"Visualising the seismic landscape"

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Abstract

The "seismic landscape" of an earthquake-prone area is the result of progressive tectonic activity penetrating the earth's surface and considers all aspects which resulted from ground deforming seismic activity during the late Quaternary. Palaeoseismology studies earthquake geological effects composing seismic landscapes in order to understand seismic activity far back into history aiming to estimate future earthquake hazards in particular regions. In order to calculate the seismic hazard potential palaeoseismologists use empirical scaling laws between earthquake magnitudes and surface faulting parameters intending to extent the history of slip on a fault. This is commonly done by identifying earthquake recurrence intervals and maximum credible magnitudes. Therefore, reliable identification of earthquake geological effects, accurate measurements, critical but transparent and reproducible analysis, and high-performance geospatial visualisation techniques covering both, developments in space and time, are crucial attributes that provide new insights to earthquake hazards and improve assessments. Here, three innovative methodologies, addressing classical palaeoseismic trenching and sea-level indicator of uncertain archival trustworthiness, complying with above mentioned attributes are presented. The approaches combine continuously new findings from palaeoseismological research with technical innovations in order to reveal precious information that fill gaps in knowledge about the seismic activity of different areas.

Firstly, obtaining a more objective palaeoseismic trench log and a 3D view of the sedimentary architecture within the trench walls is the aim of a presented multiparametric workflow. It combines conventional techniques with applications of remote sensing and geophysical GPR measurements. Its usage is highly beneficial for identifying stratigraphic units and measuring representative layer thicknesses and displacements. Furthermore, the multispectral datasets are stored allowing unbiased input for future (re-)investigations.

Secondly, tidal notches are a generally accepted sea-level marker and pose a morphological feature on which is focused in more detail. However, considering tidal notches in rifting regions for palaeoseismic studies is controversially discussed since appearing displacements from present-day sea-level yield in unrealistic information about earthquake history when applied to common scaling laws. Therefore, two different workflows aiming to detect actual palaeostrandlines in a 3-dimensional manner and visualising tidal notch
sequence development through time by numerical modelling are presented. Consequential implications for palaeoseismological studies are the possibility to identify remnants of tidal notch morphologies caused by significant overwriting of pre-existing notch generations. Hence, newly identified palaeo-sea-levels become available for reconstructing particular sea-level histories. Furthermore, the presented numerical modelling algorithm considers a dynamic late-Holocene sea-level history and distinctly points out how rapid coseismic coastal displacements and gradual sea-level changes interplay. Its application reduces doubts in using tidal notches in palaeoseismological studies.

The methodologies presented contribute to quantifying palaeoseismological archives in space and time. In particular, hanging-wall sedimentary architecture and tidal notches pose such archives that can be accessed, addressed, analysed, and correlated to other earthquake environmental effects. Fusing high-quality data and innovative workflows improves objectivity in assessment and makes palaeoseismological interpretations robust. Therefore, elaborative but objective, reproducible, clear, reliable, and innovative methodologies providing valuable inputs to scaling laws significantly improve the evaluation of a seismic landscape and thus, decisively contribute to seismic hazard assessment.


Darauffolgend werden Meeresspiegelindikatoren hinsichtlich ihrer Anwendung in der Palaeoseismologie analysiert. Innerhalb des Tidenhubs geformte Brandungskeihlen sind weitestgehend als Meeresspiegelindikator akzeptiert und können Anzeiger seismischer Ak-

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## Contents

1 Introduction

1.1 Palaeoseismology in the Mediterranean .............................................. 5
1.2 Palaeoseismological evidence in the seismic landscape .......................... 6
   1.2.1 Palaeoseismological evidence in Quaternary intermontane basins ...... 8
   1.2.2 Palaeoseismological evidence on shorelines .............................. 11
   1.2.3 Earthquake environmental effects for seismic hazard assessment ...... 16
1.3 Techniques to visualise the seismic landscape ...................................... 16
   1.3.1 Technological milestones ..................................................... 18
   1.3.2 Techniques used to visualise the surface .................................. 19
   1.3.3 Techniques used to visualise the subsurface .............................. 19
   1.3.4 Modelling techniques to visualise and quantify palaeoseismological evolution ................................................................. 20
   1.3.5 Ground-penetrating radar (GPR) ............................................ 20
   1.3.6 Terrestrial laser scanning (TLS) ............................................. 22
1.4 Guide through this thesis ............................................................... 23

2 Active tectonic setting and seismicity of the Mediterranean .................... 29

2.1 Western Mediterranean Region ....................................................... 29
2.2 Central Mediterranean Region ....................................................... 30
2.3 Eastern Mediterranean Region ....................................................... 30
2.4 Geology of the study area ............................................................. 31
   2.4.1 Geodynamical setting .......................................................... 31
   2.4.2 Geological setting ............................................................... 34

3 Multiparametric interpretation of palaeoseismological trench exposures .... 39

3.1 Why visualise trench exposures ....................................................... 40
3.2 Geological setting of the trenching sites ............................................ 44
Contents

