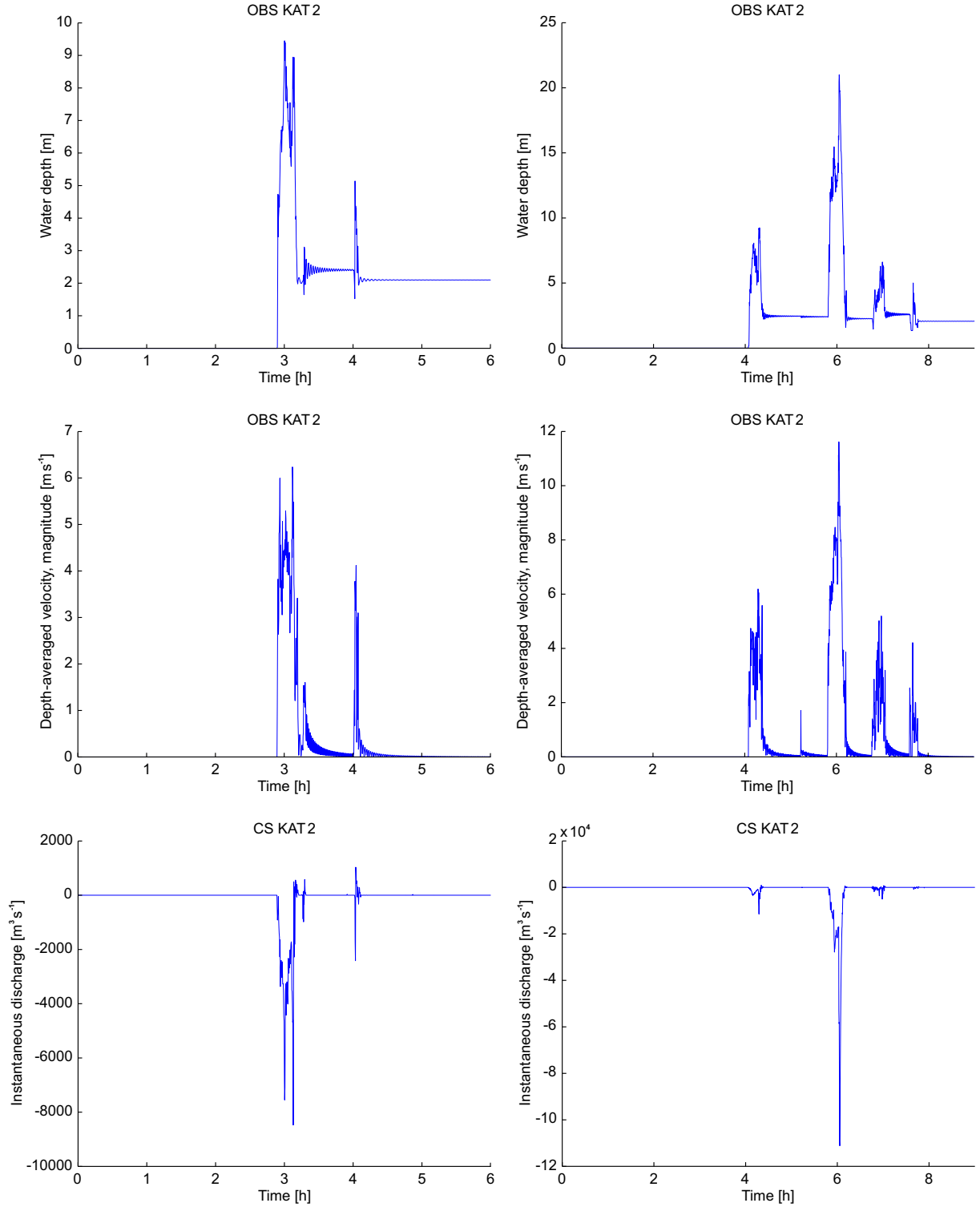


Figure 4.7: Diagrams of the water depth, depth-averaged flow velocity and instantaneous discharge as a function of time for the Katakolo Pass in Scenario_{13S} (left column) and Scenario_{13N} (right column). Water depth and flow velocity are recorded for the 30 m grid cell at OBS KAT 2. The discharge relates to CS KAT 2. Both, OBS and CS KAT 2 are illustrated in Fig. 4.2.

depth of 9.5 m and a maximum depth-averaged flow velocity of circa 6.3 m s^{-1} associated with the 6 m solitary wave at observation point (OBS) KAT 2, which is located at the top of the pass. This results in a maximum instantaneous discharge through the pass, meaning through cross section (CS) KAT 2, of about $8200 \text{ m}^3 \text{ s}^{-1}$ (Fig. 4.7, bottom left).

In Scenario_{13N}, already the second LDN-wave overflows the Katakolo Pass—comparable to the third solitary wave (cf. Figs. 4.6a–c). Inundation by the third LDN-wave is even stronger (here and in the following in view of the inundation depth and flow velocity). Although the hydrodynamics are initially identical to the 6 m solitary wave (cf. Figs. 4.6a,b,d,e), the 6 m LDN-wave finally strikes the complete area to the north of the Katakolo Pass, which makes the inundation far more turbulent and more intense (Fig. 4.6f). This is also reflected by highly increased values of the inundation parameters shown in Fig. 4.7 (right column).

In the initial situation of the DEM₃₀-Scenarios, the Katakolo Pass is covered by water minimum 10 m deep. Accordingly, strong currents through the pass in both directions are observed with the passage of the first solitary and LDN-wave (Fig. 4.8). Again, small lee eddies emerge on the corresponding side of the pass. There is barely a difference between both scenarios, except for the inundation of another bottleneck further north by the LDN-wave (cf. Figs. 4.8b,d). Flooding caused by the second and third solitary and LDN-waves is on an unprecedented scale, affecting the complete area to the north of the Katakolo Pass. Inundation dynamics are highly turbulent and undifferentiated, comparable to the third LDN-wave in Scenario_{13N} (cf. Fig. 4.6f).

In the DEM_{rec}-Scenarios (pass elevation circa 20 m a.s.l.) none of the waves inundates the pass, except the third LDN-wave, which produces similar hydrodynamics as described for the third solitary wave in Scenario_{13S} (cf. Figs. 4.6a–c).

Comparison with field traces

The numerical simulation results clearly reflect tsunami inundation as implied by the field findings (Chap. 4.3.1). The best agreement with field data is found for the DEM₁₃-Scenarios. Considering the maximum values of the inundation parameters computed for the Katakolo Pass in Scenario_{13S} (Fig. 4.7, left column), a substantial amount of sediment may be transported by the third solitary wave across the pass. This accords with the thick event layers found in the vibracores. The maximum sediment thickness of up to 5 m is recorded in vibracore AND 5 and AND 6 on both flanks of the pass (Fig. 4.3), whereas the more distant vibracores show thinner tsunami event layers. This may be due to the pass being an obstacle to approaching tsunami waters so that transported sediments are partly released on the windward (eastern) side of the pass. However, sediments will dominantly accumulate on the leeward (western) side of the pass, where the flow velocity decreases due to flow divergence. In contrast, the strongest erosion is assumed at the crest of the pass, where funnelling of tsunami waters increases the flow velocity to a maximum. Furthermore, lee eddies associated with the jet-like overflow may produce significant scouring effects on the western side of the pass, as nowadays indicated by corresponding structures in the tsunamigenic beachrock (VÖTT *et al.* 2010a, pp. 32–39; VÖTT *et al.* 2011b, p. 264).

Tsunami deposits in the environs of the Katakolo Pass mainly consist of silt, sand and gravel and may originate from the northern gulf coast several kilometres distant from the pass. This is suggested by Fig. 4.1b, which shows the path (white trajectory) of a hypothetical water particle in the depth-averaged flow from the northern gulf shelf to the western side of the Katakolo Pass associated with the 6 m solitary wave in the simulation time between 2 h 57 min and 3 h 20 min in Scenario_{13S}. The water moves across the shallow shelf, where it picks up sediment from the sea ground. After striking the hilly coastlands, the flow changes its direction to the west and finally streams across the Katakolo Pass, where most of the sediment is assumed to be released

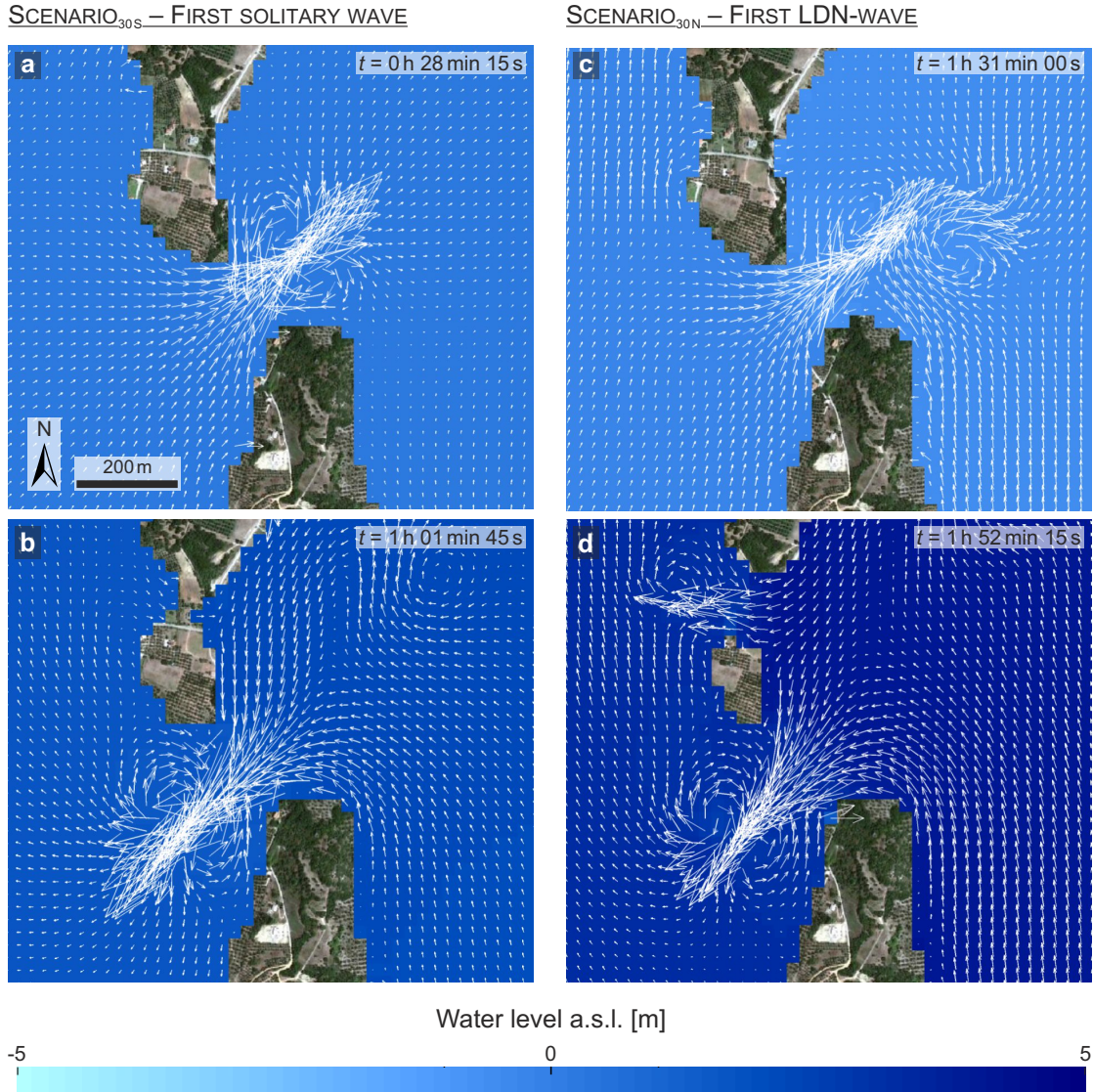


Figure 4.8: Snapshots documenting west-east and east-west orientated strong tsunami currents through the Katakolo Pass with the passage of the first, 1 m wave of the solitary (Scenario_{30S}, left column) and LDN-tsunami-wave train (Scenario_{30N}, right column) based on DEM₃₀ (Fig. 4.1c). White vectors indicate the depth-averaged flow direction and velocity. See Fig. 4.1a and Fig. 4.2 for the location and bird’s-eye view of the site respectively.

on both sides of the pass (see above). Tsunami waters are finally found swirling on the western side of the Katakolo Promontory.

The corresponding grey trajectory in Fig. 4.1b (simulation time 5 h 54 min to 6 h 5 min) indicates a similar possible transport path of sediment for the third LDN-wave in Scenario_{13N}. Nevertheless, this wave is less appropriate for a comparison with the vibracore transect as inundation is not limited to the Katakolo Pass. Instead, the 3 m LDN-wave closer accords with field data, since the hydrodynamics are nearly the same as described for the 6 m solitary wave (cf. Figs. 4.6a–c).

Due to the Katakolo Pass being covered by water in Scenario_{30S} and Scenario_{30N}, here it is more difficult to estimate the actual sedimentary and geomorphological response of tsunami waves on the site. However, thick tsunami sediments may be released on both sides of the pass

from currents driven by the 1 m solitary and LDN-wave due to the same reasons as discussed for the DEM₁₃-Scenarios. Furthermore, scour structures found on the westward side of the pass could be explained by the small lee eddies associated with the currents. Contrary to this, the turbulent and undifferentiated flooding caused by the second and third waves in the DEM₃₀-Scenarios hinders a detailed comparison with field data.

In the DEM_{rec}-Scenarios, also a 6 m LDN-wave may leave sedimentary and geomorphological field traces as found across the Katakolo Pass, this wave being the only one causing inundation of the pass in the case of the recent topography. Inundation dynamics are comparable to those observed for the 6 m solitary wave in Scenario_{13S} and are therefore assumed to have a similar geomorphological impact on the site.

4.4.2 Epitalio Pass

Simulation results

Figures 4.9a–c show the inundation of the Epitalio Pass (circa 17 m a.s.l. in DEM₁₃) caused by the third, 6 m solitary wave in Scenario_{13S} (the first two waves do not affect the pass). The wave approaches from the south-west, strikes the ridge landwards from Epitalio and is refracted and diffracted around the prominent spur of Epitalio into the Alpheios River valley (Fig. 4.9a). At the same time, a large eddy emerges on the leeward side of the Epitalio Pass. With the ridge acting as a barrier against the up-running tsunami wave, the water level continuously rises to the south-west of the pass. Finally, it is a reflection wave from the ridge to the north of the former Mouria Lagoon that approaches from the north-west and causes the pass to get inundated (Figs. 4.9a,b). Due to strong funnelling, tsunami waters reach a depth-averaged flow velocity of about 6.2 m s^{-1} directly behind the pass, but slow down to almost 2.6 m s^{-1} at vibracoring site/OBS ALP 39A, associated with a maximum inundation depth of circa 11.2 m (Fig. 4.10, left column). This results in a maximum instantaneous discharge through CS ALP 39A in the Epitalio Valley of nearly $700 \text{ m}^3 \text{ s}^{-1}$ (Fig. 4.10, bottom left).

Although Scenario_{13S} and Scenario_{13N} show quite similar results with regard to the 6 m solitary and LDN-wave respectively (cf. left and right column in Fig. 4.9), there are some vital differences which have to be mentioned. In Scenario_{13N}, flooding of the Epitalio Pass is directly ascribed to the 6 m wave crest, whereas reflection waves just intensify the inundation dynamics. Due to the stronger onshore response of the LDN-wave, funnel effects across the pass are more intense, resulting in a jet-like flow on the leeward side of the pass (Fig. 4.9d), with a maximum depth-averaged flow velocity of circa 12.5 m s^{-1} (Fig. 4.10, right centre). The jet-like flow persists for about 10 min and then merges with a strong eddy in the Epitalio Valley (Fig. 4.9e). Apart from the flow velocity, also the maximum inundation depth (24 m) and the maximum instantaneous discharge through the valley ($11\,000 \text{ m}^3 \text{ s}^{-1}$; Fig. 4.10, right column) are significantly increased compared to Scenario_{13S}. The same is true for the backflow in the Alpheios River valley (Fig. 4.9g), which shows twice the velocity (18 m s^{-1}) recorded in the solitary wave scenario. In Scenario_{13N}, the Epitalio Pass is also flooded in the inverse direction, illustrated in Fig. 4.9f. This is due to the huge water volume in the Alpheios River valley, which finally makes tsunami waters streaming across the pass from a northern direction.

In the scenarios based on DEM₃₀, the Epitalio Pass just lies above sea level. Here, the run-up of the 3 m waves cause nearly the same inundation pattern and hydrodynamics as computed for the 6 m wave in Scenario_{13N} (cf. Figs. 4.9d–g, 4.11). Thus, a jet-like flow emerges leeward to the Epitalio Pass (Figs. 4.11a,e) and merges with a large eddy in the Epitalio Valley (Figs. 4.11b,f). Subsequently, the pass is overflowed in the inverse direction (Figs. 4.11c,g) and strong backflow currents occur northwards to the Epitalio Fan in the Alpheios River valley (Figs. 4.11d,h). Once

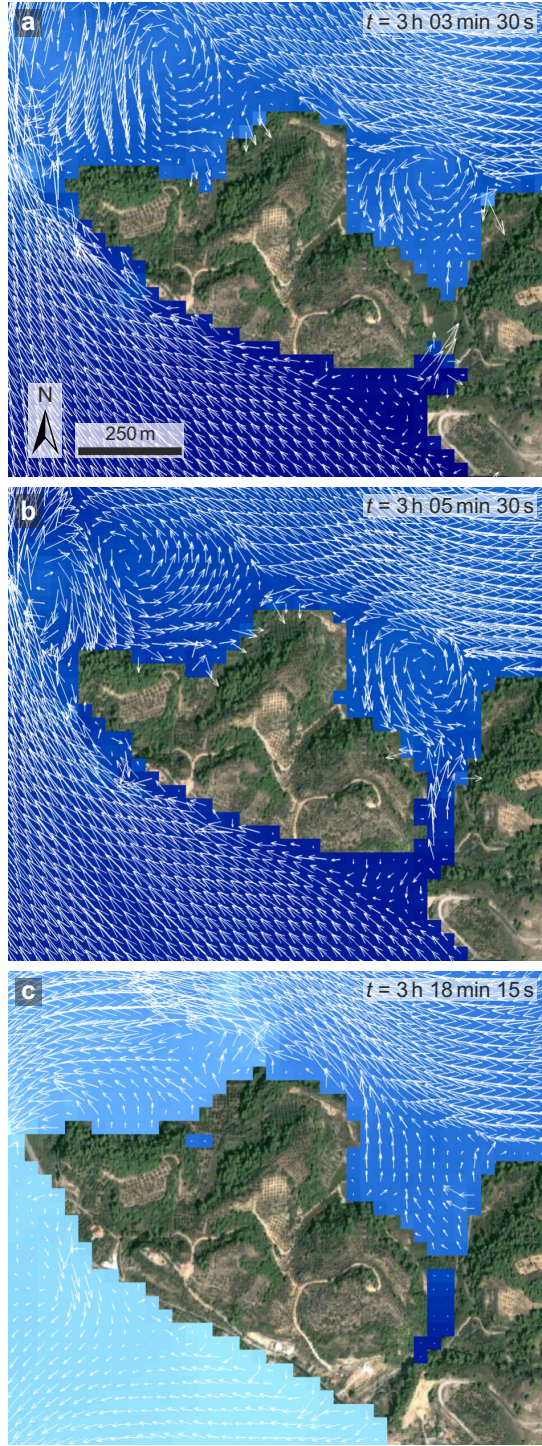
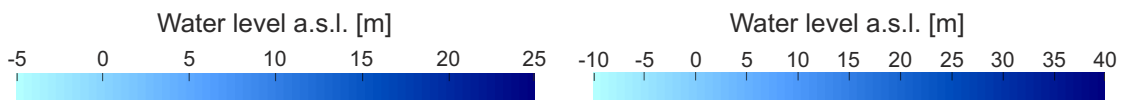
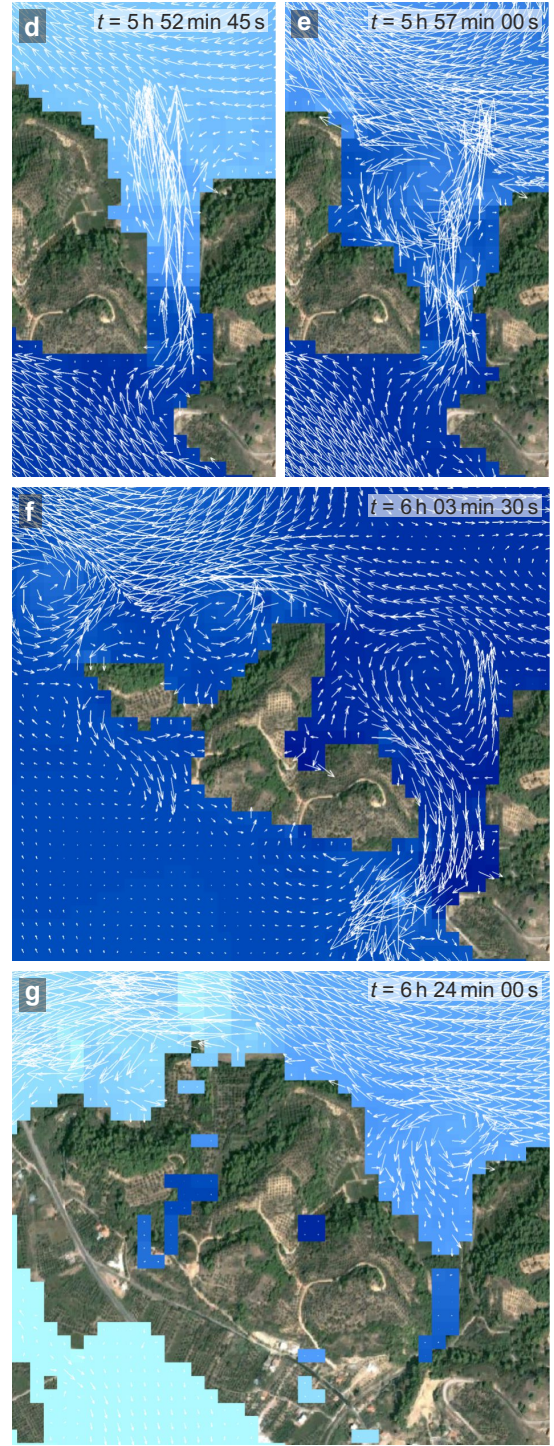
SCENARIO_{13S} — THIRD SOLITARY WAVESCENARIO_{13N} — THIRD LDN-WAVE

Figure 4.9: Snapshots documenting stepwise the inundation of the Epitalio Pass associated with the third, 6 m wave of the solitary (Scenario_{13S}, left column) and LDN-tsunami-wave train (Scenario_{13N}, right column) based on DEM₁₃ (Fig. 4.1b). White vectors indicate the depth-averaged flow direction and velocity. See Fig. 4.1a and Fig. 4.4 for the location and bird's-eye view of the site respectively.