3.2.1 The Sfaka fault (NE Crete, Greece) .................................................. 45
3.2.2 The Kaparelli Fault (Gulf of Corinth, Greece) ..................................... 45
3.3 Visualising the trench ................................................................. 45
  3.3.1 Conventional trench logging and photomosaic .................................. 46
  3.3.2 TLS measurements ...................................................................... 46
  3.3.3 Imaging Spectroscopy .................................................................. 47
  3.3.4 Ground-Penetrating Radar .......................................................... 50
3.4 Results ......................................................................................... 51
  3.4.1 Sfaka Fault, Crete ........................................................................ 51
     3.4.1.1 Trench Log ........................................................................ 51
     3.4.1.2 Imaging Spectroscopy Analysis .......................................... 51
     3.4.1.3 GPR Data Interpretation ....................................................... 54
  3.4.2 The Kaparelli fault-trench, Gulf of Corinth ....................................... 58
3.5 Discussion .................................................................................... 60
3.6 Conclusion .................................................................................... 62

4 Tidal notch identification using fuzzy logic ........................................ 69
  4.1 Tidal notches in palaeoseismology ..................................................... 70
  4.2 Investigated sites ............................................................................ 73
     4.2.1 Agios Pavlos, SW Crete ............................................................. 73
     4.2.2 Perachora Peninsula, eastern Gulf of Corinth ............................... 73
  4.3 Methodology for visualising tidal notch morphologies ...................... 75
     4.3.1 Theoretical assumptions ............................................................ 75
     4.3.2 TLS ....................................................................................... 77
     4.3.3 Edge detection ......................................................................... 79
     4.3.3.1 Canny Method .................................................................... 79
     4.3.3.2 Fuzzy Logic ....................................................................... 80
     4.3.4 Hough Transform ................................................................... 80
  4.4 Results .......................................................................................... 83
     4.4.1 Developing methods, Agios Pavlos, SW Crete ............................. 83
     4.4.2 Testing methods, Perachora Peninsula, E Gulf of Corinth ........... 86
  4.5 Discussion and Concluding remarks on tidal notch detection .......... 88

5 Numerical modelling of tidal notch sequences on rocky coasts .......... 95
  5.1 Why model coastal cliff evolution? ................................................... 95
  5.2 Contributors to notch sequencing .................................................. 99
  5.3 Dynamic notch formation ............................................................... 103
  5.4 Methodology to model a tidal notch profile .................................... 105
  5.5 Results ........................................................................................ 107
     5.5.1 Uplifting coastal regions: Western/Eastern Gulf of Corinth and eastern Sicily and Calabria .......................... 107
List of Figures

<table>
<thead>
<tr>
<th>Figure</th>
<th>Title</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.1</td>
<td>Disciplines studying past earthquakes</td>
<td>4</td>
</tr>
<tr>
<td>1.2</td>
<td>Seismicity of the Mediterranean</td>
<td>5</td>
</tr>
<tr>
<td>1.3</td>
<td>The seismic landscape</td>
<td>9</td>
</tr>
<tr>
<td>1.4</td>
<td>Photo plate earthquake geological effects</td>
<td>10</td>
</tr>
<tr>
<td>1.5</td>
<td>Photo plate sea-level indicators</td>
<td>14</td>
</tr>
<tr>
<td>1.6</td>
<td>Field techniques</td>
<td>17</td>
</tr>
<tr>
<td>1.7</td>
<td>ground-penetrating radar</td>
<td>21</td>
</tr>
<tr>
<td>1.8</td>
<td>Terrestrial laser scanning</td>
<td>22</td>
</tr>
<tr>
<td>1.9</td>
<td>Roadmap</td>
<td>24</td>
</tr>
<tr>
<td>2.1</td>
<td>Geodynamics in the eastern Mediterranean</td>
<td>31</td>
</tr>
<tr>
<td>2.2</td>
<td>Geodynamics in central Greece</td>
<td>32</td>
</tr>
<tr>
<td>2.3</td>
<td>Geology of the eastern Corinthian Gulf</td>
<td>33</td>
</tr>
<tr>
<td>3.1</td>
<td>Guide through investigated trenching areas</td>
<td>41</td>
</tr>
<tr>
<td>3.2</td>
<td>Investigated trench parts</td>
<td>43</td>
</tr>
<tr>
<td>3.3</td>
<td>Flowchart</td>
<td>44</td>
</tr>
<tr>
<td>3.4</td>
<td>Downsampling of spatial information</td>
<td>49</td>
</tr>
<tr>
<td>3.5</td>
<td>Compilation of the Kaparelli trench log and multispectral results</td>
<td>52</td>
</tr>
<tr>
<td>3.6</td>
<td>Statistical analysis of two-channel composition image</td>
<td>55</td>
</tr>
<tr>
<td>3.7</td>
<td>Varying reflectance along the trench wall</td>
<td>56</td>
</tr>
<tr>
<td>3.8</td>
<td>3D reconstruction of the Sfaka trench</td>
<td>57</td>
</tr>
<tr>
<td>3.9</td>
<td>3D reconstruction of the Kaparelli trench</td>
<td>58</td>
</tr>
<tr>
<td>4.1</td>
<td>High-performance workflow for tidal notch detection</td>
<td>70</td>
</tr>
<tr>
<td>4.2</td>
<td>Study sites for notch detection</td>
<td>71</td>
</tr>
<tr>
<td>4.3</td>
<td>Overview map of investigated sites</td>
<td>74</td>
</tr>
</tbody>
</table>
List of Figures