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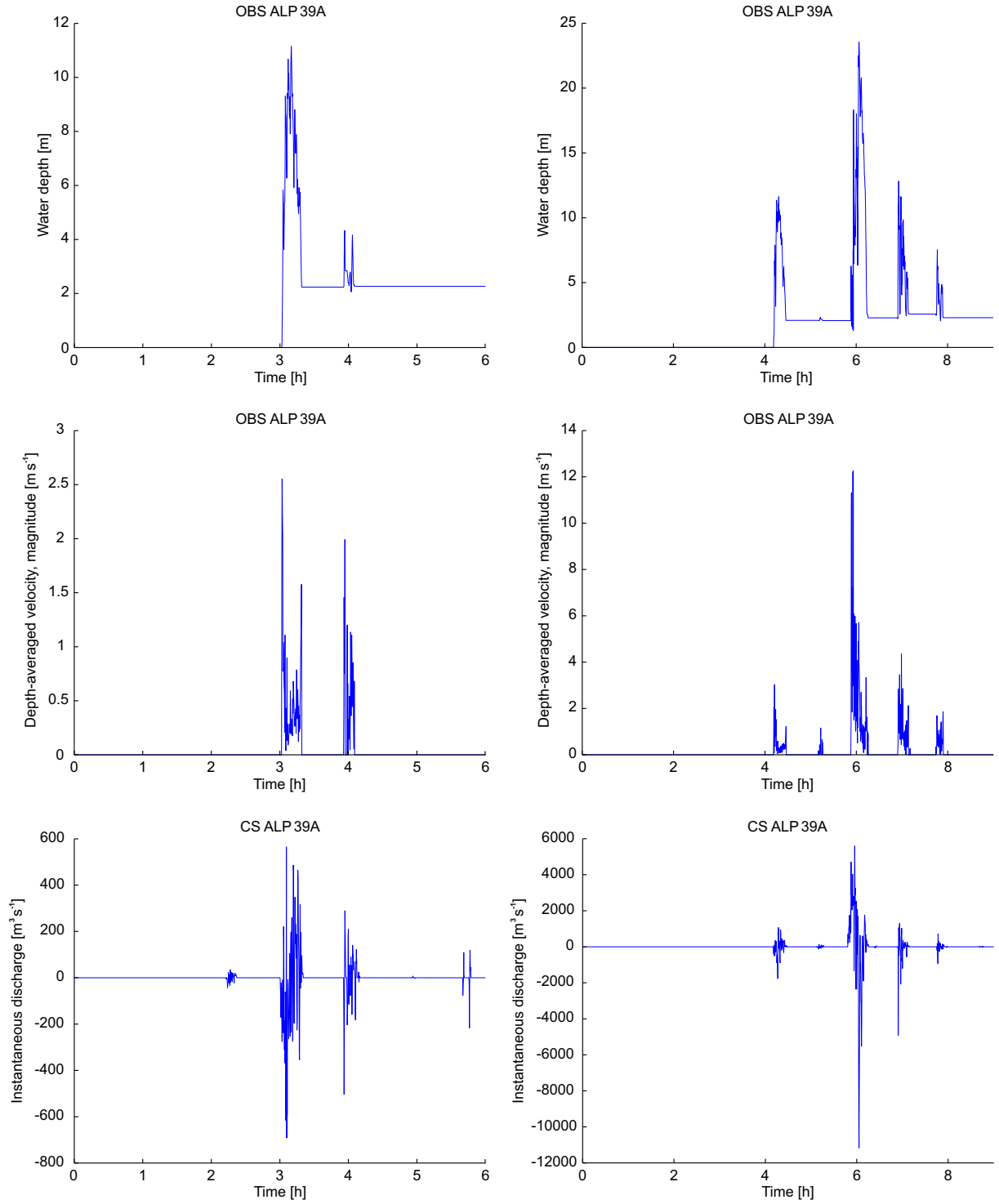


Figure 4.10: Diagrams of the water depth, depth-averaged flow velocity and instantaneous discharge as a function of time for the Epitalio Pass/Fan in Scenario_{13S} (left column) and Scenario_{13N} (right column). Water depth and flow velocity are recorded for the 30 m grid cell at OBS ALP 39A. The discharge relates to CS ALP 39A. Both, OBS and CS ALP 39A are illustrated in Fig. 4.4.

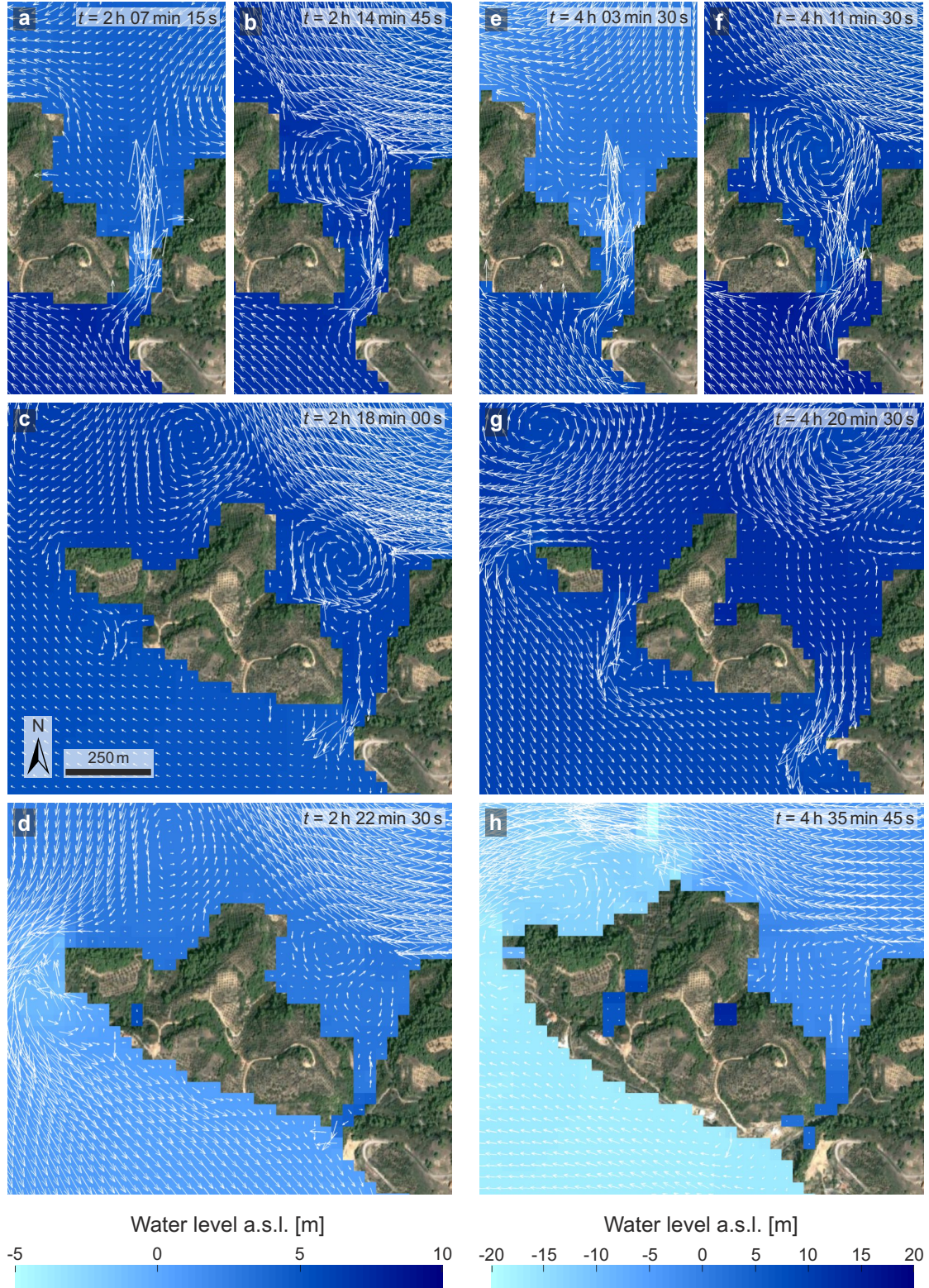
SCENARIO_{30S} — SECOND SOLITARY WAVESCENARIO_{30N} — SECOND LDN-WAVE

Figure 4.11: Snapshots documenting stepwise the inundation of the Epitalio Pass associated with the second, 3m wave of the solitary (Scenario_{30S}, left column) and LDN-tsunami-wave train (Scenario_{30N}, right column) based on DEM₃₀ (Fig. 4.1c). White vectors indicate the depth-averaged flow direction and velocity. See Fig. 4.1a and Fig. 4.4 for the location and bird's-eye view of the site respectively.

again, the LDN-wave has a stronger impact on the site, causing also inundation of another, smaller pass to the north (Fig. 4.11g).

In the DEM_{rec}-Scenarios, a persistent eddy is observed in the Epitalio Valley during the landfall of the third wave in Scenario_{recS} and Scenario_{recN}. However, the Epitalio Pass—about 30 m a.s.l in DEM_{rec}—is not inundated by any of the waves.

Comparison with field traces

The hydrodynamics observed in the numerical model are well suited to explain the genesis of the Epitalio Fan described in Chap. 4.3.2 with regard to tsunami impact. The large inundation depth, flow velocity and discharge calculated for the third solitary wave in Scenario_{13S} (Fig. 4.10, left column) suggest a clear geomorphological impact of the wave on the Epitalio Pass and environs. In accordance with field observations by VÖTT *et al.* (2015), the wave is expected to significantly erode the flanks of the Epitalio Pass on the one hand, and to build a sedimentary fan structure leewards the pass in the area of flow divergence on the other hand. A large eddy observed in the valley of Epitalio in Scenario_{13S} may strongly erode and steepen the adjacent slopes, as is also described by VÖTT *et al.* (2015) from the field (Figs. 4.3, 4.4). The authors further found the leeward front of the Epitalio Fan clearly eroded, which may be related to strong E-W-currents (up to 9 m s^{-1}) during the backflow of tsunami waters in the Alpheios River valley (Fig. 4.9c).

The lagoonal to deep water and local bedrock signal found in the event deposits of the Epitalio Fan closely agrees with the path observed in Scenario_{13S} for a water particle in the depth-averaged flow in the simulation time between 2 h 57 min and 3 h 18 min (white trajectory in Fig. 4.1b). Hypothetically, sediment erosion, transport and deposition associated with the third solitary wave are as follows: The trajectory starts to the west of the Alpheios River valley and runs in a south-eastward direction following the course of reflection waves that originate from the ridge to the north of the former Mouria Lagoon. On their way to the Epitalio Pass, tsunami waters pick up coarse-grained sediment in the shelf area. With these sediments being transported in suspension and as bed load, the water enters the Epitalio Pass, where severe erosion occurs due to funnelling and increased flow velocity. Here, allochthonous lagoonal to marine sediments mix with material eroded from the bedrock. The mixed sediments are finally released from the diverging stream leewards from the pass, producing clear intersections of high-energy deposits in the geological record of the Epitalio Fan.

Although the third LDN-wave in Scenario_{13N} shows similar hydrodynamics as described for the third solitary wave (Chap. 4.4.2), the LDN-wave even closer agrees with field observations. This is especially related to the stronger onshore response and therefore the stronger geomorphological impact of the LDN-wave, which correspond to the explicitness of the field findings. Regarding the grey trajectory in Scenario_{13N} (simulation time 2 h 40 min to 2 h 57 min; Fig. 4.1b), also the deep water signal found in the event deposits of the Epitalio Fan is easier to explain with the 6 m LDN-wave because the trajectory shows tsunami waters directly approaching from the open sea. Furthermore, the strong jet-like flow observed on the leeward side of the pass (Fig. 4.9d) may be related to channel structures recorded in geoelectrical tomograms across the Epitalio Fan (VÖTT *et al.* 2015).

In the DEM₃₀-Scenarios, a wave amplitude of 3 m for the solitary and LDN-wave is sufficient to produce hydrodynamics in close accordance with field data since those waves show the same inundation pattern as described for the 6 m wave in Scenario_{13N} (Chap. 4.4.2). The DEM_{rec}-Scenarios do not allow a concrete comparison with the field findings because the Epitalio Pass is not overflowed by tsunami waters in the numerical model. Consequently, the fan structure in the Epitalio Valley cannot be explained by those scenarios.

4.4.3 Valley of Kato Samiko

Simulation results

In Scenario_{13S} and Scenario_{13N}, with each wave an almost linear run-up and backflow is associated in the valley of Kato Samiko, which is covered by water 2 m deep at OBS SAM 1 in DEM₁₃. Whereas inundation by the 1 m waves is not significant, the 3 m and 6 m waves have a marked impact. Figure 4.12 illustrates the up-running tsunami waters (a, d), the maximum water level and run-up (b, e) as well as the backflow (c, f) in the valley for both, the 6 m solitary and LDN-wave. Snapshots of each step show the LDN-wave to stronger affect the valley, connected with hydrodynamics being more turbulent. The latter is most evident from Fig. 4.12e, which shows eddies on both sides of the valley's entrance. The stronger onshore response of the LDN-wave is also reflected by the maximum inundation depth at coring site SAM 1, which is nearly 42 m in Scenario_{13N} and 28 m in Scenario_{13S} (Fig. 4.13, top). Furthermore, the maximum instantaneous discharge through the valley (CS SAM 1) is more than three times larger in Scenario_{13N} ($5700 \text{ m}^3 \text{ s}^{-1}$) compared to Scenario_{13S} ($1700 \text{ m}^3 \text{ s}^{-1}$) (Fig. 4.13, bottom). The maximum depth-averaged flow velocity is associated with the backflow of the 6 m solitary (6 m s^{-1}) and LDN-wave (12 m s^{-1}) (Fig. 4.13, centre). Contrary to the run-up, the backflow in both scenarios is concentrated in a narrow channel (cf. Figs. 4.12a,c,d,f), which explains the high flow velocities.

In the DEM₃₀-Scenarios, the valley of Kato Samiko is largely covered by water, with OBS SAM 1 lying approximately 19 m b.s.l. As mentioned previously, each wave of both wave trains produces an almost linear run-up and backflow in the valley. Due to the lower elevation of the valley in DEM₃₀ compared to DEM₁₃, the run-up deeper penetrates inland. However, the backflow is less concentrated in the valley axis in the DEM₃₀-Scenarios. In Scenario_{recS} and Scenario_{recN}, the 1 m and 3 m solitary and LDN-waves do not reach the valley of Kato Samiko (circa 11 m a.s.l. at OBS SAM 1), whereas inundation dynamics associated with the 6 m waves are similar to those described for the DEM₁₃-Scenarios. Thus, a significant run-up (Figs. 4.14a,d) and backflow (Figs. 4.14c,f) is observed in the valley, separated by a period of maximum water level (Figs. 4.14b,e). As before, the backflow is concentrated in a narrow channel in the valley axis and causes the maximum depth-averaged flow velocity in Scenario_{recS} (4.5 m s^{-1}) and Scenario_{recN} (10.5 m s^{-1}).

Comparison with field traces

Once again, field findings are in close agreement with the numerical model. In the DEM₁₃-Scenarios, vibracoring site SAM 1 is affected by all three waves. Whereas the 1 m waves are not significant in view of the field traces, a geomorphological impact is obvious from the 3 m and 6 m waves. Judging by the inundation parameters (Fig. 4.13) and the described hydrodynamics, the backflow is assumed to significantly erode unconsolidated sediment in the valley and to build a channel structure broadening to the seaside, as is indicated with regard to the youngest event by geoelectrical tomograms running perpendicular to the valley axis (Chap. 4.3.3). Geoelectrical data and vibracore SAM 1 further document a coarse-grained channel fill deposit out of sand, gravel and abundant ceramic fragments, which may be related to the backflow slowing down with time (Fig. 4.13, centre) and thus releasing transported sediment from tsunami waters. The (shallow) marine character of the channel infill and of the older event deposits found in SAM 1 (WILLERSHÄUSER *et al.* 2015a) closely accords with the corresponding white and grey trajectories in Fig. 4.1b, which show the water particle paths associated with the 6 m solitary and LDN-wave respectively (simulation time 2 h 40 min to 3 h 12 min in Scenario_{13S} and 5 h 40 min to 6 h 5 min in Scenario_{13N}). Tsunami currents hypothetically pick up coarse-grained sediment

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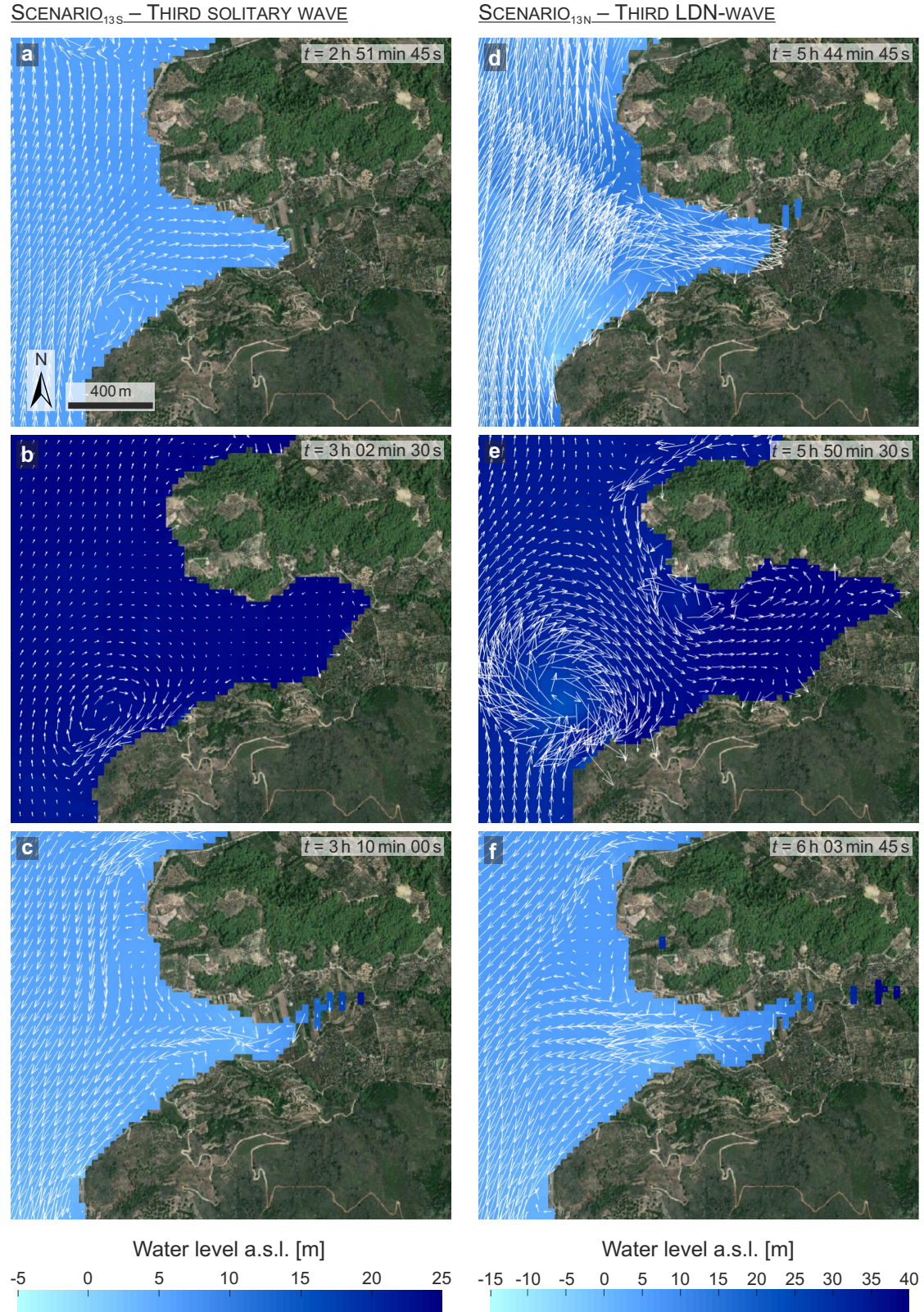


Figure 4.12: Snapshots documenting stepwise the inundation of the valley of Kato Samiko associated with the third, 6 m wave of the solitary (Scenario_{13S}, left column) and LDN-tsunami-wave train (Scenario_{13N}, right column) based on DEM₁₃ (Fig.4.1b). White vectors indicate the depth-averaged flow direction and velocity. See Fig.4.1a and Fig.4.5 for the location and bird's-eye view of the site respectively.

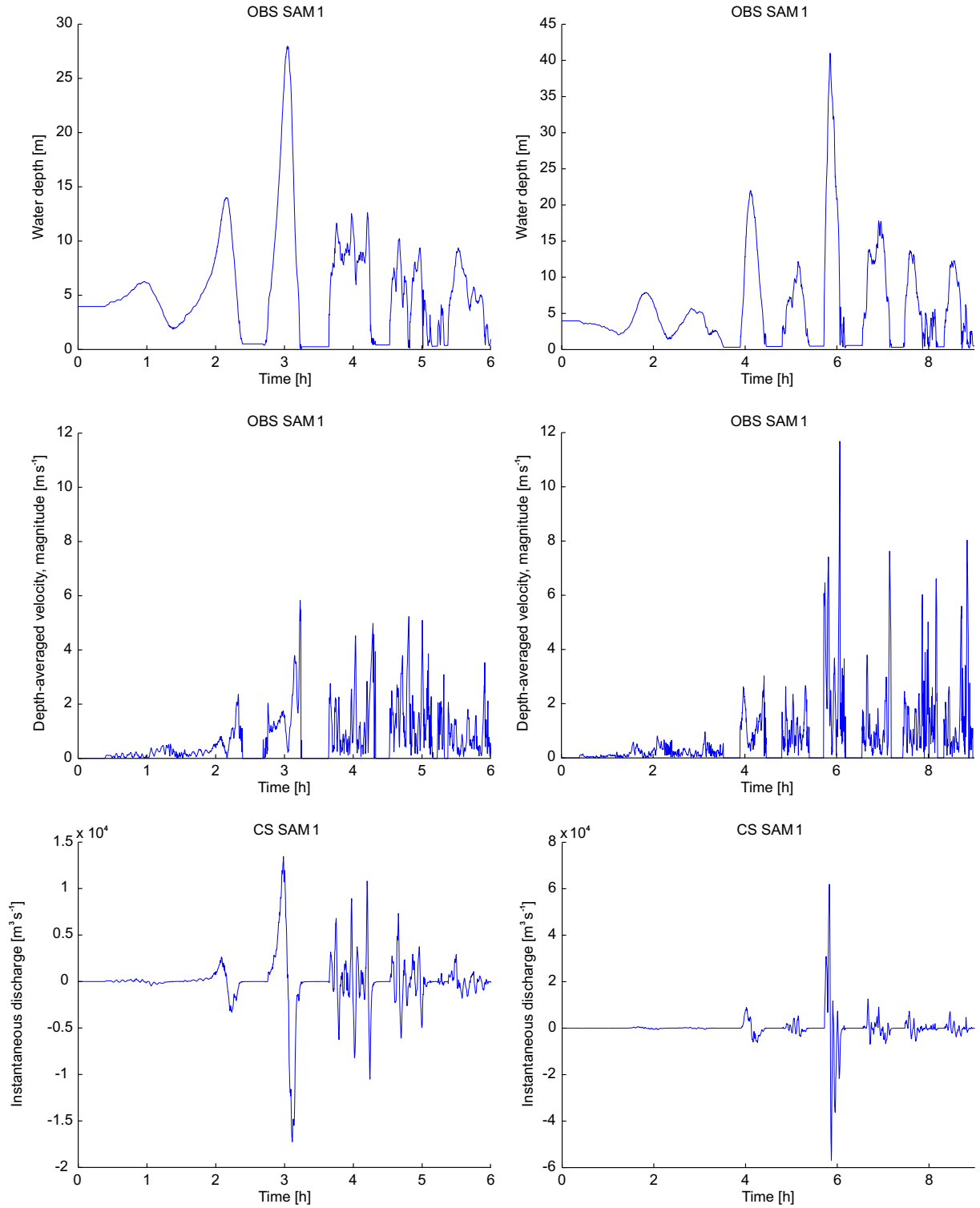


Figure 4.13: Diagrams of the water depth, depth-averaged flow velocity and instantaneous discharge as a function of time for the valley of Kato Samiko in Scenario_{13S} (left column) and Scenario_{13N} (right column). Water depth and flow velocity are recorded for the 30 m grid cell at OBS SAM 1. The discharge relates to CS SAM 1. Both, OBS and CS SAM 1 are illustrated in Fig. 4.5.

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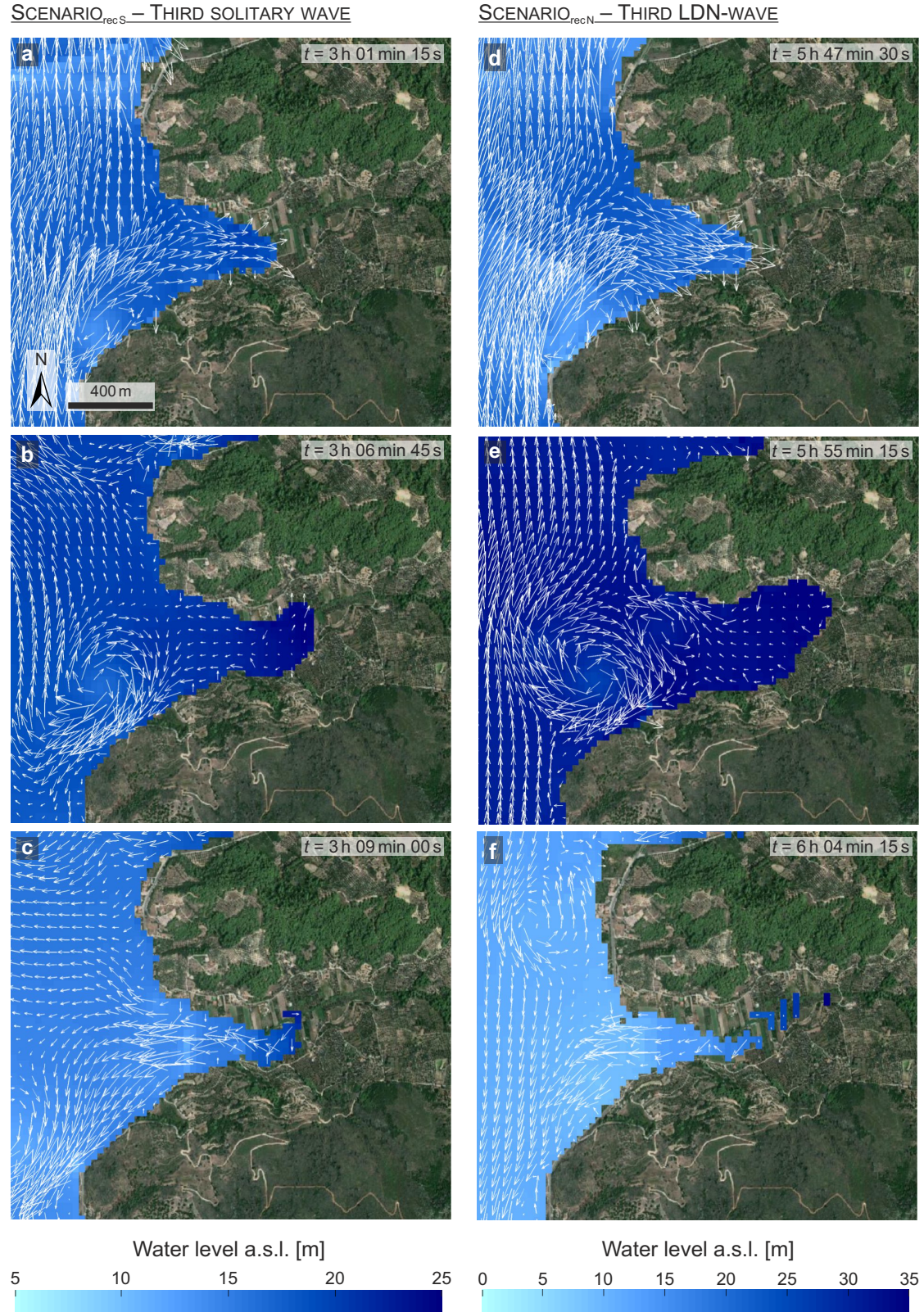


Figure 4.14: Snapshots documenting stepwise the inundation of the valley of Kato Samiko associated with the third, 6 m wave of the solitary (Scenario_{recS}, left column) and LDN-tsunami-wave train (Scenario_{recN}, right column) based on DEM_{rec} (Fig. 4.1a). White vectors indicate the depth-averaged flow direction and velocity. See Fig. 4.1a and Fig. 4.5 for the location and bird's-eye view of the site respectively.

in the shallow waters of the shelf and in the pre-event littoral zone. Subsequently, the sediment is directly transported to the valley of Kato Samiko and released in the upper valley during the maximum run-up on the one hand and in the backflow channel (as described above) on the other.

The simulation results of Scenario_{30S} and Scenario_{30N} are less suitable to explain the field traces reported by WILLERSHÄUSER *et al.* (2012, 2015a) since the backflow of tsunami waters is no longer concentrated in a narrow channel in DEM₃₀. With hydrodynamics being similar to those described for DEM₁₃ (Chap. 4.4.3), simulation results show that also the 6 m solitary and LDN-wave in the DEM_{rec}-Scenarios may cause channel formation and infill in the valley of Kato Samiko. Therefore, the field traces are consistent with tsunami landfall also in the case of the recent topography and coastline configuration.

4.4.4 Synthesis

The different hydrodynamic boundary conditions and the different palaeo-/DEMs are found to significantly affect tsunami hydrodynamics computed for the northern Gulf of Kyparissia. This applies to both, the magnitude of flooding and the inundation pattern. Due to the *N*-wave effect, all LDN-waves have a stronger onshore response in the study area than the corresponding solitary waves (for more details see Chap. 3). Furthermore, the critical wave amplitudes associated with inundation of the three key locations decrease from DEM_{rec} over DEM₁₃ to DEM₃₀, which is related to the subtraction of coastland elevations by 13 m and 30 m in the palaeo-DEMs. Therefore and because of the *N*-wave effect, Scenario_{30N} requires the lowest wave amplitudes to cause significant inundation of the key sites.

The impact of the different boundary conditions and palaeo-/DEMs on tsunami hydrodynamics and connected with this, the suitability of the different scenarios to explain tsunami field traces are summarised for each key location as follows:

- Regarding the Katakolo Pass, a 6 m solitary wave as well as a 3 m LDN-wave, both based on DEM₁₃, cause strong inundation of the pass and compare favourably with field findings by VÖTT *et al.* (2010a, pp. 32–39) and VÖTT *et al.* (2011b). The lower wave amplitude and the higher probability of LDN-waves in the gulf (Chaps. 3.4, 4.2.2) make this scenario even more realistic. In the DEM₃₀-Scenarios, even the 1 m solitary and LDN-wave are expected to have a significant geomorphological response in the area of the pass, although the general water coverage hinders detailed conclusions about the actual sedimentary impact of inundation. Considering the recent topography (DEM_{rec}), field traces could only be explained by the 6 m LDN-wave, whereas the solitary wave train does not affect the pass at all.
- DEM₁₃ requires a 6 m solitary or rather a 6 m LDN-wave to cause severe inundation across the Epitalio Pass. Geomorphological observations by VÖTT *et al.* (2015) are in preference to the LDN-wave, which closer agrees with (i) the microfossil deep water signal found in the event layers, (ii) the channel structures inherent in the Epitalio Fan, (iii) erosive traces along the leeward edge of the Epitalio Fan and (iv) the general explicitness of field findings. This is mainly due to the stronger onshore response of the LDN-wave compared to the solitary wave, and also to the fact that inundation of the Epitalio Pass is not bound to reflection waves in Scenario_{13N} but directly caused by the main tsunami wave crest. In DEM₃₀, already a 3 m solitary and LDN-wave may leave field traces as reported by VÖTT *et al.* (2015), both waves showing hydrodynamics very similar to those computed for the third wave in Scenario_{13N}. Finally, neither wave train affect the Epitalio Pass in DEM_{rec} and therefore cannot be compared with field findings.

- Differences between the six scenarios are less distinct regarding the valley of Kato Samiko. In DEM₁₃, the 3 m and the 6 m solitary and LDN-waves generate a concentrated strong backflow in the valley, which may be in direct connection with the channel structure and infill discovered in the field by WILLERSHÄUSER *et al.* (2012, 2015a). The same applies to DEM_{rec}, although the funnelled backflow is here produced by the 6 m waves only. In contrast, backflowing tsunami waters in the DEM₃₀-Scenarios are not concentrated in the valley axis but affect the whole valley area.

4.5 Conclusions

The tsunami scenarios simulated for the northern Gulf of Kyparissia in this study show marked differences concerning the hydrodynamics in the three key locations. This is related to the stronger onshore response of LDN-waves compared to the solitary waves (*N*-wave effect), and considerable topographical discrepancies between the three DEMs (Chap. 3.2.1). Differences between the scenarios are not restricted to the magnitude of flooding but also evident from a varying inundation pattern. Both may strongly affect the sedimentary and geomorphological response of tsunami waves in the study area and therefore are of great importance for the synthesis of simulation results and field data.

Significant inundation of the Katakolo and Epitalio Pass—situated on higher ground—is bound to larger wave amplitudes compared to the valley of Kato Samiko. Consequently, the valley of Kato Samiko may be affected by tsunami events the two passes are protected from. This is also suggested by field data as both passes do not show event layers that correlate with the event-related backflow channel infill in the valley of Kato Samiko. Generally, field findings compare favourably with the DEM₁₃-Scenarios, since here, all key locations show appropriate inundation characteristics. These are also the favoured scenarios in view of the palaeotopography because DEM₁₃ is based on reliable neotectonic field data documenting a minimum uplift of 13 m along the EFZ and adjacent fault zones during the past 6000 years (VÖTT *et al.* 2015). Nevertheless, there is no particular scenario or individual wave that can fully explain the field traces in all three key locations at once. This may be related to a number of reasons:

- High-energy sediments were not deposited during one and the same tsunami event but during different events, as is also implied by the wide range of age variation found for correlating event layers (Chap. 4.3).
- The event layers were modified by following inundation events or weathering processes, which may mislead the interpretation of the local hydrodynamics.
- The tectonic throw along the EFZ and adjacent fault zones is irregular and not uniform as assumed in the palaeo-DEMs. This may have serious effects on the topography in the area of the key locations, which, in turn, would strongly affect the local hydrodynamics. In a similar way, the detected tsunami deposits and overlying sediment layers, which are not removed in the palaeo-DEMs, may crucially modify the palaeotopography and, by this, the simulation results.
- The various hydrodynamic boundary conditions do not reflect the actual palaeotsunami wave characteristics in the Gulf of Kyparissia and eastern Ionian Sea, but instead tsunami waves with differing propagation directions, waveforms, wave periods or wave amplitudes have occurred.

- Numerical parameters of the hydrodynamic model, as for instance the chosen values for bottom roughness or the computed energy dissipation due to wave breaking, do not meet the actual conditions in the study area, which may have a crucial impact on the computed hydrodynamics.

Although several scenarios and tsunami waves are needed to match field findings in all key sites, the investigation of this chapter demonstrates the close agreement between numerical simulation results and palaeotsunami field traces in the northern Gulf of Kyparissia and emphasises the high reliability of hydrodynamic tsunami modelling based on palaeotopographies that changed through time due to neotectonic crust movements. Both together, the numerical model and field traces therefore allow a detailed reconstruction of palaeotsunami events in the study area.

5 Hydro- and morphodynamic tsunami simulations for the Ambrakian Gulf (Greece) and comparison with geoscientific field traces¹

Abstract Modern tsunami risk assessment uses a multi-methodological approach based on oceanic warning systems, seismological data, numerical simulation techniques and sedimentary field traces. In order to derive local tsunami risks for a particular coast, hydrodynamic numerical models that consider sediment processes and are compared and calibrated with sedimentary field data of past tsunami impacts have proven very effective. While this approach has widely been used with regard to recent tsunami events, there are no comparable investigations into pre-/historical tsunami impacts, which is the objective of this study focusing on the Ambrakian Gulf in north-western Greece.

Associated with the nearby subduction zone of the African and Eurasian Plate, the Ambrakian Gulf is located in the most active seismo-tectonic and by this most tsunamigenic area of the Mediterranean. Accordingly, palaeotsunami field studies have revealed repeated tsunami impacts on the gulf during the past 8000 years. The current study analyses 151 vibracores of the Ambrakian Gulf coast in order to evaluate tsunami signals in the sedimentary record. Based on a hydro- and morphodynamic numerical forward model of the study area, various tsunami waves are simulated with the aim of finding scenarios that compare favourably with the spatial distribution of tsunami deposits detected in the field.

Both, field data and simulation results suggest a decreasing tsunami influence from the western to the eastern Ambrakian Gulf. Various scenarios are needed to explain tsunami deposits in different parts of the gulf, as is also indicated by the field data itself. Whereas short-period tsunami waves from the south and west favourably reproduce field data in the western gulf, long-period waves from a western direction show the best agreement with tsunami sediments detected further eastwards. Tsunamis from the south-west generally do not accord with field traces. Besides the spatial sediment distribution, the numerical model also reproduces a number of essential features typical of tsunamites which were also observed in the field, including fining- and thinning-landward, basal erosional contacts and the marine character of the deposits. By contrast, the computed thickness of tsunami sediments usually lags behind the observed thickness in the field and some event layers in the study area cannot be explained by any of the simulated scenarios. This may arise from a number of reasons, e.g. the use of the average thickness of detected tsunami layers for the comparison with the computed sedimentation or changes in the local topography through time, which are not considered in the numerical model. Nevertheless, regarding the frequency of past tsunami events and their spatial dimensions indicated by both field and simulation data, a high tsunami risk has to be derived for the Ambrakian Gulf.

¹This chapter is based on the paper by RÖBKE *et al.* (2016b) submitted to *Geophysical Journal International*.

5.1 Introduction

With more than 10 % of the world's population living in coastal lowlands less than 10 m a.s.l. (McGRANAHAN *et al.* 2007, p. 22), the vulnerability of coasts to tsunami inundation is very high. Recent history has seen disastrous tsunami events in Papua New Guinea (1998), south-east Asia (2004) and Japan (2011), which have particularly emphasised the need for improved risk assessment and mitigation plans. During the last years, tsunami forecast methods have significantly extended and gained in reliability. Modern tsunami risk assessment uses a multi-methodological approach based on oceanic warning systems, seismological data, numerical simulation techniques and sedimentary field traces. In order to derive tsunami risks for a particular coast, hydro- and morphodynamic numerical tsunami models that are compared and calibrated with sedimentary field data and—if available—buoy measurements and information on the trigger mechanism of past tsunami impacts have proven very effective. Knowing the frequency and the spatial extent of pre-/historical tsunami inundation derived from field traces makes it possible to draw conclusions on future tsunami risks. However, field traces only allow an accurate reconstruction of former inundation events when combined with numerical tsunami models, as performed in numerous recent studies (e.g. BONDEVİK *et al.* 2005a; MARTIN *et al.* 2008; BORRERO *et al.* 2009; RÖBKE *et al.* 2013; SATAKE *et al.* 2013; SUGAWARA *et al.* 2013; HILL *et al.* 2014; Chaps. 3, 4). Such studies are particularly promising, if numerical models do not only compute the pure tsunami hydrodynamics but also take account of sediment processes, which allows a one-to-one comparison between field data and model (cf. SUGAWARA *et al.* 2014a). This approach was taken in a number of studies focusing on tsunami impacts in recent history, in particular the Indian Ocean tsunami 2004 in Sumatra and Thailand (GELFENBAUM *et al.* 2007; APOTSOS *et al.* 2009, 2011a,b; ONTOWIRJO *et al.* 2010, 2013; GUSMAN *et al.* 2012; LI *et al.* 2012a,b, 2014). Studies comparing sedimentary field data with hydro- and morphodynamic simulations have further been conducted with regard to the 2009 South Pacific tsunami in American Samoa (APOTSOS *et al.* 2011c) and the 2011 Tōhoku tsunami in Japan (SUGAWARA *et al.* 2014b). However, there are no studies combining sedimentary field data and morphodynamic numerical tsunami models in view of pre-/historical tsunamis, which is the approach of the following investigation.

This chapter concentrates on tsunami impacts in the Ambrakian Gulf (north-western Greece) during the last 8000 years. The Ambrakian Gulf is a marine inlet of the eastern Ionian Sea in coastal Epirus/Akarnania with a maximum N-S- and W-E-extent of about 20 km and 35 km respectively (Figs. 1.1, 5.1). The gulf is connected to the Ionian Sea by the Strait of Preveza, a 600 m wide shallow channel less than 25 m deep (UKHO 2010). Maximum water depths are found in the eastern gulf with roughly 65 m. The Ambrakian Gulf forms a tectonic basin dominated by N-S crustal extension along WNW-ESE trending normal faults (CLEWS 1989, p. 454). According to palaeogeographical studies by POULOS *et al.* (1995, 2005), TZIAVOS (1997), JING & RAPP (2003), BROCKMÜLLER *et al.* (2005) and KAPSIMALIS *et al.* (2005), the gulf was isolated from the Ionian Sea during the last ice age (Weichselian or Würm period), but became a marine inlet in the early Holocene (epoch of the last 10 000 years). By the 4th millennium BC, the ongoing rise in sea level had resulted in the complete flooding of the present river deltas in the northern gulf, whereas the coastline in the western, southern and eastern gulf approximately reached its present position. Subsequently, considerable sediment input by the rivers Louros and Arachthos created the large delta complex nowadays found in the northern gulf (Fig. 5.1). Since the 5th millennium BC, tectonic crust movements have resulted in a relative uplift by 6 m of the north-western gulf compared to the south-eastern part (BROCKMÜLLER *et al.* 2005, pp. 46, 47).

Located in the most tsunamigenic region in the entire Mediterranean, i.e. western Greece (Chap. 1.2; SOLOVIEV *et al.* 2000, p. 14; SCHIELEIN *et al.* 2007, pp. 164, 193–199; RÖBKE *et al.*

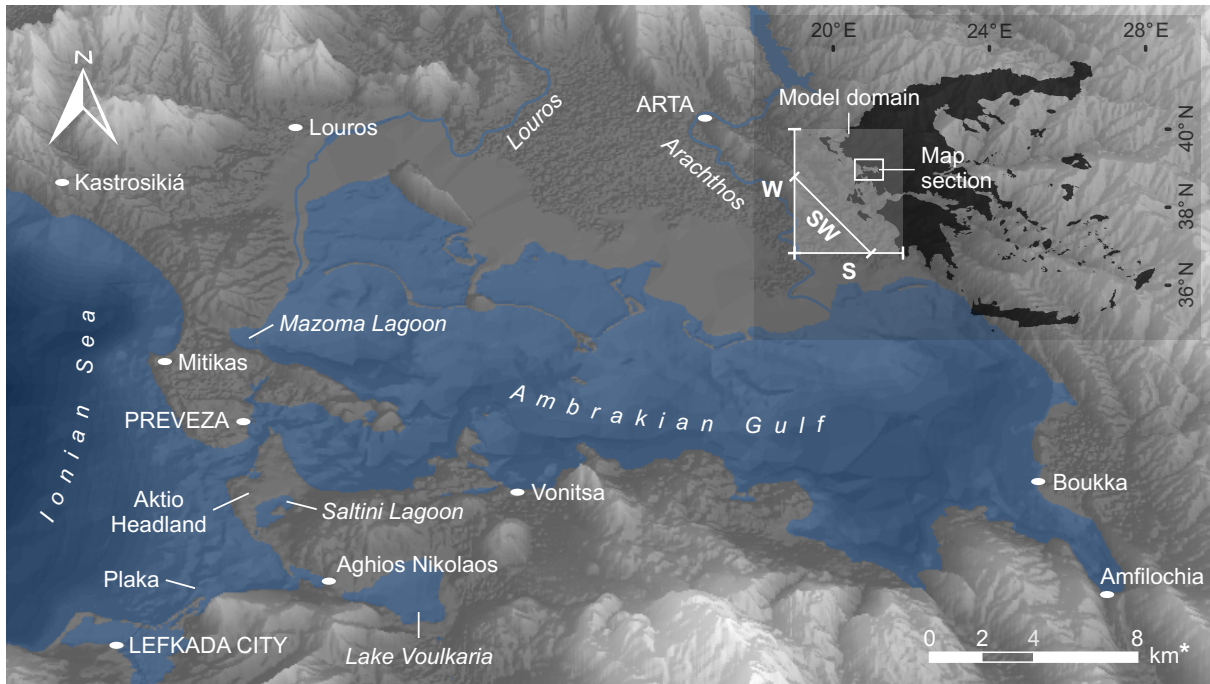


Figure 5.1: Bird's eye view of the Ambrakian Gulf, including the opening to the Ionian Sea, the coastal flats in the north and the hilly shore in the south-east. The figure is based on a 3-D excerpt of the digital elevation model employed for numerical tsunami simulations in this study. The overview map of Greece in the top right-hand corner shows the numerical model domain and the western, south-western and southern hydrodynamic model boundaries. The blue surface indicates the water surface. *Due to the 3-D view, the scale bar gives the true distance only in the centre of the figure.