4.4 Theoretic assumptions of notch identification ........................................ 76
4.5 Data acquisition and processing ............................................................... 78
4.6 Fuzzy set edge detection ........................................................................... 81
4.7 Comparison of applied edge detection algorithms ..................................... 82
4.8 Hough Transform ....................................................................................... 84
4.9 Feature extraction at Agios Pavlos ............................................................ 85
4.10 Results from Perachora Peninsula ............................................................ 86
5.1 Plate tectonics in the Mediterranean Basin ............................................... 96
5.2 Examples for deeply incised notches withstand ing cliff collapse in the Medi ter ranean .................................................. 100
5.3 Logic tree for tidal notch sequence evolution ........................................... 102
5.4 Assumptions leading to a quadric polynomial to cover requirements of tidal notch shape description ........................................ 104
5.5 Biasing sea-level curve and coastal uplift resulting in a tidal notch profile . 106
5.6 Sea-level curves of the Mediterranean ....................................................... 108
5.7 Time slices of notch development in the Gulf of Corinth ........................... 111
5.8 Time slices of notch development in southern Italy ................................ 112
5.9 Time slices of notch development in the Aegean Sea ............................... 113
5.10 Time slices of notch development in tectonically stable regions ............... 114
5.11 Time slices of notch development in subsiding regions ........................... 116
5.12 Time slices of notch development in the e. Gulf of Corinth considering coseismic activity .................................................. 118
5.13 Time slices of notch development in the w. Gulf of Corinth considering coseismic activity .................................................. 120
5.14 Comparing the model to reality ............................................................... 127
6.1 Application of TLS and Sfm in scientific work ........................................... 135
6.2 Geodetic constraints of Perachora Peninsula ............................................. 139
6.3 Notches of Perachora Peninsula .............................................................. 141
6.4 Tidal notch submerged at Schinos ............................................................. 144
List of Tables

3.1 Comparison of Photomosaic and HRDBSM values, and those from a composition of both ........................................... 53

5.1 Model parameters .......................................................... 107
5.2 Model parameters comprising coseismic displacements .............. 119

6.1 Statistically generated earthquake history of the Perachora Peninsula 145
6.2 Input parameters applied for tidal notch modelling ....................... 146
6.3 Statistically generated earthquake history of the Perachora Peninsula incl. offshore activity .............................................. 146
Success consists of going from failure to failure without loss of enthusiasm
W. Churchill (1874-1956)

So much universe, and so little time
T. Pratchett (1948-2015)
Palaeoseismology aims to understand seismic activity far back into history in order to estimate future earthquake hazards in a particular region [Ambraseys, 2009]. While other data sets are based on historical records palaeoseismological studies use late Quaternary geological and geomorphological data obtained from the seismic landscape [Wallace, 1981; Serva et al., 1997; Michetti and Hancock, 1997]. Instrumental data and historical catalogues are predominantly used in seismic hazard assessment (SHA). However, the time period these catalogues generally cover (Fig. 1.1) is too short to consider large recurrence intervals of particular faults [e.g. Wesnousky, 1986; Machette, 2000; Grützner et al., 2013] and do not encompass numerous earthquake cycles [Yeats and Prentice, 1996]. Consequently, the sample from the statistical elaboration of historical records is incomplete and a large number of faults would not have ruptured during the period in which the historical record is considered complete [Stucchi et al., 2004, 2013; Woessner and Wiemer, 2005; Guidoboni and Ebel, 2009; Grützner et al., 2013; Papanikolaou et al., 2015]. Therefore, palaeoseismological studies are often undertaken in order to extend the history of slip on a fault, to identify earthquake recurrence intervals, and maximum credible magnitudes [McCaulpin, 2009], and hence, produce valuable input for the accurate calculation of seismic hazard potentials of active fault zones [Reicherter et al., 2009].

Most of the required information on past earthquakes is provided from several kinds of coseismic ground deformation archived in different types of geological and geomorphological records [Michetti and Hancock, 1997]. As a result of progressive tectonic activity penetrating the earth’s surface, the seismic landscape considers all aspects which resulted from ground deforming seismic activity during the late Quaternary. Hence, coseismic (e.g. fault scarps) as well as post-seismic (e.g. colluvial wedges) expressions, on-fault effects (e.g. unconformities) as well as off-fault effects (e.g. uplifted/subsided shorelines), and all other permanent ground deformations directly related to tectonic stress [Dramis and Blumetti, 2005] form individual archives within the seismic landscape of an earthquake-prone area.

Beside the assumption recurring earthquakes are not randomly distributed through
time and space [Allen, 1975], scaling laws between earthquake magnitude and surface faulting parameters are a major tool in seismic hazard assessment. However, the investigation of sites for the possible hazard of surface fault rupture is a deceptively difficult task since the evidence for identifying active fault traces is generally subtle or obscure. Individual modification of palaeoseismic archives at certain rates and hence of varying significance occurs due to climatically driven and/or anthropomorphic processes. Mason et al. [2016] indicate the effect of natural erosion and anthropogenic activity on palaeoseismological estimates. Climatic changes such as from glacial to interglacial conditions pose a tremendous impact on the preservation of geomorphologic expressions. For instance, for the Mediterranean a common theory is that in postglacial times, the improved climatic conditions reduced erosion rates allowing fault scarpes caused by repeated coseismic surface displacements to be preserved [Benedetti et al., 2002; Papanikolaou et al., 2005; Reichelter et al., 2011]. Furthermore, from ~18 - 8 ka BP rapid sea-level rise linked to melting ice-sheets and water input to the oceans after the last glacial maximum (LGM) resulted into complex coastline changes refusing the development of distinct sea-level markers which can be used to infer coastal coseismic history [Pirazzoli, 1991].

Although earthquakes and associated natural hazards are not as frequently present in the world of media as in Japan or the western coastline of south-America, especially the eastern part of the Mediterranean basin depicts an excellent study area for investigating the composition of local seismic landscapes and its constituent parts. Ongoing seismic activity in different stress regimes (Fig. 1.2), dry-summer temperate climates, and microtidal conditions are advantageous circumstances to greatly produce and preserve archives of coseismic activity. Furthermore, long historical records have proven useful in associating seismicity with these palaeoseismological archives [McCalpin, 2009]. This poses best pre-conditions to study single earthquake-generated archives and assess associated hazards, and to carry out multidisciplinary studies to take full advantage of geological earthquake evidence. But also the development of new methodologies benefits from obvious physical expressions of a seismic facies [Michetti and Hancock, 1997] since it is easy to calibrate innovative processes towards improved identification or visualising the evolution of palaeoseismological archives.

Palaeoseismology is a rather young discipline in geoscience and one of its major challenges in future research is to build detailed relations between various categories of co-
seismic effects on the environment and earthquake magnitudes [Michetti et al., 2005]. Therefore, critical analysis of high-resolution data of geological and geomorphological features obtained from increasing possibilities of visualisation is required to enhance their identification, to measure their evolution in 4D, to store data sets allowing unbiased input for future (re)investigations, and to provide new insights to earthquake hazards. Combining continuously new findings from palaeoseismological research with technical innovations reveals precious information that fill gaps in knowledge about the seismic activity of different areas.

1.1 Palaeoseismology in the Mediterranean

Palaeoseismological studies aim to enhance understanding of seismic activity in a particular region commonly during the Quaternary. General guidelines on gathering information on past earthquakes are, among others, provided by Yeats and Prentice [1996], Michetti and Hancock [1997], Serva et al. [1997], Michetti et al. [2004], Dramis and Blumetti [2005], Michetti et al. [2005], Caputo and Helly [2008], Reicherter et al. [2009]. A comprehensive handbook on methods and fields of applications is provided by McCalpin [2009]. Regional studies cover almost every tectonically active region in the world. In the Mediterranean region studies focus on Iberia [e.g. Reicherter, 2001; Reicherter et al., 2003; Masana et al., 2005; Gracia et al., 2006; Silva et al., 2009; Grützner et al., 2010; Carbonel et al., 2013; van Balen et al., 2015], the Maghreb region [e.g. Swan, 1988; Galindo-Zaldívar et al., 2009; Heddar et al., 2013; Boulton et al., 2014], Italy [e.g. Roberts et al., 2002; Palumbo et al., 2004; Papanikolaou et al., 2005; Papanikolaou and Roberts, 2007; Galli et al., 2008; Ercoli et al., 2013; Wilkinson et al., 2015], Greece and Balkan peninsula [e.g. Pavlides, 1996; Collier et al., 1998; Meyer et al., 2007; Kokkalas et al., 2007; Papanikolaou and
Papanikolaou, 2007; Papanikolaou et al., 2013; Grützner et al., 2016b], Turkey [e.g. Rockwell et al., 2001; Cetin et al., 2003; Boulton and Whittaker, 2009; Fraser et al., 2010], and Israel and the Levant [e.g. Ben-Menahem, 1991; Amit et al., 2002; Zilberman et al., 2005; Daeron et al., 2007; Wechsler et al., 2014]. Additionally, there are several studies dealing with newly developed methodologies related to palaeoseismology preferably tested in above mentioned regions [e.g. Reiss et al., 2003; Wilkinson et al., 2010; Wiatr et al., 2013; Reitman et al., 2015; Schneiderwind et al., 2016].

1.2 Palaeoseismological evidence in the seismic landscape

Palaeoseismology uses geological and geomorphological evidence for deductive reasoning on earthquake events and then, gathering information on the event itself. First, geological structures that can exclusively be produced by an earthquake were summerised as "seismites" by Vittori et al. [1991]. To avoid misunderstanding the term refers to deformation structures in sediments induced by seismic shaking, today [Seilacher, 1969; van Loon, 2014]. In order to consider all permanent effects on the natural environment that are widespread generated by a crustal earthquake, Serva et al. [1997] preferred to use the term "seismic landscape" (Fig. 1.3).