2013, pp. 70–77), the Ambrakian Gulf and its environs were frequently affected by tsunami landfall in the past. Besides historical accounts, this is documented by sedimentary field traces in the area. According to detailed field studies by VÖTT *et al.* (2006, 2007, 2008, 2009b, 2010a) and MAY *et al.* (2007, 2012a), major tsunamis struck the Ambrakian Gulf in the 6th millenium BC or shortly afterwards, between 2870 cal BC and 2350 cal BC, around 1000 cal BC, between 395 cal BC and 247 cal BC, in the 4th century AD and around 840 cal AD. Further events occurred in the last 700 years. Sedimentary signatures of most of these impacts were also found in close vicinity to the gulf on the nearby island of Lefkada (VÖTT *et al.* 2006, 2008, 2009a; MAY *et al.* 2012a,b) and in the Bay of Palairos (VÖTT *et al.* 2011a) and seem to be related to supraregional tsunami events in the Mediterranean. The wider study area is the northernmost region in the eastern Mediterranean where sedimentary signatures of the 365 AD Crete tsunami were found (e.g. VÖTT *et al.* 2009a; MAY *et al.* 2012b). Besides field studies, hydrodynamic numerical tsunami simulations were computed for the coastal area between Preveza and Lefkada City (FLOTH *et al.* 2009, 2010, 2013). In this initial approach, simulation results, which were based on a small-scale numerical model of the area, were compared with local tsunami field traces.

The current study aims to (i) analyse and evaluate systematically tsunami signals in the sedimentary record of 151 vibracores taken all along the Ambrakian Gulf within previous palaeotsunami field studies, (ii) derive the scale of pre-/historical tsunami landfall in the gulf based on field data, (iii) set up a hydro- and morphodynamic numerical tsunami forward model, which takes account of various tsunami wave scenarios and considers sediment processes, (iv) compare and calibrate the numerical model with the spatial distribution of detected tsunami deposits in

the field and find tsunami scenarios that replicate field data in order to (v) assess tsunami risks for the Ambrakian Gulf.

5.2 Methodology

5.2.1 Field work

Within in the framework of previous palaeogeographical studies, about 151 vibracores were drilled alongshore the Ambrakian Gulf using a handheld engine-powered Atlas Copco coring device (type Cobra mk1). The final coring depth at each location (maximum 18 m below ground surface) was reached, when either pre-Holocene deposits or bedrock appeared or the sedimentary facies did not change for several metres. Vibracores were first documented by digital photographs and subsequently described regarding their sedimentary features, such as grain size, texture, colour, fossil content and pedogenetic properties (cf. Ad Hoc-AG Boden 2005). The exact location of all vibracores was determined using a Topcon DGPS-device (type HiPer Pro).

5.2.2 Evaluation of field data

A number of vibracores taken in the Ambrakian Gulf had already been analysed in view of the regional palaeogeography and palaeotsunami signatures by VÖTT *et al.* (2006, 2007, 2008, 2009b, 2010a, 2011a) and MAY *et al.* (2007, 2012a). Following these investigations, in the current study, all 151 vibracores were analysed and evaluated systematically with regard to tsunami candidate layers based on a set of sedimentary criteria which have been associated with onshore tsunamites worldwide (e.g. DAWSON 1994; DAWSON & SHI 2000; DAWSON & STEWART 2007a, pp. 173–180; DAWSON & STEWART 2007b, pp. 578–581; DOMINEY-HOWES *et al.* 2006; MORTON *et al.* 2007; FUJIWARA 2008; SHIKI *et al.* 2008; VÖTT & MAY 2009, pp. 45, 46; SHANMUGAM 2012; cf. Chap. 2.4). Such physical criteria are directly ascribed to the hydrodynamic characteristics of tsunami inundation and particularly include

- basal erosional unconformities,
- fining-upward sequences,
- high content of marine macrofossils,
- clayey to silty rip-up clasts from underlying units,
- mud caps at the top of event layers,
- fining- and thinning-landward,
- a general allochthonous, high-energy character of the sediment (cf. Fig. 5.2).

For some vibracores, also geochemical, microfaunal and palynological analyses had been conducted, which further facilitated the identification of tsunami deposits in the study area (BROCK-MÜLLER *et al.* 2005; JAHNS 2005; VÖTT *et al.* 2006, 2007, 2008, 2009b, 2010a, 2011a; MAY *et al.* 2007, 2012a; cf. Chap. 2.4). Findings from these studies together with physical sedimentary features described above and correlations between related event strata at neighbouring vibracoring sites allowed a systematic and standardised identification of tsunami deposits along the entire Ambrakian Gulf coast. In this process, also storm events were taken into account as a potential mechanism to leave high-energy deposits. However, a thorough investigation into the possible

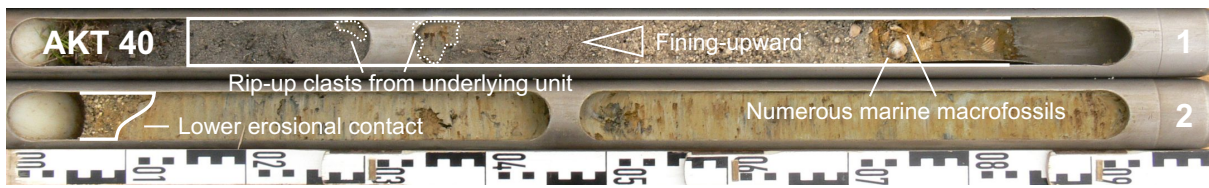


Figure 5.2: Vibracore AKT 40 showing sedimentary features typical of tsunami deposits, including a basal erosional unconformity, numerous marine macrofossils, normal grading and clayey to silty rip-up clasts from underlying units. Such features, among others, were referred to in order to identify tsunami candidate layers in 151 vibracores alongshore the Ambrakian Gulf shown in Figs. 5.4–5.6.

influence of storm surges on the Ambrakian Gulf and a careful analysis of field data led to the following key conclusions, which strongly suggest the event deposits being of tsunamigenic origin (VÖTT *et al.* 2007, pp. 51–53; VÖTT *et al.* 2008, pp. 116–119; VÖTT *et al.* 2011a, pp. 233, 234):

- Although the prevailing wind and waves in the Ionian Sea and western Greece are from the west and north-west and despite the occasional appearance of so-called Medicanes in the Mediterranean during winter (hurricane-like but significantly weaker cyclones), there are no written accounts of extraordinary storm surges causing severe inundation of the Ambrakian Gulf or the west coast of Greece. By contrast, multiple sources of data give evidence of several tsunami impacts on the study area and environs (Chaps. 1.2 and 5.1).
- The geomorphological appearance of the Ambrakian Gulf coast, in particular the very narrow beaches do not imply a recent to subrecent influence of storm surges beyond the supralittoral zone, where most of the high-energy deposits were detected. This closely accords with regional buoy and tide gauge data and studies by GHIONIS *et al.* (2015) and POULOS *et al.* (2015, pp. 43–45) that revealed an expected maximum run-up of wind generated waves of only 2.82 m a.s.l. for northern Lefkada, located less than 10 km to the south-west of Aktio Headland at the entrance to the Ambrakian Gulf. For the latter, comparable run-up heights can easily be expected due to similar bathymetric conditions and the same wave climate. Regarding the inner gulf coast, even lower maximum run-ups of wind-generated waves have to be assumed given the isolation of the gulf from the open Ionian Sea and the short maximum fetch in the gulf of less than 40 km (cf. Fig. 5.1). These arguments also hold true for prehistorical times when the Ambrakian Gulf was even more isolated from the Ionian Sea than it is today (Chap. 5.1).
- Vibracores in the study area did not reveal any sedimentary characteristics which are typically associated with tempestites, i.e. in particular rhythmic intercalations of storm sand sheets into nearshore lagoonal or limnic muds showing planar stratification and climbing ripples (e.g. MORTON *et al.* 2007; Chap. 2.4).
- Judging by dating approaches in previous studies, several event layers in vibracores of the Ambrakian Gulf correlate well with tsunami deposits described from other coasts in western Greece (Chap. 1.2) and by this point to the same tsunami events.

Owing to the large number of vibracores and identified event deposits alongshore the Ambrakian Gulf, the data set was strongly simplified in order to achieve a clear and meaningful overview of field traces in the study area and to allow a detailed comparison with simulation results. This simplification could not be done on the basis of specific tsunami events as correlations between related event strata could only be drawn for selected vibracores for which geochronological data was available. Instead, three different parameters were determined, namely (i) the total

number and (ii) the cumulative thickness of all tsunami candidate layers per vibracore, both divided by the coring depth, as well as (iii) the average thickness of all tsunami candidate layers per vibracore (Chap. 5.3.1). The total number and cumulative thickness of tsunamite candidates per vibracore strongly depend on the local coring depth (0.72 m to 18 m below ground surface), in that longer vibracores may contain more event strata than shorter vibracores. Consequently, both quantities were divided by the coring depth yielding the relative number and relative cumulative thickness of event layers, which allowed the evaluation of all vibracores independent from the coring depth. This was especially important with regard to Aktio Headland, where bedrock is close to ground surface and vibracores generally show a small coring depth (< 5 m) resulting in fewer event strata compared to other areas in the Ambrakian Gulf.

The relative number of tsunamite candidates per vibracore gives a rough indication of the local susceptibility to tsunami inundation measured by the number of tsunami impacts, whereas the average layer thickness rather implies the mean intensity of inundation (i.e. inundation depth and flow velocity) per event for each location (cf. LI *et al.* 2014, pp. 2264, 2265). Consequently, the relative cumulative thickness per vibracore may be indicative of both, the frequency and the intensity of tsunami inundation events. The most appropriate parameter for comparing field traces with numerical simulation results in this study was the average tsunami candidate layer thickness per vibracore. This parameter indicates the average sedimentary impact of a single tsunami event most adequately and can directly be compared to the amount of sediment accumulation or erosion computed for 18 different single tsunami wave scenarios in the numerical model (Chap. 5.2.4). Data of all three parameters was classified according to the Jenks optimisation method, which yields the minimum in-class variance but maximises the variances between all classes (JENKS 1967; JENKS & CASPALL 1971; MCMASTER 1997). This classification method turned out to be most appropriate with regard to field data of this study as it demonstrates natural breaks in the frequency distribution of the three chosen parameters.

5.2.3 Digital elevation model

The DEM used for tsunami simulations in this study covers the eastern Ionian Sea and the west coast of Greece, including the Ambrakian Gulf (Fig. 5.1). The domain stretches from 36.83° N to 40.00° N and from 19.00° E to 21.73° E ($\cong 233$ km $\times$ 352 km). The DEM was derived from several sources of data. The offshore topography of the open sea is based on the 30 arc-second ($= 926.1$ m) interval grid of GEBCO (IOC, IHO & BODC 2003). Bathymetric data of the shelf and the Ambrakian Gulf was digitised from several nautical charts on the scale of 1:93 213 to 1:300 000 (Imray Laurie Norie & Wilson Ltd 2011; Eagle Ray Publications 2010b; UKHO 2010). SRTM elevation data version 4 with a resolution of 90 m (JARVIS *et al.* 2008) and 30 m Google Earth elevation data (Google Inc. 2013; generated using the eSurvey Earth website <http://www.esurveyearth.com/>) were used to derive the onshore topography of the wider study area and alongshore the Ambrakian Gulf respectively. Among all freely available elevation data, both data sets are most appropriate for hydrodynamic simulations in the area because both closely accord with DGPS-based elevation data measured in the field and altitude markings in topographical maps. All elevation data was edited and assembled to an overall DEM with a maximum resolution of 30 m using ArcGIS Desktop 10.2 by Esri.

5.2.4 Hydro- and morphodynamic numerical model

In this study, a hydro- and morphodynamic numerical tsunami forward model was created using the Delft3D modelling suite (version 4.01.00.rc.12) developed by the Dutch Institute Deltares. Delft3D is a validated software suite to simulate both, off- and onshore tsunami propagation,

including sediment transport (APOTSOS *et al.* 2011a). It has been successfully employed for 1-D (APOTSOS *et al.* 2011d, 2012), 2-D (RÖBKE *et al.* 2012, 2013; CHACÓN-BARRANTES *et al.* 2013; PERÍÁÑEZ & ABRIL 2014a,b; Chaps. 3, 4) and 3-D tsunami simulations (GELFENBAUM *et al.* 2007; APOTSOS *et al.* 2009, 2011a,b,c).

All simulations were calculated using the program module Delft3D-FLOW, which solves the non-linear shallow water equations of unsteady flow and transport phenomena based on the full Navier Stokes equations for incompressible free surface flow (for this and the following see Deltare 2014, pp.177–180). The module is designed for flow phenomena where the horizontal spatial and temporal scales are much larger than the vertical scales, such as tidal waves, storm surges or tsunamis. In Delft3D-FLOW, the non-linear shallow water equations are solved in two (depth-averaged) or in three dimensions. As the water density in the oceans can approximately be regarded as vertically homogeneous, the depth-averaged calculation mode is appropriate for tsunami simulations conducted here.

As this study investigates tsunami propagation from the deep sea to the coastline and subsequent overland flow, the numerical model has to deal with energy dissipation due to wave-induced turbulence. Wave energy decay is particularly related to wave breaking and bottom roughness associated with the oscillatory water motion at the sea bottom. The production of turbulent energy due to both processes is considered in the numerical tsunami model based on appropriate extra source terms in the turbulent kinetic energy and dissipation equations of the basic turbulence models behind Delft3D-FLOW (for further methodological details see Deltare 2014, pp.236, 237; cf. APOTSOS *et al.* 2011a, pp.2099, 2116, 2117).

This study only considers the pure hydro- and morphodynamics of tsunami waves, whereas trigger mechanisms and the process of wave generation were not simulated. Accordingly, the numerical model is based on various hydrodynamic boundary conditions describing tsunami waves with different propagation directions, wavelengths and wave amplitudes in order to derive wave scenarios that accord with sedimentary field data (cf. PRITCHARD & DICKINSON 2008; GELFENBAUM *et al.* 2007; APOTSOS *et al.* 2009, 2011a,b,c,d, 2012). This approach was taken because in most cases, no correlation could be drawn between palaeotsunami field traces and written tsunami accounts for the Mediterranean. Consequently, no information was available neither on related trigger mechanisms nor on tsunami wave characteristics. Furthermore, there is no single specific tsunami source with regard to the Ambrakian Gulf, but instead various potential trigger mechanisms exist. Such are nearby submarine earthquakes and landslides but also remote earthquakes and mass wastings in the central and eastern Mediterranean Sea (especially in the area of the Hellenic Trench) as well as volcanic eruptions of Mount Etna (Sicily, Italy) or in the Aegean Sea (RÖBKE *et al.* 2013, pp.74–77). Thus, the number of potential tsunami sources is too large to be considered in this study. Instead, artificial waves that are realistic in view of the (supra)regional physical geography were created.

The hydrodynamic boundary conditions of the numerical model were determined according to the isosceles leading-depression (LDN) *N*-wave by TADEPALLI & SYNOLAKIS (1994, 1996) (see Eq.2.14), which is a widely accepted waveform to describe tsunamis (for this paragraph see Chap.2.5). The LDN-wave is based on the observation that the sea often recedes before a tsunami wave crest strikes the coast. This particularly occurs when tsunamis are triggered near the coast and the waves have not enough time to evolve into leading-elevation *N*-waves or solitary waves. Consequently, LDN-waves are likely to appear in the Ambrakian Gulf. LDN-waves with two different wave periods or wavelengths respectively and three different wave amplitudes were created, these wave parameters having the strongest influence on the onshore tsunami response. Although tsunamis usually appear as wave trains (BRYANT 2008, pp.27, 28; BORMANN 2012, pp.5, 6; Chap.2.2), all LDN-waves were simulated as single waves in order

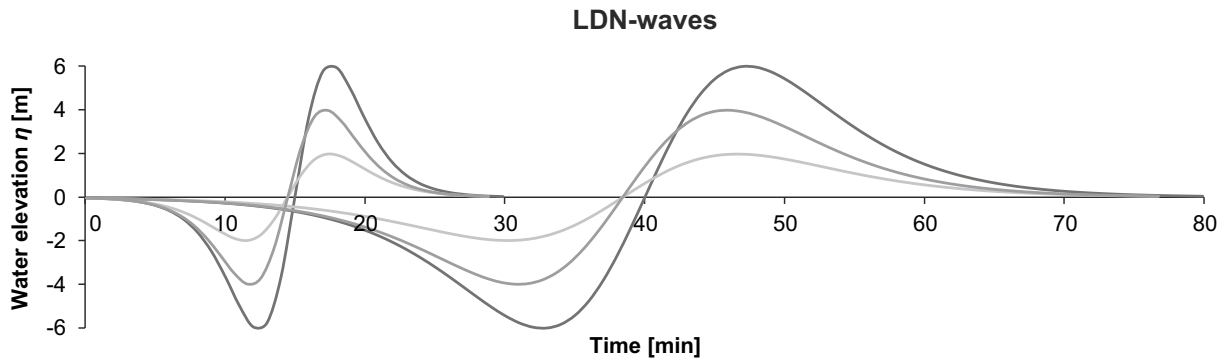


Figure 5.3: Leading-depression N -waves according to Eq. 2.14 with wave amplitudes of 2 m, 4 m and 6 m and wave periods of 30 min and 80 min. These waves form the hydrodynamic boundary conditions of the numerical tsunami model of this study at the southern, south-western and western model boundary. The fixed wave periods of 30 min and 80 min were realised according to the 1 %-definition by SYNOLAKIS *et al.* (2008, p. 2211). See text for further explanations.

to investigate their individual morphodynamic impact depending on the wave amplitude, wave period and propagation direction.

According to the energy peaks at periods between 30 min and 80 min measured for the Indian Ocean tsunami 2004 in the open sea (RABINOVICH *et al.* 2011a, p. 4; RABINOVICH & EBLÉ 2015, p. 23), we applied short and long LDN-waves with a wave period of 30 min (α values in Eq. 2.14 between 0.138 and 0.15) and 80 min (α values in Eq. 2.14 between 0.049 and 0.055) respectively (Fig. 5.3). Such wave periods are characteristic of tsunamis triggered by submarine landslides and earthquakes (BOLT *et al.* 1975, p. 134; HELAL & MEHANNA 2008, p. 788). Although N -waves are non-periodic waves and have an infinite wave period or wavelength respectively, a fixed wave period of 30 min and 80 min was realised according to the 1 %-wavelength-definition by SYNOLAKIS *et al.* (2008, p. 2211) (cf. CHACÓN-BARRANTES *et al.* 2013, p. 16). This definition says that the wavelength is defined as the distance between the wave's front and tail, where the water level is 1 % of the wave amplitude.

Based on the wave periods of 30 min and 80 min, wave amplitudes of 2 m, 4 m and 6 m were chosen (Fig. 5.3) in order to consider small, moderate and extreme tsunami waves according to the Imamura-Iida magnitude (IDA 1963; Tab. 2.1). Historical accounts (e.g. SOLOVIEV *et al.* 2000; SCHIELEIN *et al.* 2007), field data (e.g. PIRAZZOLI *et al.* 1999; BRYANT 2005, pp. 238–240; VÖTT *et al.* 2011b, 2015; HADLER *et al.* 2015a; WILLERSHÄUSER *et al.* 2015a,b), numerical simulations (e.g. PARESCHI *et al.* 2006a,c; SHAW *et al.* 2008; YOLSAL-ÇEVİKBİLEN & TAYMAZ 2012; PERÍAÑEZ & ABRIL 2014a; SAMARAS *et al.* 2015) and the regional physiogeographical setting (e.g. FERENTINOS 1991, 1992; PAPAACHOS 1996; POULOS *et al.* 1999; BENETATOS *et al.* 2004, 2005) suggest such tsunami wave amplitudes to be realistic in the central and eastern Mediterranean Sea (cf. RÖBKE *et al.* 2013, p. 74–77 and references therein).

Three tsunami propagation directions were considered in the numerical model, i.e. south, south-west and west, which are the favourite wave propagation directions with regard to the Ambrakian Gulf. The exact location of the southern, south-western and western model boundary from where the waves started are indicated in the overview map in Fig. 5.1. Combining the three propagation directions (S, SW, W) with two wave periods (30 min, 80 min) and three wave amplitudes (2 m, 4 m, 6 m) results in a total of 18 scenarios simulated in this study (Tab. 5.1).

All simulations are based on a two-part computational grid — an overall grid and a detailed, nested grid. The overall grid has a resolution of 900 m and covers the entire domain of the DEM (Chap. 5.2.3). The nested grid with a resolution of 90 m stretches from 38.72° N to 39.24° N and

Table 5.1: Overview of the 18 simulated tsunami scenarios, resulting from three propagation directions, two wave periods and three wave amplitudes.