Surficial features which are produced as a consequence of earthquakes that (at least partially) persist through time can be categorised as primary and secondary evidence (Fig. 1.4) in accordance to their triggering factors. Primary effects are those produced directly by tectonic deformation resulting coseismic slip along a fault plane. Evidence of secondary order are essentially the result of both seismic shaking and gravitational stress [Dramis and Blunetti, 2005]. Coseismic earthquake geological effects may include [e.g. McCalpin, 2009; Reichert et al., 2009]:

**Primary effects:**

- **Surface ruptures** are instantaneous geomorphic expressions occurring on-fault. A minimum magnitude threshold for surface rupturing is about 5.5 to 6.0 and the focal depth should not be deeper than ~20 km [e.g. Michetti et al., 2005]. Furthermore, surface ruptures might occur distributed over a significant fault width across strike. Such fault zones often include secondary surface ruptures, such as antithetic, en-echelon, and release faults. This is also why earthquake ruptures of subsequent events may not occupy the same trace. In correlation with the earthquake magnitude surface ruptures and displacements occur at different scales [e.g. Wells and Coppersmith, 1994]. The ESI 2007 scale lists surface ruptures from intensity VII on [Michetti et al., 2007]. For palaeoseismological studies in the Mediterranean it is assumed that surface ruptures preserved as fault scarps represent cumulative coseismic activity since the last glacial maximum (LGM) ~15±3 ka ago [Benedetti et al., 2002; Papanikolaou et al., 2005; Reichert et al., 2011]. Mason et al. [2016] demonstrate the variability of preserved originality along fault strike and give a guidance on where to derive deformation rates.
Seismic uplift/subsidence occurs instantaneous and affects also off-fault regions as for example shorelines. This type of deformation is rarely observed for intensity VIII events but commonly occurs during events of a higher order [Michetti et al., 2007]. It should be noted that, while uplift/subsidence is timed to the causative coseismic activity, the formation of indicators (e.g. shorelines) is dependent on erosive processes and hence, requires considerable time-periods. Furthermore, local topographic expressions (e.g. marine terraces) are seldom located directly at the causative fault and hence express only a proportion of cumulative coseismic displacements. When considering a varying erosional base [Schneiderwind et al., submitted] deformation rates can be inferred from palaeostrandlines [e.g. Leeder et al., 2003; Stiros et al., 2000; Shaw et al., 2008; Roberts et al., 2013; Schneiderwind et al., 2017].

**Secondary effects:**

**Mass movements** (e.g. rock fall, debris, rock slide, landslide, Sackung) may occur instantaneous (up to 6h after shaking) and postseismic (mostly retrogressive beyond the fault zone) as well as on-fault and off-fault and show up at intensity IV [Michetti et al., 2007]. It should be noted that besides cohesion, gravitational forces have exponential effects on slope stability. Thus, steep, normal fault controlled slopes are more likely to fail especially along structural discontinuities such as the fault plane itself [e.g. Keefer, 1984; Galen et al., 2014; Stahl et al., 2014]. Papathanassiou et al. [2017] mapped the density of landslides caused by the November 17th, 2015 Lefkada (Greece) earthquake and conclude that the majority of landslides was strongly related to pre-existing discontinuities that behaved as sliding planes.

**Liquefaction processes** are the result of transforming a granular material from a solid state to a liquified state as a consequence of ground-shaking and increasing pore water [e.g. Minos-Minopoulos et al., 2015]. Depending on local conditions, liquefaction may be frequent to occur in the epicentral area during intensity VIII events [Michetti et al., 2007]. However, liquefaction phenomena are also reported for events of Magnitude M5.0 [Papadopoulos, 1993; Papathanassiou et al., 2005].

**Dike injections** into low permeable capping layers on-fault and off-fault as stratigraphic expressions of coseismic activity. Weinberger et al. [2016] suggest that displacement along horizontal bedding planes is a viable mechanism to absorb deformation in well-bedded stata near the surface. Therefore, the authors derive estimates on coseismic horizontal slip by sheared clastic dike injections in the Dead Sea Basin.

**Ground cracks** that open the very shallow subsurface coseismically and form sub-vertical in downward-tapering zones. They show up from intensity IV but only where the lithology (e.g. loose alluvial deposits or saturated soils) and/or
morphology (e.g. slopes or ridge crests) support the development of millimeter-wide cracks. Events of higher intensities (e.g. X) are frequently accompanied by metre-scale open ground cracks affecting areas in the order of 5,000 km² [Michetti et al., 2007; Reicherter et al., 2009]. For instance, discontinuous cracks over a total length of ~400m have been recorded at the hanging-wall of the Mt. Vettore fault (central Italy) after the August 24, 2016 Amatrice earthquake event (M6.0) [Livio et al., 2016].

Tsunami deposits and erosional unconformities caused by tsunamis. Depending on the shape of sea floor and coastline, tsunamis may reach the shores with runups exceeding 5 m flooding flat areas for thousands of metres inland during intensity X earthquakes [Michetti et al., 2007]. Sedimentological evidence of past tsunamis given by coarse grained intervals from fining up and thinning up sequences and clastic marine layers including shell debris, foraminifera and rip-up clasts has been found throughout Mediterranean [Reicherter and Becker-Heidmann, 2009; Reicherter et al., 2010b,a].

Surface faulting and fracturing are effects that might result from both tectonic deformation and seismic shaking/gravitational stress [Dramis and Blumetti, 2005].

1.2.1 Palaeoseismological evidence in Quaternary intermontane basins

As a result of repeated earthquakes producing a cumulative appearance of geological effects, the geomorphology progressively changes and forms out typical landforms (geomorphic evidence), such as fault escarpments, horsts, graben, rift valleys and elongated ridges, and palaeoshorelines different from present-day sea-level. These geomorphologies mainly result from primary surface ruptures. For instance, fault scarpes are the projection of the fault plane onto the surface. Therefore, the relationship between surface displacement, rupture length, and earthquake magnitude follows empirical regression [e.g. Wells and Coppersmith, 1994]. Stratigraphic evidence is given by deposits and sedimentary structures, such as displaced palaeosols.

Geomorphological analyses in palaeoseismological studies [e.g. Silva et al., 2003; Ganas et al., 2005; Pérez-Peña et al., 2010; Boulton et al., 2014] generally aim to assess rates of tectonic and seismic activity by numerous measurements over a large area, but often do not provide precise ages for palaeoeartquakes [McCaffrey, 2009]. The stratigraphic approach uses trenching methods which intend to accurately measure fault displacements and infer earthquake recurrence by dating deformed strata [e.g. Pantosti et al., 1996; Rockwell et al., 2000; Reicherter et al., 2003; Grützner et al., 2010; Rockwell et al., 2010; Ferrater et al., 2016]. However, such detailed mapping is costly and time consuming and is typically undertaken only for selected sites which are assumed to be representative for the studied tectonic structure. These sites are constrained by the availability of stratigraphic evidence (sedimentation on both footwall and hanging wall, or continuous sedimentation in the hanging wall with the extraction of colluvial wedges etc.). As a reconnaissance technique, geophysical methods are commonly used in order to detect buried stratigraphic evidence.
1. Introduction

Figure 1.3: Schematic block diagram of Quaternary intermontane basins showing hierarchical classification of palaeoseismic evidence from repetitive surface rupturing normal faulting earthquakes. Both, primary and secondary earthquake environmental effects compose a seismic landscape including seismo-tectonic and seismo-gravitational landforms [modified after Dramis and Blumetti, 2005].
1. Introduction

**Figure 1.4**: Photo plate of primary and secondary geological evidence for tectonic activity.

a) Hill slope highly affected by repeated coseismic events (Loutraki fault, eastern Gulf of Corinth). White arrows indicate bedrock fault scarp, b) bedrock normal fault scarp (Loutraki fault, eastern Gulf of Corinth), c) uplifted terraces (Perachora Peninsula, eastern Gulf of Corinth), d) displaced stratigraphy (Corinth Canal), e) ground cracks (Atacama Desert, Chile), and f) head-scarps (dashed white lines) of subsequent landslide events (Messara Basin, Crete).
and hence, to locate the best trenching site [e.g. Chow et al., 2001; Green et al., 2003; Reiss et al., 2003; Vanneste et al., 2008; Grützner et al., 2010; Roberts et al., 2010]. Since each approach holds subject to ambiguity in interpretation, multidisciplinary approaches combining remote sensing, geophysical methods, geomorphological and sedimentological analyses [e.g. Bubeck et al., 2015; Grützner et al., 2016b; Mason et al., 2016; Schneiderwind et al., 2016] help to identify past earthquakes from primary and secondary evidence and allow determining tectonic and seismic activity more precisely.

1.2.2 Palaeoseismological evidence on shorelines

Especially in extensional tectonic settings seismic activity is often accompanied by vertical land displacements. Palaeoseismological studies in coastal areas use the relative sea-level as a regional reference datum and focus on the identification of fossil palaeoshorelines which might have been rapidly displaced from palaeo-sea-level due to coseismic slip. The list of sea-level indicators (Fig. 1.5) may include [from Pirazzoli, 1991; Shennan et al., 2015]:

Erosional geomorphological indicators:

- **Pools** that may form out in cemented sediments [e.g. Kelletat, 2006]

- **Notches** on steep, plunging cliffs are the result of biological, physical, and chemical erosion at mean sea-level [e.g. Pirazzoli, 1986; Kershaw and Guo, 2001; Trenhaile, 2015]. Espacially 'tidal notches' in micro-tidal regimes pose a preferred sea-level marker since these form distinct morphological and ecological erosional features developed only within the tidal range [Antonioli et al., 2015]. When tidal notches appear different from present-day sea-level, they are commonly used in palaeoseismological studies in order to infer coseismic history. By using high-resolution data provided by terrestrial laser scanning, Schneiderwind et al. [2017] are able to identify multiple morphological evidence even in between major emergence. However, a varying relative sea-level during the late Holocene has to be taken into account for reconstructing seismic activity [Schneiderwind et al., submitted].