LDN-wave from	Wave period	Wave amplitude	Scenario
South	30 min	2 m	S2S
		4 m	S4S
		6 m	S6S
	80 min	2 m	S2L
		4 m	S4L
		6 m	S6L
South-west	30 min	2 m	SW2S
		4 m	SW4S
		6 m	SW6S
	80 min	2 m	SW2L
		4 m	SW4L
		6 m	SW6L
West	30 min	2 m	W2S
		4 m	W4S
		6 m	W6S
	80 min	2 m	W2L
		4 m	W4L
		6 m	W6L

from 20.56° E to 21.20° E ($\cong 148 \text{ km} * 166 \text{ km}$). The boundary conditions of the nested model were determined by bilinear interpolation of the results computed at corresponding monitoring stations in the overall model (for further methodological details see Deltares 2014, pp. 677–683). For both the overall and nested model, the Riemann type of boundary was applied so that outgoing waves could cross the open boundary in order to minimise unintended wave reflections (VERBOOM & SLOB 1984; Deltares 2014, pp. 205–209). According to the CFL stability condition, the calculation step was 0.6 s in both, the overall and the nested model. All scenarios were computed for the time period characterised by significant changes in water level and significant sediment transport, which was about 10 h in each scenario.

Based on field surveys, photographs and Google Earth satellite images (Google Inc. 2013), appropriate values of Manning’s n coefficient were determined in order to take account of the bottom roughness in the model domain. n -values ranged from 0.015 (sea-floor, pastures) over 0.03 (moderate vegetated or developed land) to 0.07 (densely forested landscapes) (cf. HILLS & MADER 1997, p. 388; BRYANT 2008, pp. 45, 46).

With the aim of simulating tsunami morphodynamics, a realistic off- and onshore sediment stratigraphy had to be determined. This was only done for the detailed, nested model because no significant sediment transport was found for the overall model owing to the large water depths. Delft3D-FLOW considers both, suspended load and bedload transport of non-cohesive sediments as well as suspended load transport of cohesive sediments (for further methodological details see Deltares 2014, pp. 323–373; LESSER *et al.* (2004)). In Delft3D-FLOW, cohesive sediments are defined by a sediment diameter of less than 0.063 mm and are no further distinguished, meaning that silt and clay are treated as one grain size. In the case of non-cohesive sediments, however, any grain sizes between fine and coarse sand ($0.063 \text{ mm} \leq \varnothing < 2 \text{ mm}$) can be considered.

Based on these specifications, the sediment stratigraphy of the nested model was determined according to sedimentary field data of the study area. The 151 onshore vibracores taken within this study (Chap. 5.2.1) revealed that the thickness of unconsolidated sediment exceeds 10 m in

Table 5.2: Overview of the average on- and offshore sediment composition of the upper 10 m below surface/sea bottom for the nested model of the Ambrakian Gulf based on statistical analyses of 151 vibracores of the study area and sedimentary data taken from literature (see text for further explanations).

Grain size	Average onshore sediment composition of the upper 10 m below surface [%]	Average offshore sediment composition of the upper 10 m below sea bottom [%]
<i>Cohesive</i>		
Clay and silt	67	80
<i>Non-cohesive</i>		
Fine sand	9	20
Middle sand	7	0
Coarse sand	8	0
Gravel	8	0

most places. Regarding the offshore area, an even larger thickness has to be assumed, e.g. up to 45 m in the inner Ambrakian Gulf (KAPSIMALIS *et al.* 2005, p. 404). Nevertheless, for both, the onshore and offshore parts of the nested model, an initial sediment layer thickness of 10 m was chosen since the maximum amount of sediment erosion was less than 4 m in all simulated scenarios. Statistical analyses of all 151 vibracores alongshore the Ambrakian Gulf led to an average onshore sediment composition for the upper 10 m below surface of about 67 % clay and silt ($\varnothing < 0.063$ mm), 9 % fine sand ($\varnothing = 0.1315$ mm), 7 % middle sand ($\varnothing = 0.415$ mm), 8 % coarse sand ($\varnothing = 1.315$ mm) and 8 % gravel ($\varnothing \geq 2$ mm) (Tab. 5.2). Although the average grain size of gravel is 32.5 mm, a sediment diameter of 2 mm had to be chosen instead as this is the maximum grain size supported in Delft3D-FLOW. Judging by sediment cores, sediment samples and seismic profiles taken by TZIAVOS (1997), POULOS *et al.* (1995, 1999, 2005), KAPSIMALIS *et al.* (2005) and KARAGEORGIS *et al.* (2006) in the Ambrakian Gulf, its lagoons and in the adjacent eastern Ionian Sea, an average offshore sediment composition of 80 % clay and silt as well as 20 % fine sand was found for the upper 10 m below sea bottom (Tab. 5.2). Taking the average on- and offshore sediment fractions as a basis for the sediment stratigraphy in the nested model resulted in a 10 m thick initial sediment layer, in which all corresponding grain sizes are homogeneously mixed.

Cohesive sediment erosion and deposition in Delft3D-FLOW was computed according to the Partheniades-Krone formulations (PARTHENIADES 1965), whereas computation of non-cohesive sediment processes was based on the approach of VAN RIJN *et al.* (2001) (for further methodological details see Deltares 2014, pp. 323–373). The specific density and dry bed density of cohesive sediment was set to 2650 kg m^{-3} and 500 kg m^{-3} respectively, with a settling velocity of 0.25 mm s^{-1} . The critical bed shear stress was 1000 N m^{-2} for sedimentation and 0.5 N m^{-2} for erosion of cohesive sediment, and a corresponding erosion parameter of $0.0001 \text{ kg m}^{-2} \text{ s}^{-1}$ was chosen. Independent of the grain size, the specific density and dry bed density of non-cohesive sediment were set to 2650 kg m^{-3} and 1600 kg m^{-3} respectively.

As an important feature of tsunami inundation, the numerical model also considered the effect of sediment concentration on fluid density by adding the mass of all transported sediment per unit volume and subtracting the corresponding mass of displaced water (UNESCO 1981; cf. Deltares 2014, pp. 324, 325). Furthermore, the hindered sediment settling velocity due to other sediment particles in highly loaded water was considered according to RICHARDSON & ZAKI (1997) (cf. Deltares 2014, pp. 324, 325), based on a reference density for hindered settling of 1600 kg m^{-3} for all grain sizes. Finally, changes in the on- and offshore topography due to

sediment erosion and deposition were taken into account during simulation by updating the elevation of the bed in each grid cell after every calculation step (Deltares 2014, pp. 366–369).

5.3 Results

5.3.1 Distribution of tsunami deposits alongshore the Ambrakian Gulf

The analysis and evaluation of all 151 vibracores revealed a consistent geographical distribution of tsunami deposits alongshore the Ambrakian Gulf. All three determined parameters, i.e. the total number and the cumulative thickness of tsunami candidate layers per vibracore both divided by coring depth (yielding the relative number and relative cumulative thickness of event layers), as well as the average thickness of tsunami candidate layers per vibracore, indicate a significant decrease of tsunami influence in the Ambrakian Gulf from west to east. Regarding Fig. 5.4, the relative number of event strata, i.e. the number of tsunami candidate layers per coring metre per vibracore, ranges from 0 to >0.67 . High values (>0.29) are found around the entrance of the Ambrakian Gulf, including the Mazoma Lagoon, Aktio Headland, the Bay of Aghios Nikolaos and Lake Voulkaria, but significantly drop in the inner gulf with maximum 0.14 at its eastern shore. Thus, the number of tsunami impacts clearly decreases with increasing distance to the gulf's entrance. The same picture is found for the relative cumulative thickness of tsunami candidate layers with high values (>0.24) concentrated in the western gulf and low values (≤ 0.24 m) in the eastern part (Fig. 5.5). On a long-term basis, tsunamis therefore have resulted in higher sediment accumulation in the west compared to the east. This closely agrees with the spatial distribution found for the average tsunami candidate layer thickness, which ranges from 0 m to >1.94 m in the study area (Fig. 5.6). Again, the western Ambrakian Gulf shows the maximum values (frequently >0.76 m), whereas there are no values larger than 0.76 m in the inner and eastern gulf. Consequently, the average sediment accumulation per single tsunami event significantly decreases from the west to the east.

5.3.2 General inundation dynamics in the Ambrakian Gulf

Prior to the comparison of field data and simulated morphodynamics, a short summary of the basic computed tsunami inundation dynamics in the Ambrakian Gulf of all scenarios is given. Figure 5.7 exemplarily illustrates the inundation of the area in Scenario S4S. Due to the leading trough of the simulated LDN-wave, in each scenario, a significant withdrawal of the shoreline between Kastrosikiá and Lefkada City is observed before the wave crest makes landfall (cf. Fig. 5.7a; see Fig. 5.1 for places). Independent from the chosen boundary, all wave crests strike the Ambrakian Gulf from an almost western direction (cf. Fig. 5.7b). Whereas this is the expected direction for waves from the west, waves from the south-west and south are strongly diffracted and refracted around Lefkada Island, giving the impression that also these waves originate from the western model boundary. Once the water depth drops below 1000 m, strong wave shoaling occurs, which, together with funnel effects in the inlet-shaped Ionian Sea towards the Ambrakian Gulf (cf. Fig. 1.1), causes a significant increase in wave amplitude by up to factor three (cf. Fig. 5.7b). This results in maximum run-up heights of more than 40 m in the valleys between Kastrosikiá and Mitikas (Scenario W6L).

In all scenarios, tsunami waves completely inundate Aktio Headland (ground surface maximum 6 m a.s.l.) and subsequently enter the Ambrakian Gulf through the main waterway, the Strait of Preveza (cf. Figs. 5.7b,c). Simultaneously, tsunami waters reach the gulf via a small depression (ground surface maximum 14 m a.s.l.) to the west of the Mazoma Lagoon. Furthermore, each tsunami wave gets strongly funnelled in the Bay of Aghios Nikolaos and strikes

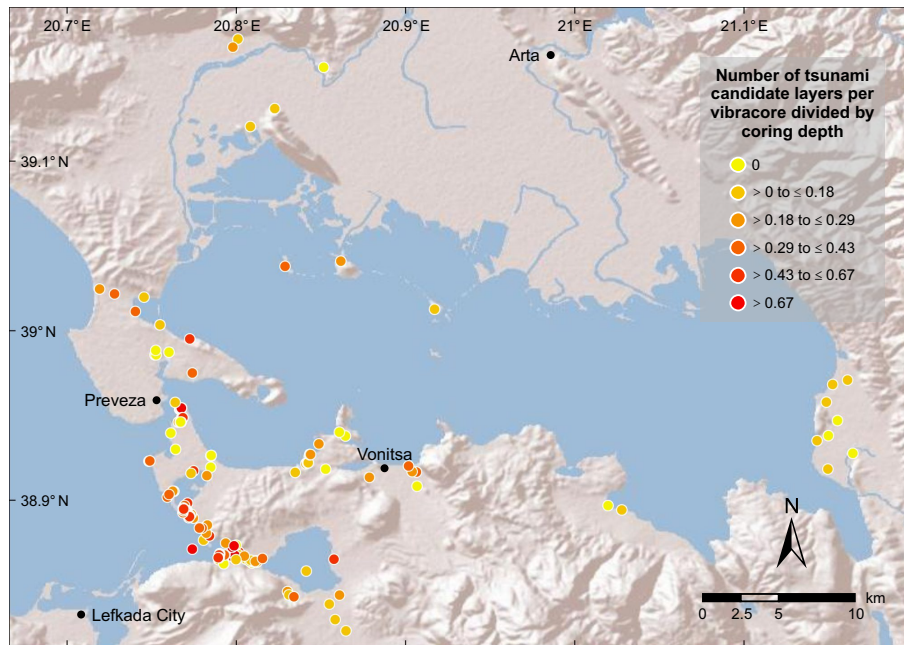


Figure 5.4: Geographical distribution of the total number of tsunami candidate layers per vibracore (ranging from 0 to 6) divided by coring depth (ranging from 0.72 m to 18 m) alongshore the Ambrakian Gulf. The identification of tsunami deposits in all 151 vibracores is based on several physical sedimentary criteria typical of tsunamites (Chap. 5.2.2 and Fig. 5.2; cf. Chap. 2.4). Furthermore, geochemical, micro-faunal and palynological analyses conducted for selected vibracores in previous studies helped to detect event layers.

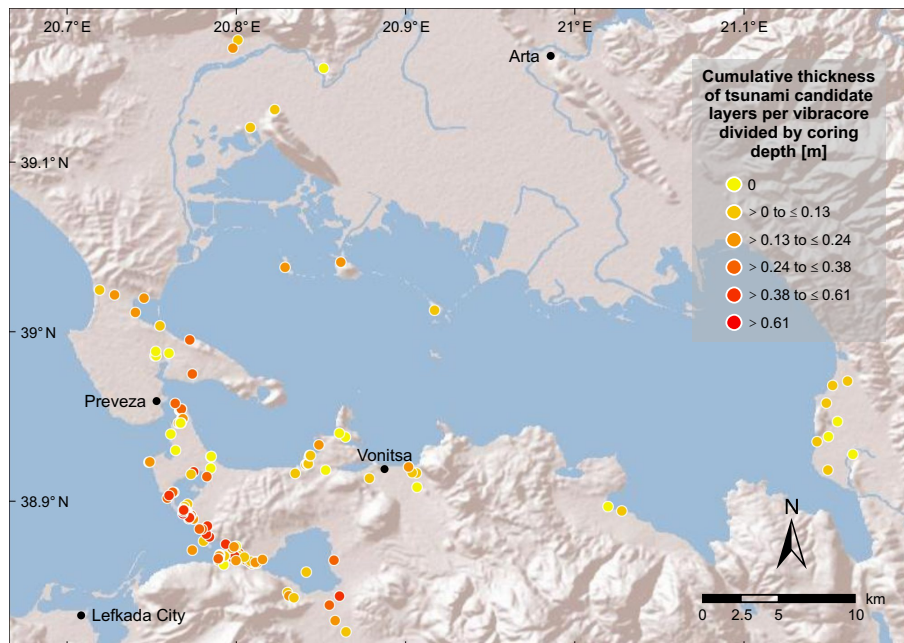


Figure 5.5: Geographical distribution of the cumulative thickness of tsunami candidate layers per vibracore (ranging from 0 m to 5.14 m) divided by coring depth (ranging from 0.72 m to 18 m) alongshore the Ambrakian Gulf. The identification of tsunami deposits in all 151 vibracores is based on several physical sedimentary criteria typical of tsunamites (Chap. 5.2.2 and Fig. 5.2; cf. Chap. 2.4). Furthermore, geochemical, microfaunal and palynological analyses conducted for selected vibracores in previous studies helped to detect event layers.

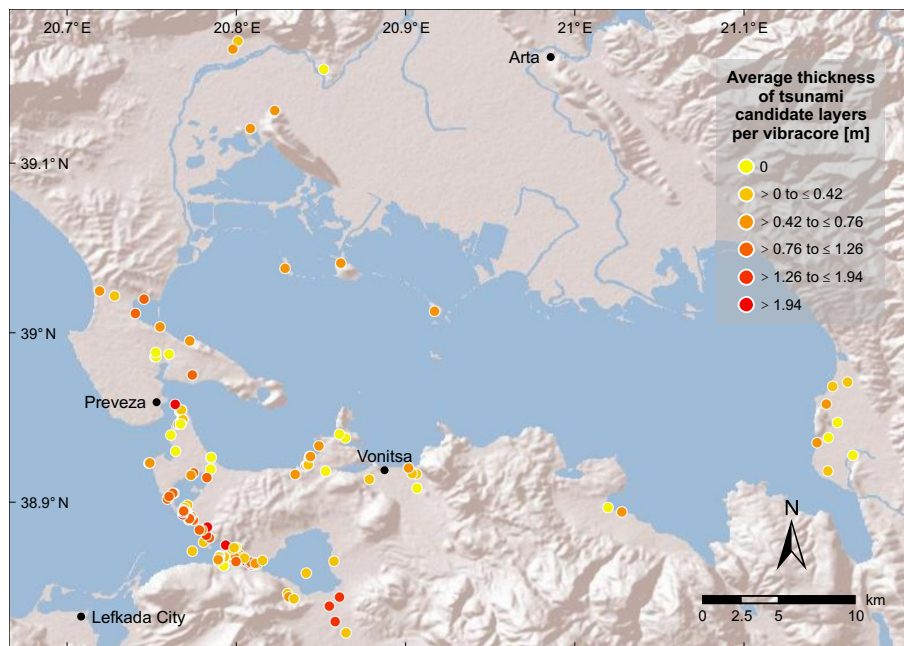


Figure 5.6: Geographical distribution of the average thickness of tsunami candidate layers per vibracore alongshore the Ambrakian Gulf. The identification of tsunami deposits in all 151 vibracores is based on several physical sedimentary criteria typical of tsunamites (Chap. 5.2.2 and Fig. 5.2; cf. Chap. 2.4). Furthermore, geochemical, microfaunal and palynological analyses conducted for selected vibracores in previous studies helped to detect event layers. The average tsunami layer thickness is compared with the computed sedimentation in the numerical model in Chap. 5.3.3.

Lake Voulkaria. Once the waves have reached the inner gulf, the area around Vonitsa and the narrow sandspit systems as well as the wide coastal flats in the northern gulf get inundated (cf. Figs. 5.7c,d). Due to flow divergence, the eastern gulf is hardly affected by tsunami waves; considerable flooding only occurs in the moderate and extreme scenarios (wave amplitudes of 4 m and 6 m) in the Boukka coastal plain. Even after a simulation time of 10 h, no significant backflow can be observed. Instead, tsunami waters gradually flow back into the Ionian Sea mainly via the Strait of Preveza.

Generally, tsunamis from the west cause the strongest inundation of the Ambrakian Gulf, whereas waves from the south have the weakest impact (here and in the following measured by the inundation depth, flow velocity and area of inundation). Tsunamis originating from the south-western model boundary result in an intermediate onshore response. Besides the propagation direction, the onshore tsunami response strongly depends on the wave period and wave amplitude; for both wave parameters, a positive correlation with the inundation intensity is found. As a result, Scenario W6L—representing a tsunami wave from the western model boundary with a wave amplitude of 6 m and a wave period of 80 min—shows the strongest onshore response in the Ambrakian Gulf, whereas Scenario S2S shows the weakest impact.

5.3.3 Sedimentary field traces versus computed morphodynamics

In the following, the results of the simulated tsunami morphodynamics are described and compared with tsunami field traces presented in Chap. 5.3.1. Maps of the Ambrakian Gulf show the computed sediment erosion (blue colours) and accumulation (yellow to red colours) associated with tsunami inundation after a simulation time of 10 h in each scenario (Figs. 5.8–5.13; notice

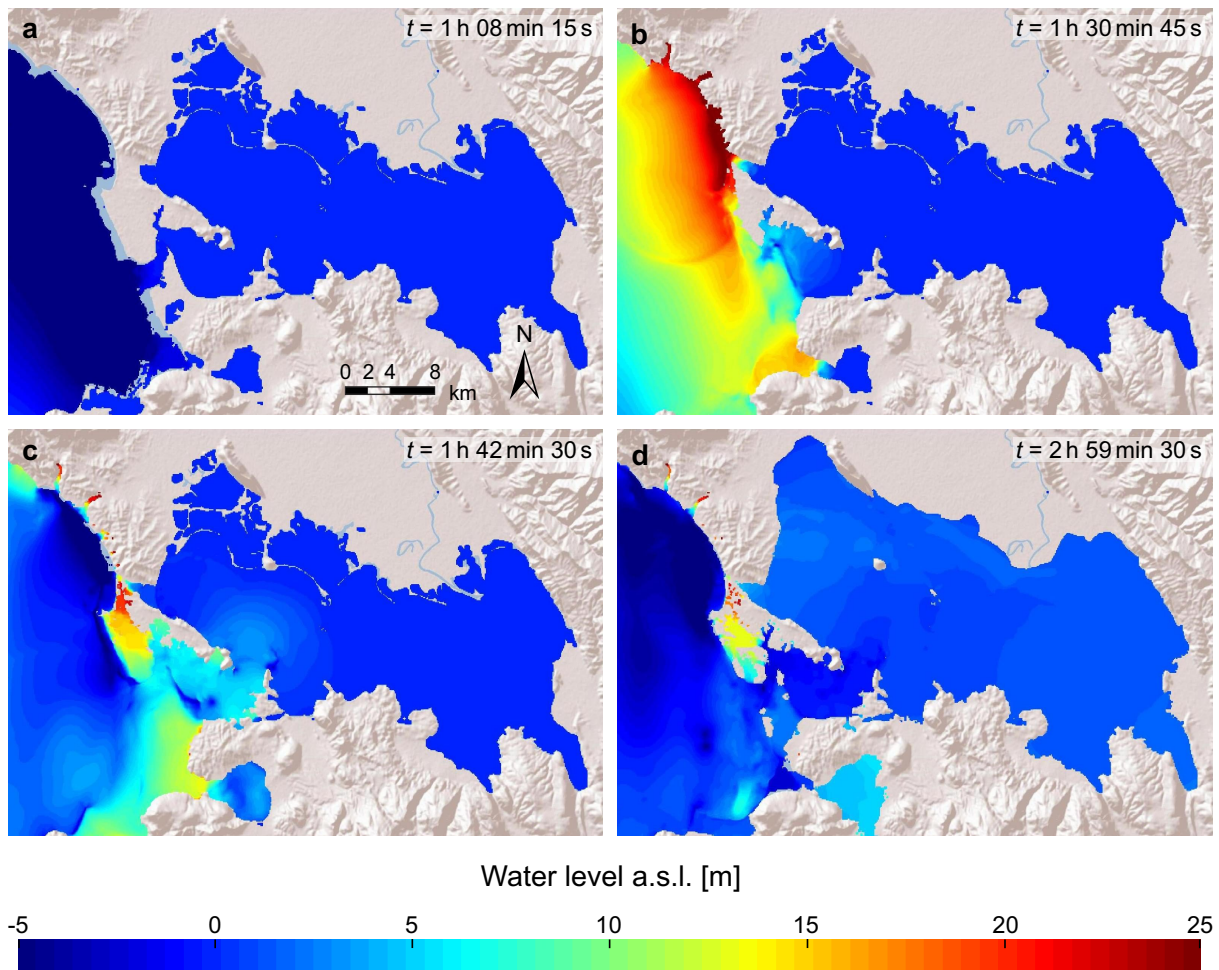


Figure 5.7: Snapshots documenting stepwise the inundation of the Ambrakian Gulf in Scenario S4S, which is based on a single tsunami wave from a southern direction with a wave amplitude of 4 m and a wave period of 30 min. (a) shows the receding shoreline due to the leading wave trough of the simulated LDN-wave. The following wave crest (b) causes maximum run-up heights of up to 25 m in the valleys between Kastrosikiá and Mitikas, complete inundation of Aktio Headland and water levels of more than 15 m in the Bay of Aghios Andreas. In (c), the tsunami wave enters both, the Ambrakian Gulf via Aktio Headland and Lake Voulkaria. (d) illustrates the maximum inundation extent around Lake Voulkaria and in the northern gulf, including the sandspit systems and large delta areas.

the varying scale bar). This period covers most of the inundation process, including the main backflow. The maps further contain information on the average thickness of tsunami candidate layers per vibracore already illustrated in Fig. 5.6. This parameter allows a direct comparison with the sedimentary impact computed for different single tsunami waves (Chap. 5.2.2). Due to the large number of computed scenarios (18), only those scenarios are presented that show the best match with field data, namely Scenario W2S, S4S, W4S, W4L, S6S and W6L.