- **Wave-cut platforms** are the result of deep submarine abrasion while sea level is stationary [e.g. Bradley, 1958; Swift, 1968; Trenhaile, 2000]. For instance, Roberts et al. [2009] used raised wave-cut platforms in the eastern Gulf of Corinth (central Greece) and dated corals found in palaeoshoreface deposits to infer long-term slip rates of the coast uplifting normal fault.

Depositional geomorphological indicators:

- **Beaches** as coastal depositional environment [e.g. Gerald M. Friedman, 1961; Gischler and Lomando, 1997]
1. Introduction

Estuary or lagoon floors formed by topographic balancing changing from erosional to depositional environment close to the shore [e.g. Brinson et al., 1995; Lambbeck and Chappell, 2001; Shennan and Horton, 2002]

Beachrocks are cemented coastal sedimentary formations lithified by the precipitation of carbonate cements [e.g. Vousdoukas et al., 2007]. For instance, a multiproxy sedimentological and geomorphological approach applied to the Lefkada barrier and beachrock system (NW Greece) by May et al. [2012] enables the authors to infer information on extrem wave events that contributed to coastal and environmental changes

Coral conglomerates occurring as subhorizontal-surfaces flagstones of cemented coral rubble located behind adjacent outer reef flats [e.g. Darwin, 1842; Montaggioni and Pirazzoli, 1984]

Biological indicators:

Coral reef ecosystems comprising of different but interconnected habitats dependent on the exchange of seawater, sediments, nutrients, and marine fauna [e.g. Darwin, 1842; Fairbanks, 1989]

Barnacles, sessile encruster living in shallow and tidal waters [e.g. Laborel and Laborel-Deguen, 1994]. Since the habitat of barnacles is distinctly terminated the upper limit of occurrence is suitable for deriving relative sea-level changes at high confidence [Morhange et al., 2001]

Sea urchins living in the infralittoral zone below low tide levels pose a strong contributor to coastal bioerosion [e.g. Laborel and Laborel-Deguen, 1994]

Limpets living within the tidal range as important agents of bioerosion [e.g. Laborel and Laborel-Deguen, 1994]

Chitons living next to limpets in the midlittoral zone [e.g. Laborel and Laborel-Deguen, 1994]

Sponges living always submerged posing an active erosive agent [e.g. Laborel and Laborel-Deguen, 1994]

Lithophaga are boring bivalves responsible for a rapid erosion of limestones at sea-level [e.g. Laborel and Laborel-Deguen, 1994]. Pirazzoli et al. [1994b] point out, Lithophaga and other vermetid shells are killed by uplift at a site and remain in growing position. Only when these are sheltered from erosion their organic material can be dated. However, sediment infill to the lithophaga boreholes or other bioconstructions capping lithophaga agents after their displacement will affect radiocarbon dating results.
Sedimentological and stratigraphical indicators:

Foraminifera revealing even low-amplitude sea-level fluctuations in stratigraphic deposits [e.g. Gehrels, 1999]. In tsunami research, micropalaeontological analysis of drill cores has the potential to distinctly differentiate terrestrial from coastal sediments. For instance, varying microfauna associations, showing a combination of foraminifera from different habitats, may be interpreted as indication for event layers [Mathes-Schmidt et al., 2013].

Pollen and spores of terrestrial plants [e.g. Godwin, 1940; Ellison, 1989]

Diatoms are certain types of algae predominantly living in nutrient-rich waters [e.g. Zong and Horton, 1999]

Ostracods occur in the upper layer of the sea-floor as part of the benthos [e.g. Yokoyama et al., 2000]

Mollusca, especially those from shallow waters [e.g. Milliman and Emery, 1968; Ávila, 2000]

Saltmarshes are coastal ecosystems located on a gently dipping shore immersed during tidal highstands [e.g. Milliman and Emery, 1968]

Archaeological indicators:

"Submerged forests" as indicator for coastal subsidence and/or sea-level rise [e.g. Godwin and Newton, 1938; Brinson et al., 1995]

Submerged structures that used to have their foundations on dry land (e.g. coastal quarries, houses, tombs). The Roman Kenchreai harbor is located on the west coast of the Saronic Gulf and appears submerged with respect to present-day sea-level. Due to its location directly at the hanging-wall of the Kenchreai fault subsidence could be related to coseismic activity or Holocene sea-level rise. However, geomorphological analysis has shown that the fault dies out at the harbor. Therefore, Koukouvelas et al. [2017] suggest the drowning of the harbor is the result of late Holocene sea-level rise.

Harbour installations dislocated from present-day sea-level [e.g. Stiros and Blackman, 2014]

Fish tanks dislocated from present-day sea-level. Mourtzas [2012] points out, that fish tanks are relics of the Roman domination and form a sensitive sea-level indicator at present.