In all scenarios based on a low wave amplitude of 2 m, the sedimentary onshore response of tsunami waves is mainly restricted to Aktio Headland, the Bay of Aghios Nikolaos and the entrance of Lake Voulkaria. Only in Scenario W2L, significant sediment erosion and accumulation is also observed in the inner Lake Voulkaria and to the east of Aktio Headland. The best agreement with vibracore data of the Ambrakian Gulf is found for the simulated western tsunami with a short wave period of 30 min (Scenario W2S; Fig. 5.8). Although the absolute

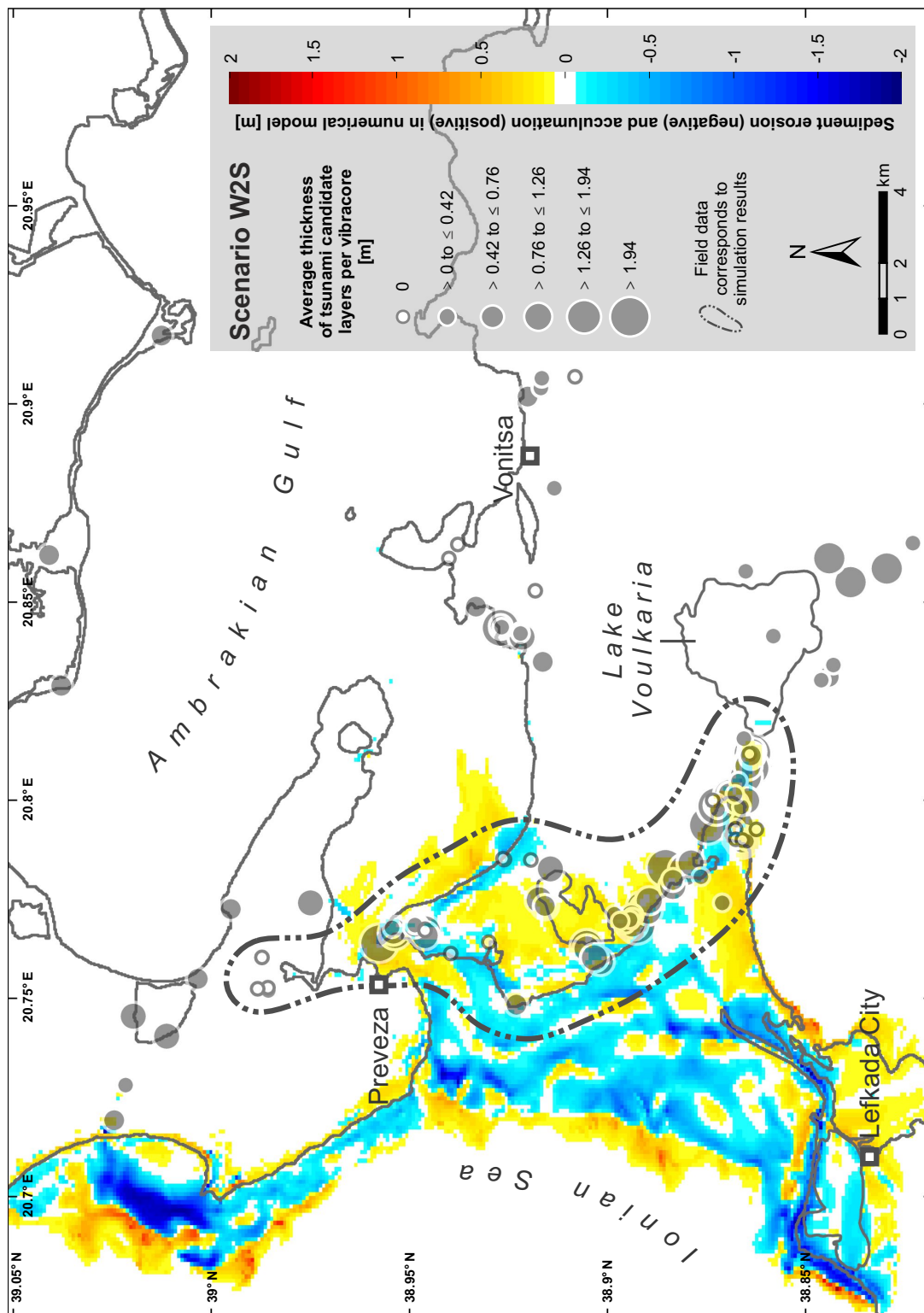


Figure 5.8: Comparison of the average thickness of tsunami candidate layers per vibracore (cf. Fig. 5.6) with computed sediment erosion and accumulation in the numerical model in Scenario W2S for the Ambrakian Gulf after a total simulation time of 10 h. Scenario W2S is based on a single tsunami wave from a western direction with a wave amplitude of 2 m and a wave period of 30 min. The dashed-dotted line indicates the area in which the average tsunamite thickness detected in the field and the computed sedimentation of tsunami inundation are generally consistent.

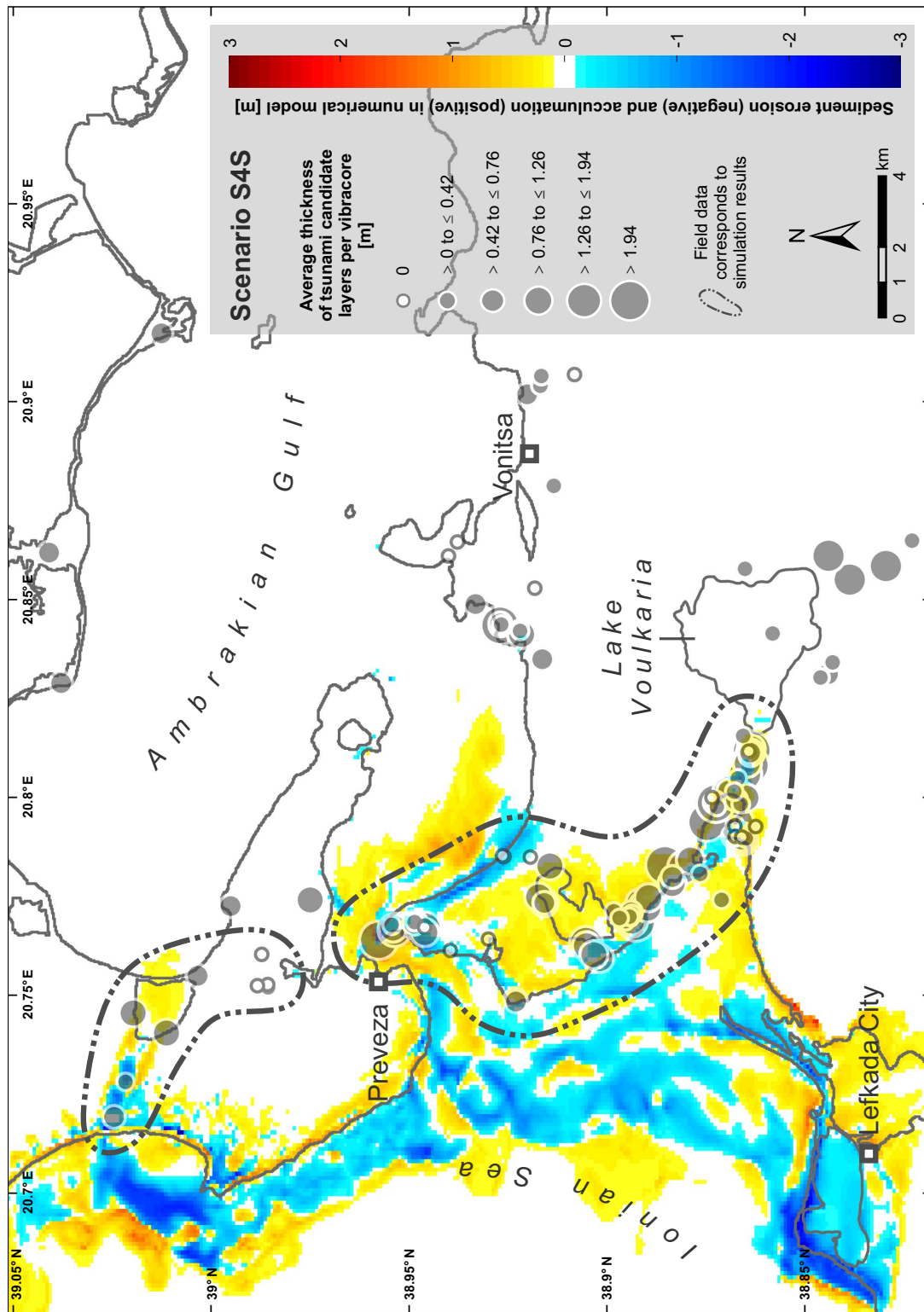


Figure 5.9: Comparison of the average thickness of tsunami candidate layers per vibracore (cf. Fig. 5.6) with computed sediment erosion and accumulation in the numerical model in Scenario S4S for the Ambrakian Gulf after a total simulation time of 10 h. Scenario S4S is based on a single tsunami wave from a southern direction with a wave amplitude of 4 m and a wave period of 30 min. The dashed-dotted line indicates the area in which the average tsunamite thickness detected in the field and the computed sedimentation of tsunami inundation are generally consistent. See Fig. 5.7 for the computed hydrodynamics of this scenario.

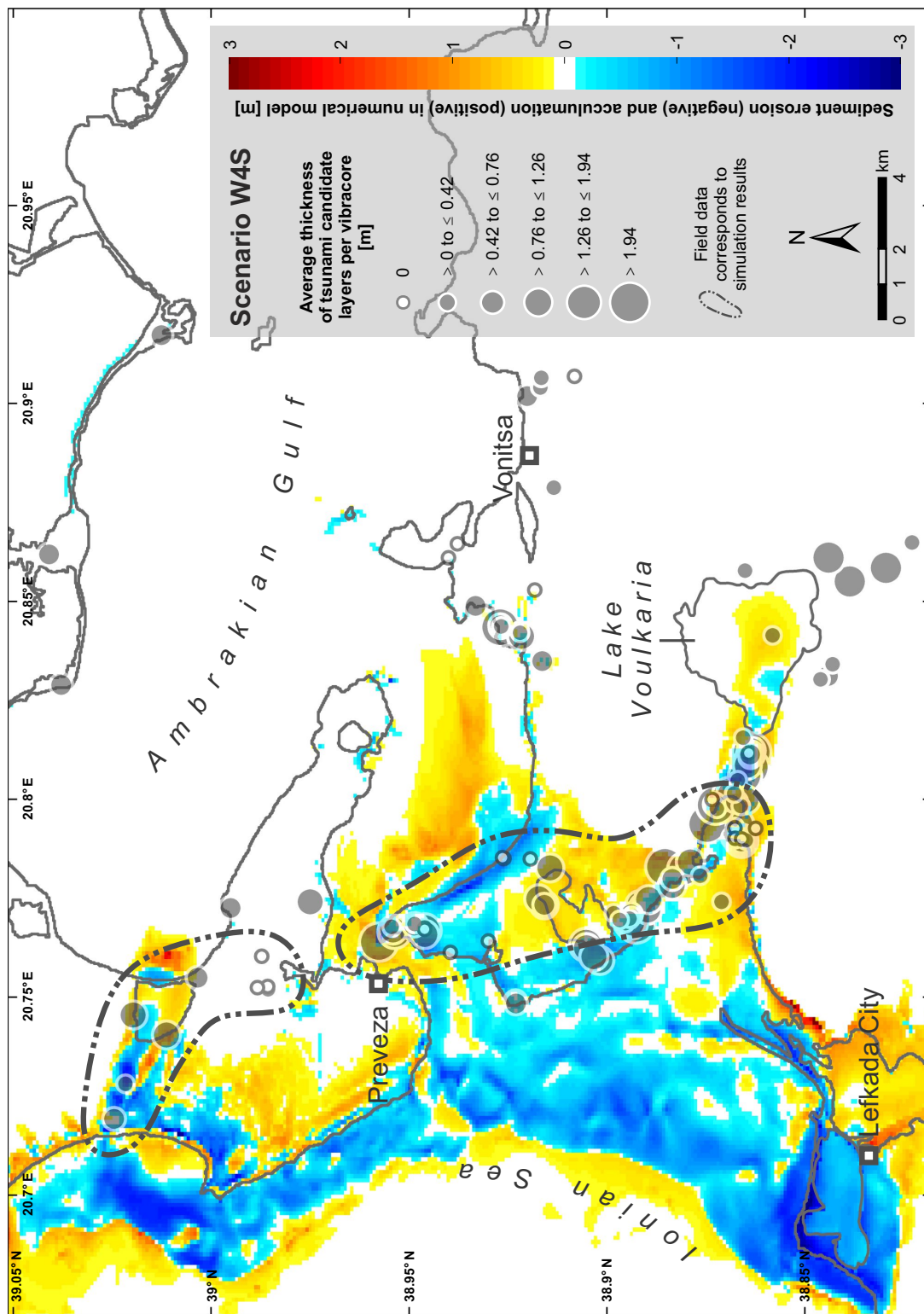


Figure 5.10: Comparison of the average thickness of tsunami candidate layers per vibracore (cf. Fig. 5.6) with computed sediment erosion and accumulation in the numerical model in Scenario W4S for the Ambrakian Gulf after a total simulation time of 10 h. Scenario W4S is based on a single tsunami wave from a western direction with a wave amplitude of 4 m and a wave period of 30 min. The dashed-dotted line indicates the area in which the average tsunamite thickness detected in the field and the computed sedimentation of tsunami inundation are generally consistent.

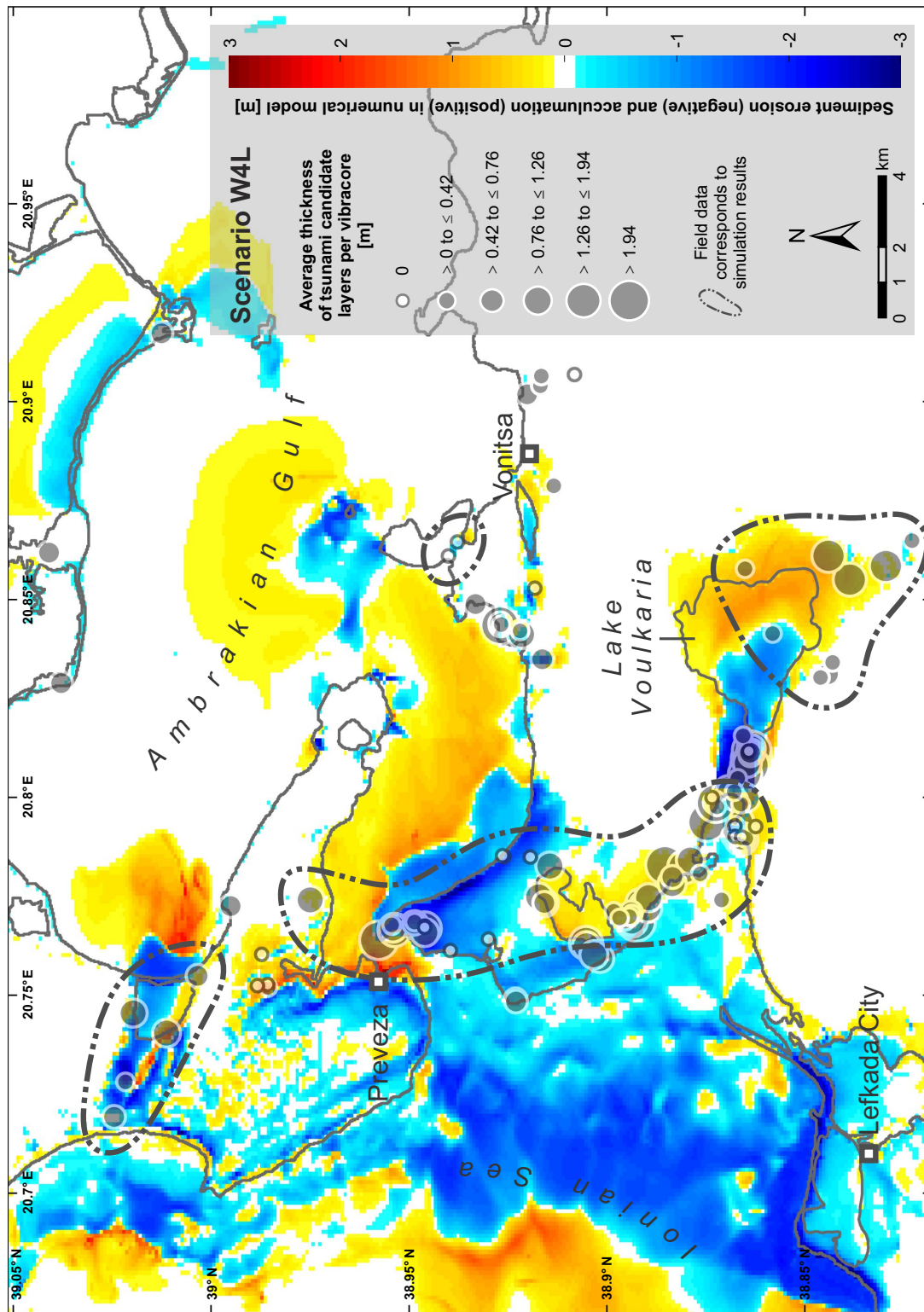


Figure 5.11: Comparison of the average thickness of tsunami candidate layers per vibracore (cf. Fig. 5.6) with computed sediment erosion and accumulation in the numerical model in Scenario W4L for the Ambrakian Gulf after a total simulation time of 10 h. Scenario W4L is based on a single tsunami wave from a western direction with a wave amplitude of 4 m and a wave period of 80 min. The dashed-dotted line indicates the area in which the average tsunamite thickness detected in the field and the computed sedimentation of tsunami inundation are generally consistent.

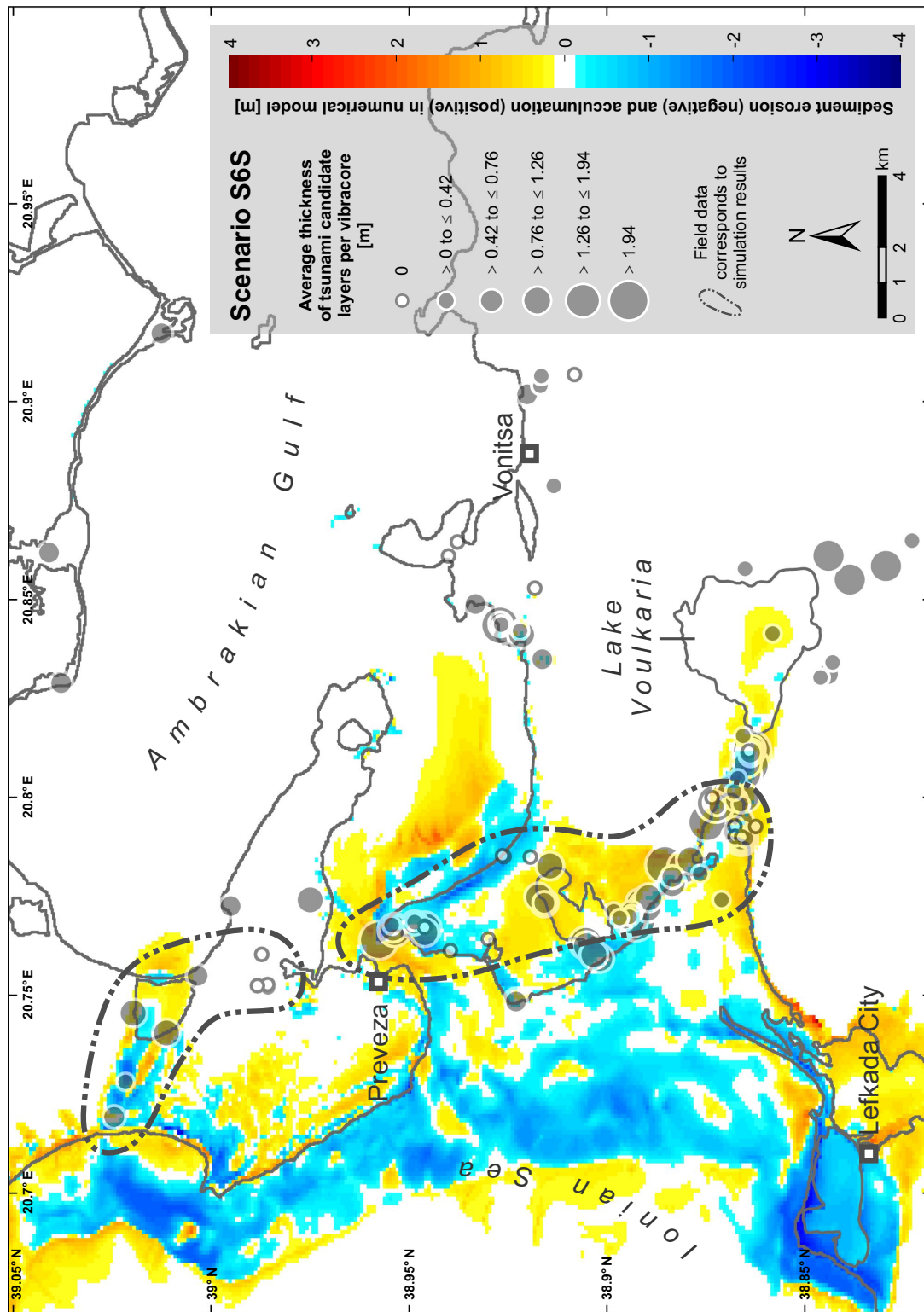


Figure 5.12: Comparison of the average thickness of tsunami candidate layers per vibracore (cf. Fig. 5.6) with computed sediment erosion and accumulation in the numerical model in Scenario S6S for the Ambrakian Gulf after a total simulation time of 10 h. Scenario S6S is based on a single tsunami wave from a southern direction with a wave amplitude of 6 m and a wave period of 30 min. The dashed-dotted line indicates the area in which the average tsunamite thickness detected in the field and the computed sedimentation of tsunami inundation are generally consistent.

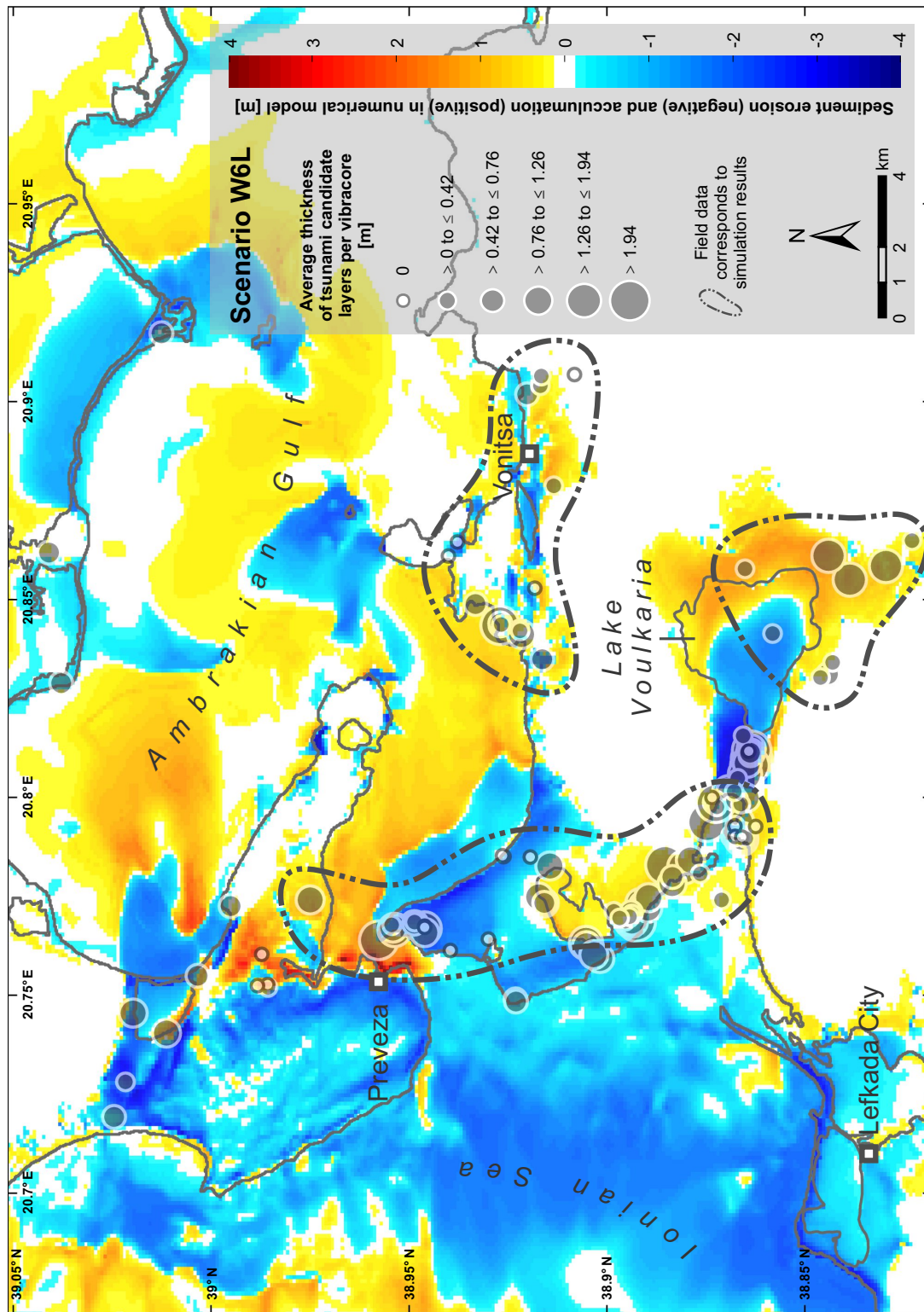


Figure 5.13: Comparison of the average thickness of tsunami candidate layers per vibracore (cf. Fig. 5.6) with computed sediment erosion and accumulation in the numerical model in Scenario W6L for the Ambrakian Gulf after a total simulation time of 10 h. Scenario W6L is based on a single tsunami wave from a western direction with a wave amplitude of 6 m and a wave period of 80 min. The dashed-dotted line indicates the area in which the average tsunamite thickness detected in the field and the computed sedimentation of tsunami inundation are generally consistent.

sedimentation computed in this scenario (≤ 0.5 m) clearly lags behind the local average tsunami candidate layer thickness per vibracore (≤ 2.67 m), both show the same spatial extent. Areas of computed sediment erosion or no sediment accumulation in north-western and north-eastern Aktio Headland as well as to the north of Preveza coincide with local vibracore data showing no event layers. Sediment accumulation in the numerical model is mostly found at the northern tip and western shore of Aktio Headland, around the Saltini Lagoon, along the Bay of Aghios Andreas and at the entrance of Lake Voulkaria; for these areas also vibracore data indicates dominant tsunami sediment deposition. Regarding the offshore tsunami morphodynamics, in all scenarios based on a 2 m wave amplitude, significant sediment processes are not observed before the water depth drops below 60 m. Consequently, most of the deposited tsunami sediments originate from the shallow marine to littoral zone landward from the -60 m contour.

Although Scenario W2S is found in close accordance with field data of Aktio Headland and the Bay of Aghios Nikolaos, Scenario S4S is the favourite scenario with regard to the entire western part of the Ambrakian Gulf, including the Mazoma Lagoon (Figs. 5.7, 5.9). Again, vibracores containing no tsunami candidate layers are generally located in areas that show dominant erosion or no accumulation in the numerical model, whereas vibracores with thick event strata are mostly found in areas of computed sediment accumulation. Although still too small compared to the average layer thickness determined per vibracore, the computed thickness of tsunami deposits in Scenario S4S has significantly increased (up to 1 m).

According to the stronger hydrodynamic onshore response of western wave scenarios (Chap. 5.3.2), in Scenario W4S (Fig. 5.10), sediment erosion and accumulation affects a larger area and shows higher values compared to Scenario S4S. As a result, the computed sediment thickness (up to 1.5 m) nearly matches the average tsunami candidate layer thickness per vibracore, where both, simulation results and vibracores indicate significant sediment accumulation. Areas showing consistent model and field data are similar to those in Scenario S4S, with the exception of western Aktio Headland and the entrance of Lake Voulkaria. Here, numerical results suggest dominant sediment erosion, which is in contrast to relatively thick event layers found in the local vibracores. In return, Scenario W4S is the first scenario that explains sedimentation in the midst of Lake Voulkaria.

The morphodynamics computed for Scenario W4L (Fig. 5.11) are far stronger compared to the scenarios described above. Whereas the western part of the Ambrakian Gulf (except for the Saltini Lagoon and the Bay of Aghios Nikolaos) mostly show sediment erosion, significant sedimentation is particularly observed to the east of the Mazoma Lagoon and Aktio Headland, in the inner gulf and along the eastern shore of Lake Voulkaria. For the latter, the computed sediment thickness in the numerical model (up to 1.10 m) accurately replicates the average event layer thickness determined for corresponding vibracores. Despite the clear sedimentary impact on the inner gulf, Scenario W4L is not appropriate to explain vibracore data around Vonitsa and the sandspit systems in the north. Here, the numerical model either suggests no significant sedimentation or even dominant erosion respectively. Strong morphodynamics in Scenario W4L are also observed offshore the Ambrakian Gulf, where sediment processes already start clearly seaward from the -60 m contour.

Scenario S6S (Fig. 5.12) shows a similar spatial distribution of computed sediment erosion and accumulation, including the absolute sediment thickness, as was described for Scenario W4S. Sediment processes in Scenario W6L are on an unprecedented scale starting already landward from the -100 m contour and reaching far into the inner and even eastern Ambrakian Gulf (Fig. 5.13). Comparable to Scenario W4L, in Scenario W6L, the western part of the Ambrakian Gulf is mostly characterised by sediment erosion, which is generally not consistent with vibracore data. Instead, a good accordance between simulated tsunami morphodynamics and field traces

is found for the eastern and southern shore of Lake Voulkaria and for the inner Ambrakian Gulf around Vonitsa. For the latter, Scenario W6L is the only scenario that plausibly explains tsunami event strata detected in the field. However, simulation results again do not match vibracore data of the sandspit systems, the delta plain in the north and the Boukka coastal plain at the very eastern end of the gulf (the latter two locations are not displayed in Figs. 5.8–5.13; see Fig. 5.6).

5.4 Discussion

Based on the presented simulation results, a number of systematic hydro- and morphodynamic features of tsunami inundation in the Ambrakian Gulf can be drawn together:

- Tsunamis from a western direction show the strongest hydro- and morphodynamic impact on the Ambrakian Gulf. This is closely related to strong funnelling of tsunami waters in the eastern Ionian Sea between the islands of Corfu and Cephalonia for waves from the west. In contrast, funnel effects are less distinct for waves from a south-western and especially southern direction (cf. VÖTT *et al.* 2008, pp. 107, 119). Furthermore, the central Ionian Islands, in particular Lefkada, Cephalonia and Zakynthos, seem to give some protection for the Ambrakian Gulf against waves from these two directions.
- The sedimentary response of all short-period tsunami waves ($T = 30$ min) is concentrated on a comparatively narrow area mainly covering the western part of the Ambrakian Gulf (Figs. 5.8–5.10, 5.12). These waves result in high sediment accumulation especially between the Mazoma Lagoon and Lake Voulkaria. By contrast, for all long-period waves ($T = 80$ min), significant sediment processes are observed for a much wider area stretching from the open Ionian Sea (landward from the -100 m contour) up to the eastern shore of the Ambrakian Gulf (Figs. 5.11, 5.13). Associated with the longer reach of long-period tsunami waves, here, the western part of gulf, including the area of the Mazoma Lagoon, Aktio Headland and the entrance of Lake Voulkaria, is dominated by erosion and the main sedimentation is found further east or clearly offshore in the Ionian Sea. Nevertheless, maximum values of erosion and accumulation are only slightly larger for long- than for short-period waves, the same wave amplitude provided.
- Computed sediment accumulation — where not influenced by natural obstacles (e.g. cliffs or hills) — is clearly characterised by thinning-landward in all scenarios. This is due to the general decrease of inundation depth and flow velocity and by this of the transport capacity of inundating tsunami waters. Thinning-landward is most obvious in Scenario W4L (Fig. 5.11) and W6L (Fig. 5.13) along the eastern shore of Lake Voulkaria and in Scenario W6L also in the area of Vonitsa. The spatial analysis of the various sediment fractions in the numerical model further reveals significant fining-landward of tsunami deposits. Finally, in particular in the moderate and extreme scenarios ($A \geq 4$ m), significant erosion is observed prior to the deposition of coarse-grained material, which results in erosional contacts at the base of such sediments. All three, thinning- and fining-landward as well as basal erosional unconformities are essential features typical of tsunami deposits (Chap. 5.2.2).
- The main sedimentary response of tsunamis in the Ambrakian Gulf arises from the wave run-up and not from the backflow. This is connected to the extensive flat flooding zone in the northern Ambrakian Gulf and therefore the gentle slope of the water surface, which prevents high flow velocities during the backflow.

Generally, the numerical model accurately reflects the decreasing tsunami influence from the western to the eastern parts of the Ambrakian Gulf, as is indicated by sedimentary field data (Chap. 5.3.1). However, there is no specific simulated scenario which can totally explain the spatial distribution of detected tsunami deposits all along the gulf at once. Instead, various scenarios are needed to reproduce tsunami field traces in different parts of the study area. This is also suggested by the field data itself because most vibracores contain several tsunami candidate layers that point to different tsunami events in the past (VÖTT *et al.* 2006, 2007, 2008, 2009b, 2010a, 2011a; MAY *et al.* 2007, 2012a; Chap. 5.1). According to these studies and regarding the plethora of possible tsunami trigger mechanisms in the central and eastern Mediterranean (Chap. 5.2.4; RÖBKE *et al.* 2013, pp. 74–77), these tsunami impacts were probably related to various sources and therefore have resulted in different sediment accumulation in the study area. Consequently, the average thickness of tsunami candidate layers per vibracore used for the comparison with simulation results does not reflect the actual sedimentary response of a particular tsunami event and therefore cannot be accurately replicated by single tsunami waves simulated in the numerical model.

Nevertheless, the simulation results suggest that short-period tsunami waves ($T = 30$ min) from the south or west with any wave amplitude are in favour of field traces of the western Ambrakian Gulf, including the area of the Mazoma Lagoon, Aktio Headland, the Bay of Aghios Andreas and the entrance to Lake Voulkaria (Figs. 5.8–5.10, 5.12). In contrast, 4-m- to 6-m-amplitude long-period waves ($T = 80$ min) especially from a western direction, are the favourite tsunami scenarios in respect to sedimentary field data around Lake Voulkaria and Vonitsa in the inner gulf (Figs. 5.11, 5.13). The sedimentary response of tsunami waves from the south-west, however, generally do not match the spatial distribution of tsunami deposits detected in the field.

The marine to shallow marine origin of detected tsunami deposits in the study area can be explained by all simulated scenarios, which show that most sediment in the numerical model originate from the shallow marine to littoral zone landward from the -60 m contour. These findings corroborate published field and numerical findings of previous studies (e.g. MORTON *et al.* 2007, pp. 187–194; APOTSOS *et al.* 2011b, p. 8; APOTSOS *et al.* 2011c, p. 9; SUGAWARA *et al.* 2014a, p. 299; SUGAWARA *et al.* 2014b, p. 19). Related tsunami event strata at adjacent vibracoring sites in the study area frequently show thinning- and fining-landward tendencies (VÖTT *et al.* 2006, 2007, 2008, 2009b, 2010a, 2011a; MAY *et al.* 2007, 2012a), which are also recorded in the simulated scenarios. Furthermore, field observations of erosional unconformities at the base of most tsunami candidate layers can be explained by previous erosion and subsequent sediment accumulation widely observed in the numerical model.

Besides these general common features of simulation results and field traces, the numerical model also reproduces local tsunami inundation dynamics for particular key sites in the western part of the Ambrakian Gulf which were reconstructed on the basis of vibracore data in previous studies:

- According to VÖTT *et al.* (2008, p. 120), the central and northern part of the former Plaka sandspit system, located to the south-west of the Bay of Aghios Nikolaos (Fig. 5.1), was completely destroyed by tsunami waves over a distance of more than 5 km. This closely agrees with all simulated scenarios showing dominant erosion in this area.
- Computed morphodynamics corroborate strong local erosional forces during tsunami land-fall in the western, northern and eastern part of Aktio Headland, as documented by distinct erosional contacts in vibracores (VÖTT *et al.* 2007, p. 54).
- VÖTT *et al.* (2007) and VÖTT *et al.* (2011a, pp. 236, 237) suggest that tsunami waters

struck the Bay of Aghios Nikolaos, inundated Lake Voulkaria and subsequently flooded the Bay of Palairos located almost 5 km to the south. Inundation of Lake Voulkaria from a western direction is reproduced by all simulated scenarios. Scenario W4L and W6L (Figs. 5.11, 5.13) further show inundation and significant morphodynamics in the adjacent Bay of Palairos.

- According to VÖTT *et al.* (2009b, p. 30), the so-called Cleopatra Canal — connecting the Bay of Aghios Nikolaos with Lake Voulkaria — was originally dug out by tsunami waves and was filled again during a subsequent tsunami event. Erosion of the canal can easily be explained by Scenario W4S, W4L, S6S and W6L (Figs. 5.10–5.13), whereas the sediment infill might be reflected by Scenario W2S and S4S (Figs. 5.8, 5.9).

With some exceptions, it is rather the spatial extent of deposits that agree with vibracore data than the absolute thickness of computed sedimentation. The latter usually lags behind the average tsunami candidate layer thickness determined per vibracore, except for the strongest scenarios (W4L, S6S, W6L). This may be related to a number of reasons:

- In the case of several thin and only one or two exceptional thick tsunami candidate layers per vibracore, the average layer thickness would pretend a thickness too large compared to most tsunami layers in the vibracore, which is therefore not reproduced in most simulated scenarios.
- If the upper part of an event layer has been eroded and subsequently covered by younger tsunami deposits, the complete sequence may be interpreted as only one tsunami candidate layer resulting in an overestimation of the average layer thickness per vibracore. This may further explain the exceptionally large average layer thickness of more than 1 m found for some vibracores especially in the western part of the Ambrakian Gulf.
- The absolute thickness of computed sedimentation is affected by various parameters in the numerical model, including the predetermined bottom roughness, values of the critical bed shear stress, sediment stratigraphy as well as the small-scale topography (cf. APOTSOS *et al.* 2009; APOTSOS *et al.* 2011b, pp. 4, 16; APOTSOS *et al.* 2011c, p. 9; LI *et al.* 2012a, 2014). For example, a general overestimation of the critical bed shear stress would result in less sediment entrainment by waves, and consequently in less sediment deposition in the numerical model than observed in reality.

In contrast to the absolute thickness, the spatial extent of tsunami deposits in both, the numerical model and the field is more resistant to modifications. This is because even a very thin event strata would still be indicative of tsunami influence. Therefore, the spatial extent of tsunamites is generally more appropriate for the comparison of simulation results and field data than the absolute thickness (cf. LI *et al.* 2014, p. 2272).

Erosion particularly due to subsequent tsunami impacts does not only strongly affect the absolute thickness but also the general spatial distribution of tsunami deposits in the field (e.g. WHEATCROFT 1990; WHEATCROFT & DRAKE 2003; SZCZUCIŃSKI 2012; SPISKE *et al.* 2013). In extreme cases, tsunami event layers can be completely removed from the geological record, which is an important process to be considered in this study. Especially the area of the Mazoma Lagoon, Aktio Headland and the entrance of Lake Voulkaria show dominant erosion in some simulated scenarios, which is also documented by distinct erosional unconformities in the local vibracores (cf. VÖTT *et al.* 2007, p. 54). Particularly in such cases, the number as well as the cumulative and average thickness of tsunami candidate layers must be regarded as minimum values.

All simulated scenarios are based on single tsunami waves in order to investigate the morphodynamic effects of different wave periods, wave amplitudes and propagation directions. However, most tsunamis appear in the form of wave trains consisting of a series of waves (Chap. 2.2). Subsequent waves during one and the same tsunami event may redeposit sediment that was released by previous waves (cf. e.g. APOTSOS *et al.* 2011b, p. 15; SUGAWARA *et al.* 2014a, pp. 298, 299), which would result in a sediment distribution not replicated by any of the simulated scenarios in this study.

Given the wide time span of detected tsunami impacts and associated deposits in the Ambrakian Gulf of about 8000 years, long-term changes in the local topography as described in Chap. 5.1 may further account for discrepancies between simulation results and field data (cf. Chaps. 3, 4). This is particularly true for the northern and eastern shores of the Ambrakian Gulf, which experienced major topographical changes due to tectonic and fluvio-deltaic processes and where no scenarios can explain tsunami deposits detected in the field. However, this study takes account of the average sedimentary signal of all detected tsunami impacts in the last 8000 years and does not consider deposits associated with a particular single event in pre-/history due to an incomplete age model of the sediments (Chap. 5.2.2).

Finally, the comparison of morphodynamic simulation results with sedimentary field data of the study area is limited by the spatial resolution of the numerical model. Whereas the simulations yield the average sedimentation for 90 m by 90 m grid cells, vibracores represent the geological record for a specific coring site only. This may result in further inconsistency between model and field (cf. SUGAWARA *et al.* 2014b, p. 28).

5.5 Conclusions

According to the general decrease of tsunami wave energy during the landfall (Chap. 2.2), sedimentary vibracore data indicates a decreasing tsunami influence from the western to the eastern Ambrakian Gulf for both, the frequency and the intensity of events in the last 8000 years. This is accurately reflected by the hydro- and morphodynamic numerical model showing that the sedimentary tsunami response decreases from the west to the east and that the inner and eastern gulf is only affected by extreme impacts. As indicated by the field data itself, several different tsunami wave scenarios are needed to explain detected tsunami deposits in different parts of the study area. Generally, short-period tsunami waves ($T = 30$ min) from the south or west with any wave amplitude (2 m, 4 m, 6 m) are highly consistent with field data of the western Ambrakian Gulf. In contrast, long-period tsunamis ($T = 80$ min) from a western direction with 4 m to 6 m amplitudes show the best agreement with vibracore data collected in more central parts of the Ambrakian Gulf. Simulated tsunamis from the south-west generally do not match tsunami field traces. Therefore and because most potential tsunami sources in the central and eastern Mediterranean are located to the west (Mount Etna, Calabrian Arc) and south or south-east (Hellenic Trench) of the Ambrakian Gulf, it can be assumed that no major tsunamis from a south-western direction struck the area in the past.

The simulated scenarios do not only replicate the spatial distribution of tsunami candidate layers; they further show a number of essential sedimentary features typical of tsunamites, which were also detected in many vibracores of the study area. Such features are thinning- and fining-landward, basal erosional contacts and the marine to shallow marine origin of tsunami deposits. Furthermore, the numerical model accurately reproduces local tsunami inundation dynamics which were derived from field data for particular key sites in the western Ambrakian Gulf by previous sedimentary studies.

Nevertheless, there are some discrepancies between simulation results and field traces. In

particular, the computed sediment thickness in the numerical model usually lags behind the average thickness of tsunami candidate layers per vibracore. Also, there is no simulated scenario which can explain tsunami sedimentation indicated by vibracore data in the northern and eastern Ambrakian Gulf. Such inconsistencies may arise from (i) the comparison of single simulated tsunami waves with the average layer thickness per vibracore, (ii) tsunami candidate layers that give the impression of being related to a single tsunami event but actually relate to two or more events, (iii) false assumptions about the basic parameters in the numerical model (e.g. bottom roughness), (iv) hydrodynamic boundary conditions that do not represent realistic tsunami waves in the study area (e.g. single tsunami waves instead of tsunami wave trains), (v) changes in the local topography during the last 8000 years, which are not considered in the numerical model and finally from (vi) the limited spatial resolution of the model. Especially inconsistencies arising from (i) and (v) could be reduced in future studies, provided that detected tsunami candidate layers can be dated and correlated for most of the vibracores (cf. Chaps. 3, 4). Then, event strata of a particular pre-/historical tsunami impact could be compared with numerical simulations that consider the contemporary topography of the study area.

As demonstrated in this study, comparing and calibrating numerical models with sedimentary field data and computing sediment processes significantly increase the quality of numerical tsunami simulation results as a basis for modern tsunami risk assessment. Given the frequency of tsunami impacts in the past as indicated by sedimentary field data and historical accounts and considering the computed tsunami scenarios, a high tsunami risk has to be derived for the Ambrakian Gulf.

6 Synthesis and conclusions

This PhD study investigated tsunami impacts having occurred in the last 8000 years in two marine inlets of the west coast of Greece — the northern Gulf of Kyparissia and the Ambrakian Gulf. Forming the basis of the conducted studies in Chaps. 3–5, Chap. 2 provided a thorough treatise on the tsunami phenomenon, including tsunami wave physics, trigger mechanisms, sedimentary field traces and modelling techniques. Chapter 3 methodologically investigated the effects of different hydrodynamic boundary conditions (solitary and leading depression *N*-waves) and different palaeo-/topographies on the onshore tsunami response in a numerical model of the northern Gulf of Kyparissia. Based on the same methodology and simulation results, small-scale palaeotsunami inundation dynamics as indicated by field traces for three key locations in the northern Gulf of Kyparissia were reconstructed in Chap. 4. Regarding the second study area, the Ambrakian Gulf, various tsunami wave scenarios were simulated in Chap. 5, using a hydro- and morphodynamic numerical model in order to find scenarios that reproduce the spatial distribution of tsunami deposits detected all along the coast.

6.1 Tsunami risk assessment for the study areas

The findings of Chaps. 3–5 allow not only to reconstruct the dynamics, the extent and associated frequency of pre-/historical tsunami inundation in both study areas, but at the same time also to derive plausible tsunami wave scenarios for the future in view of a risk assessment. This has impressively been demonstrated by similar palaeotsunami studies in north-east Japan prior to 2011, which predicted a future tsunami impact for the area in the order of the 2011 Japan tsunami based on the analysis of the historical Jōgan tsunami in 869 AD (Chap. 1.1). Although such findings have never been implemented in regional zoning and mitigation plans, this example shows that palaeotsunami studies significantly contribute to a reliable tsunami risk assessment. Accordingly, in the following, the inundation extent and probability of possible future tsunami impacts in the northern Gulf of Kyparissia and the Ambrakian Gulf are discussed based on the synthesis of palaeotsunami field traces and numerical simulations presented in Chaps. 3–5.

Fig. 6.1 shows the expected minimum (deep red) and maximum (deep red plus light red) area of inundation for future tsunami impacts in the northern Gulf of Kyparissia, judging by the extent of past tsunami events indicated by tsunami field traces and hydrodynamic simulations (Chaps. 3, 4). Both areas derive from tsunami scenarios based on digital elevation model DEM₁₃, which is the most reliable representation of the contemporary topography associated with regional palaeotsunami inundation. In DEM₁₃, a moderate tsunami in the form of a 3 m solitary wave from a western direction with a wave period of 58 min is required to cause significant inundation of at least one of the three key locations in the northern gulf showing sedimentary palaeotsunami signatures — the valley of Kato Samiko. Applying this wave scenario to the recent coastland topography results in an expected minimum area of future tsunami inundation (deep red), meaning that this area would be affected in the case of the weakest tsunami impact indicated by field data for pre-/historical times. However, hydrodynamics in accordance with sedimentary tsunami field traces of all three investigated key sites in DEM₁₃ are associated with a stronger wave impulse, a 6 m LDN-wave from a western direction with a wave period of 92 min.

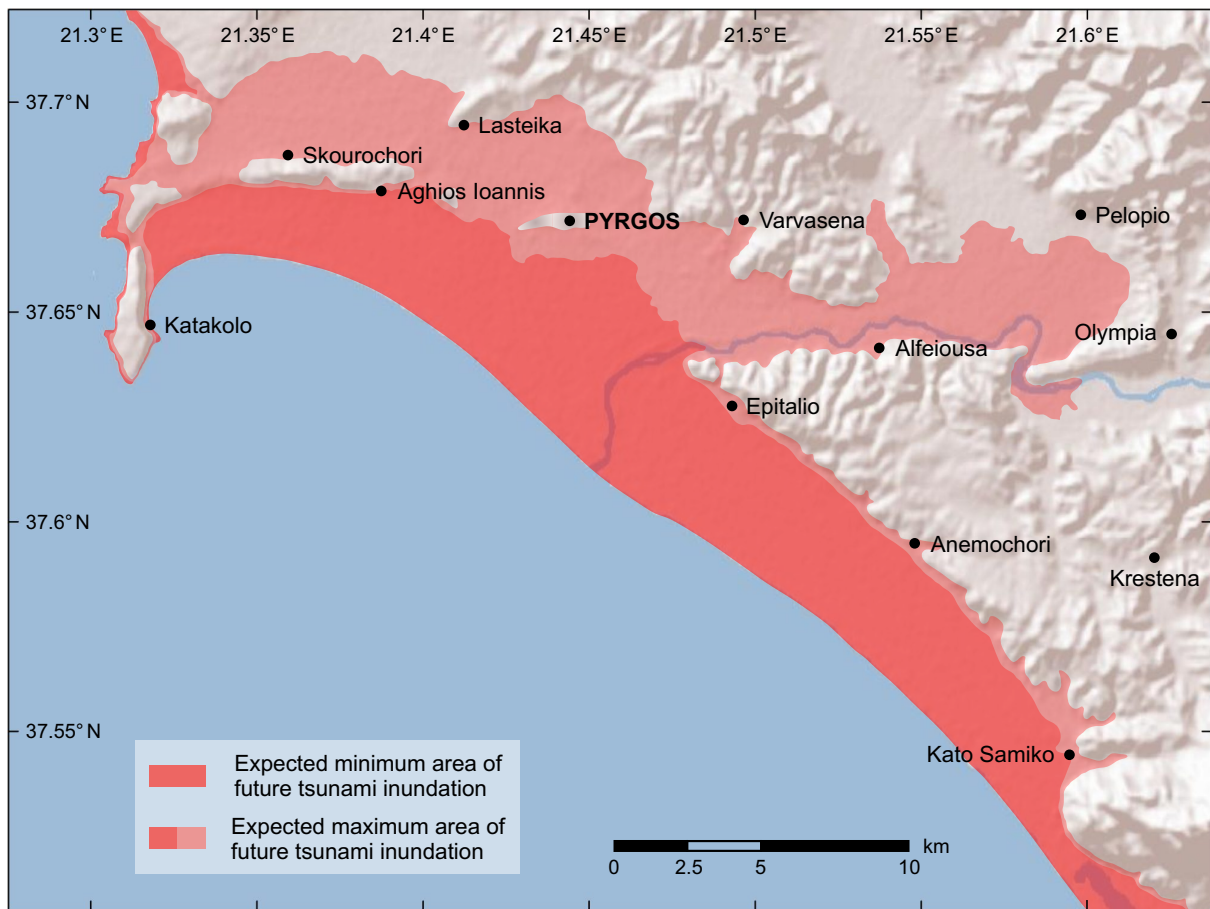


Figure 6.1: Area of possible inundation in the northern Gulf of Kyparissia by future tsunami impacts as implied by sedimentary tsunami field traces and hydrodynamic tsunami simulations presented in Chap. 3 and Chap. 4. Deep red defines the expected minimum area of inundation meaning the area that would be affected in the case of the weakest tsunami impact indicated by the field data for pre-/historical times. This impact is associated with a 3 m solitary wave from the west with a wave period of 58 min. Accordingly, the deep red plus light red area display the expected maximum inundation extent, which relates to an extreme tsunami event indicated by field traces (6 m LDN-wave from the west with a wave period of 92 min). Note that the map shows the major cities and villages only, excluding a large number of further villages and tourist centres along the coast in the area inundated.

In view of the recent topography of the northern gulf, this extreme scenario yields the expected maximum area of inundation (deep red plus light red) indicated by field traces.

As is evident from Fig. 6.1, a moderate tsunami (deep red) would result in vast inundation of the coastal lowlands in the northern Gulf of Kyparissia. This area gives home to numerous small villages, tourist centres and the larger villages of Katakolo, Aghios Ioannis and Kato Samiko, which are therefore at high risk of tsunami impacts. By contrast, the regional capital of Pyrgos, located on a small rise in the coastal lowlands, is well protected. Affecting the same region as before but additionally vast areas to the north-west of Pyrgos and in the lower Alpheios River valley, the expected maximum inundation extent (deep red plus light red) is far larger. In this extreme scenario, several villages to the north-west of Pyrgos, Pyrgos itself except for the inner centre, the villages of Epitalio and Anemochori plus a large number of villages in the lower Alpheios River valley would receive strong flooding.

The probability of future moderate to extreme tsunami events associated with the expected

minimum and maximum inundation extent in the northern Gulf of Kyparissia (Fig. 6.1) is difficult to estimate. According to SOLOVIEV *et al.* (2000, p. 14) and SCHIELEIN *et al.* (2007, pp. 164, 193–199), the average tsunami recurrence interval for the west coast of Greece in general is 53.9 and 50.6 years since 373 BC respectively, independent from the impact intensity (Chap. 1.2). However, moderate to extreme tsunami impacts as indicated by field data of the northern Gulf of Kyparissia are considerably less frequent. Summarising the number of not relating tsunami event layers detected in the three key locations of the study area yields a total of minimum 10 tsunami events since the beginning of the 6th millennium BC (Chap. 4.3), which results in an average recurrence interval of moderate to extreme tsunamis in the area of about 800 years. This period must be regarded as a minimum return interval as (i) deposits of further pre-/historical tsunami impacts might have disappeared and are consequently not retained in the geological record and (ii) not all tsunami events must have left sedimentary signatures in the key locations.

Using hydro- and morphodynamic simulations and field data presented in Chap. 5, an expected minimum and maximum area of future tsunami inundation can also be estimated for the Ambrakian Gulf (Fig. 6.2). Such estimations, however, do not consider the possible palaeo-landscape of the area but are solely based on the recent topography. The expected minimum area of inundation (deep red) derives from a moderate tsunami impact in the form of a 4 m LDN-wave from a southern direction with a wave period of 30 min. This is the weakest wave impulse associated with a similar sediment distribution in the numerical model as detected in vibracores in the western part of the Ambrakian Gulf. In this scenario, inundation is particularly observed for the city of Lefkada and its environs, the village of Aghios Nikolaos, entire Aktio Headland, large parts of the Preveza Peninsula including Preveza and Mitikas, the village of Kanali, the entire sandspit system in the northern Ambrakian Gulf, large parts of the Boukka coastal plain in the east and the village of Vonitsa. Tsunami deposits as found in the more inner parts of the Ambrakian Gulf require a stronger wave impulse to be reproduced in the numerical model—a 6 m LDN-wave from the west with a wave period of 80 min, which must be regarded as an extreme scenario. This scenario is associated with the expected maximum area of future tsunami inundation in the Ambrakian Gulf (deep red plus light red). Again, all major villages and cities of the region are inundated plus large parts of the coastal flats in the northern gulf, including the villages of Louros, Aneza and Nechoiri. Furthermore, the entire coastal plain of Boukka and the Bay of Vonitsa receive extensive flooding, and even the Bay of Palairos, located to the south of Lake Voulkaria, is affected. However, the city of Arta is again well protected against tsunami inundation.

As is the case for the northern Gulf of Kyparissia, moderate to extreme tsunami events in the Ambrakian Gulf are much less frequent than implied by the average recurrence intervals calculated by SOLOVIEV *et al.* (2000, p. 14) and SCHIELEIN *et al.* (2007, pp. 164, 193–199) for tsunamis of any intensity along the west coast of Greece. In total, field data suggests at least 8 tsunami impacts in the study area with the oldest event having occurred in the 6th millenium BC or shortly afterwards (Chap. 5.1). This results in an average return interval of moderate to extreme tsunamis in the Ambrakian Gulf of about 1000 years, which again has to be regarded as a minimum estimate due to the same reasons as mentioned above. Consequently, moderate to extreme tsunami impacts have struck both study areas, the Ambrakian Gulf and the northern Gulf of Kyparissia, with a similar frequency. On the one hand, this is due to the fact that some tsunami events have affected both marine inlets, as is indicated by the field data. On the other hand, both areas face a similar physiogeographical offshore setting with regard to seismicity and the submarine relief, which may result in a similar magnitude and frequency of tsunamis in the northern Gulf of Kyparissia and the Ambrakian Gulf.

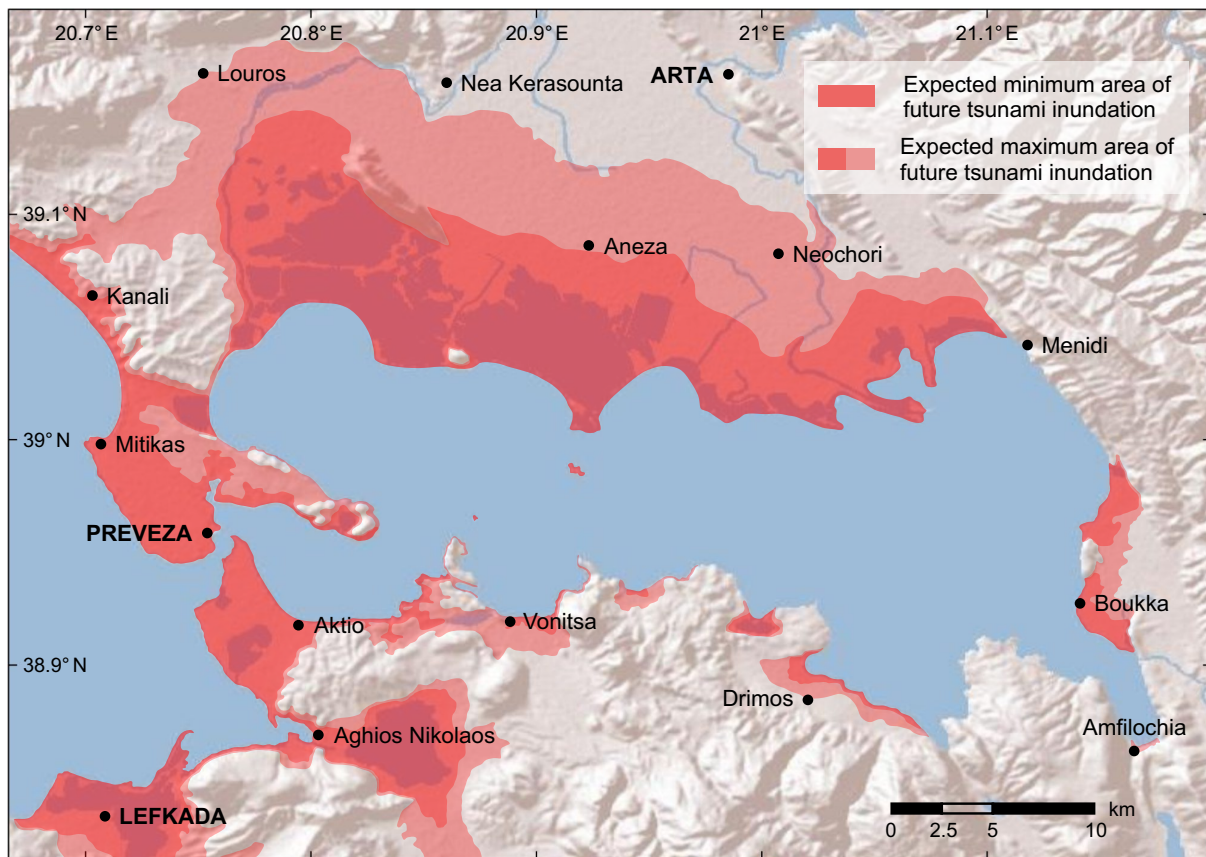


Figure 6.2: Area of possible inundation in the Ambrakian Gulf by future tsunami impacts as implied by sedimentary tsunami field traces and hydro- and morphodynamic tsunami simulations presented in Chap. 5. Deep red defines the expected minimum area of inundation meaning the area that would be affected in the case of the weakest tsunami impact indicated by the field data for pre-/historical times. This impact is associated with a 4 m LDN-wave from the south with a wave period of 30 min. Accordingly, the deep red plus light red area display the expected maximum inundation extent, which relates to an extreme tsunami event indicated by field traces (6 m LDN-wave from the west with a wave period of 80 min). Note that the map shows the major cities and villages only, excluding a large number of further villages and tourist centres along the coast in the area inundated.

6.2 Numerical simulation of palaeotsunamis as a base for risk assessment

Numerical tsunami simulations that take account of sedimentary palaeotsunami field traces significantly contribute to an understanding of regional tsunami risks, as is demonstrated by the current PhD thesis. However, this approach may demand a large number of assumptions which have to be made in the numerical model. With increasing age of the regarded tsunami impacts, also the uncertainty of the model generally increases. In order to still achieve reliable simulation results and by this a reliable risk assessment, the following items should be considered when calibrating and comparing numerical models with palaeotsunami field data (Fig. 6.3):

- Considering tectonic crust movements, littoral dynamics and sea level changes, the palaeogeography and palaeotopography associated with past tsunami impacts in the study area have to be reconstructed, which allows to create corresponding palaeo-DEMs as a basis for numerical simulations. For a realistic small-scale topography, detected tsunami deposits

6.2 Numerical simulation of palaeotsunamis as a base for risk assessment

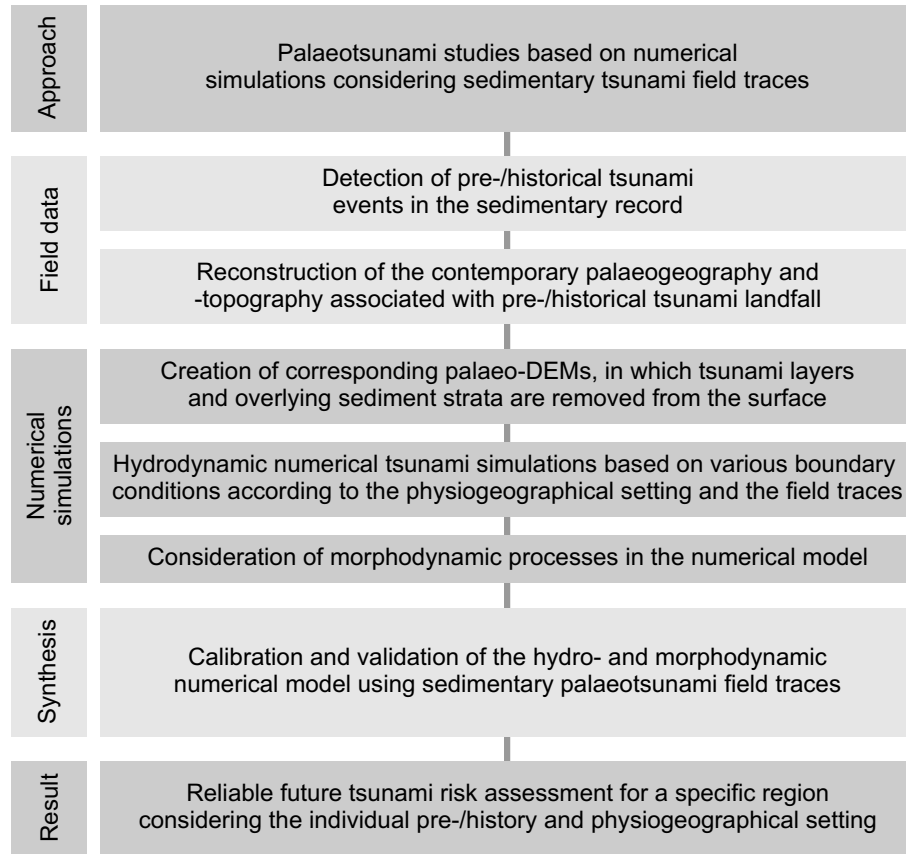


Figure 6.3: Schematic diagram summarising the idealised approach as derived from this PhD study to assess regional tsunami risks based on hydro- and morphodynamic numerical simulations considering sedimentary palaeotsunami field traces.

and overlying sediment layers should be subtracted from the surface in the palaeo-DEMs, a sufficient density of field data provided.

- The hydrodynamic boundary conditions, i.e. in particular the tsunami waveform, wave amplitude, wave period and propagation direction, are an essential input parameter of numerical tsunami models. If there is no information available on the trigger mechanism of pre-/historical tsunamis, various hydrodynamic boundary conditions according to the regional physiogeographical setting and the sedimentary field data have to be considered in the numerical model.
- For a direct comparison with field traces, numerical models should take account not only of the hydrodynamics but also of the morphodynamics of tsunami inundation. Morphodynamic models require information on the local sediment stratigraphy of the study area, which can be derived from vibracore data.
- Besides the hydrodynamic boundary conditions, there is a large number of further parameters to be determined and calibrated in hydro- and morphodynamic numerical tsunami models. Such parameters especially include the bottom roughness, the exact position of the open boundaries, the handling with wave reflections along open boundaries as well as the critical bed shear stress and other important morphodynamic parameters.

6 *Synthesis and conclusions*

Taking account of as many of the above listed points as possible will significantly increase the reliability of the numerical tsunami model. However, with each additional parameter or detail considered in the model, also the number of possible variations increases. In order to still understand the specific effects of each input variable on the simulation results and to bring into line numerical model and field data, often a large number of different scenarios has to be simulated. Following this approach, most of the points discussed above have been considered in the numerical tsunami models of the northern Gulf of Kyparissia and the Ambrakian Gulf presented in Chaps. 3–5. Although there is still room for further improvement in the created models (Chaps. 3.5, 5.5), close agreement could be achieved between the simulation results and field findings, which allows a reliable tsunami risk assessment for both study areas.

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