Emerged structures (e.g. ship-wrecks, ship anchors, harbour foundations) [e.g. Stiros and Blackman, 2014]
1. Introduction

Once former shorelines are identified the elevation difference to the present datum is usually measured, and dependent on the type of sea-level indicator also dated. Among archaeological indicators, the most common indicators used for palaeoseismological studies are coastal geomorphological features [e.g. Keraudren and Sorel, 1987; Dumas et al., 1993; Armijo et al., 1996; McNeill and Collier, 2004; Roberts et al., 2009]. In particular, in the microtidal Mediterranean Sea tidal notches have been used to infer Holocene palaeosea-level positions at decimetre confidence [e.g. Pirazzoli et al., 1982, 1989, 1991, 1994a; Stewart and Vita-Finzi, 1996; Rust and Kershaw, 2000; Kershaw and Guo, 2001; Evelpidou et al., 2012b; Antonioli et al., 2015; Goodman-Tchernov and Katz, 2016]. Indicator-dependent is the information on time of formation. Biological indicators are unlikely to resist subsequent erosion through time, however, if preserved they are available for radiocarbon dating which is one the most trustworthy and easiest dating methods. Roberts et al. [2009] successfully applied \(^{234}\text{U}^{230}\text{Th}\) dating to corals preserved in uplifted shorefaces [see also Houghton et al., 2003] and reconstructed Quaternary slip rates since 340 ka. Indirect dating approaches are commonly applied to tidal notch exposures. In a conceptual approach Boulton and Stewart [2015] compared local sea-level curves with associated regional uplift estimates and concluded that the highest elevation tidal notch on uplifting coasts dates to 6,000 years BP. Not until that time did the rate of eustatic sea-level rise decrease to 1 mm/yr and reach gravitational equilibrium with the continental lithosphere [Caminati et al., 2003; Stocchi et al., 2005]. Utilising this information measured notch elevations are then typically compared to the linear regression of a seismic cycle and reported earthquake events. However, the usage of sea-level marker in palaeoseismological studies is controversially discussed. In rifting regions, observable metre-scale offsets between individual tidal notches and/or recent sea-level are not attestable with modern palaeoseismological scaling laws. Furthermore, Antonioli et al. [2015] concluded in their comprehensive analyses on tidal notches in the Mediterranean Sea that the average notch height appears always higher than the local tidal range implying driving forces for notch development are not limited to the tidal range as generally stated [e.g. Pirazzoli, 1986; Pirazzoli et al., 1989; Pirazzoli and Evelpidou, 2013; Evelpidou and Pirazzoli, 2016]. However, the interplay between notch widening due to gradual sea-level changes and coseismic uplifts which do not exceed the tidal range has not been acceptably investigated, yet.

**Figure 1.5:** Photo plate of selected geomorphological and biological sea-level indicators. a) Algae rims (arrows) at the ancient harbour of Sougia, Crete, b) estuary floor (Sougia, Crete), c) raised notch photographed at low-tide (Perachora Peninsula, eastern Gulf of Corinth), d) raised beachrock (arrows) at Myloko Bay, Perachora Peninsula, eastern Gulf of Corinth, e) grazing chitons (arrows) (Perachora Peninsula, eastern Gulf of Corinth), f) limpets (Perachora Peninsula, eastern Gulf of Corinth), g) sea-urchins at low-tide level (dashed white lines) (Strava, Perachora Peninsula, eastern Gulf of Corinth), and h) Lithophaga borings within the tidal range (Skalosia Bay, Perachora Peninsula, eastern Gulf of Corinth).
Figure 1.5: See facing page.
1.2.3 Earthquake environmental effects for seismic hazard assessment

The needs for a general classification, quantification and measure for a wide variety of geological, hydrological, botanical and geomorphic palaeoseismological evidence for different intensity degrees have led to the development of the ESI 2007 scale [e.g. Michetti et al., 2007; Reicherter et al., 2009; Papanikolaou et al., 2009]. This scale provides an important tool for seismic hazard assessment for two reasons. Firstly, traditional intensities are generally based on human parameters. Furthermore, intensities derived from effects on humans and human environments predominantly reflect the economic development and the cultural setting of the area that experienced the earthquake [Serva, 1994]. Other than that, macroseismic intensity forms a major seismic hazard parameter describing environmental damage patterns after earthquake events without subjective influences from human parameters. Due to independent mapping of seismic hazards, earthquake environmental effects (EEE) and their quantification can directly be used through the ESI 2007 scale into seismic hazard assessment [Papanikolaou et al., 2009; Papanikolaou, 2011]. Secondly, a scenario of expected earthquake environmental effects for certain regions can be made if an earthquake Magnitude and its ESI 2007 scale intensity are known and vice versa. This helps to address problems relating to the incompleteness of the historical records and obtains higher spatial resolution and realistic source locality distances by using empirical relationships between coseismic slips, rupture lengths, earthquake magnitudes [e.g. from Wells and Coppersmith, 1994], and intensity distributions on geologically constrained active faults [Papanikolaou, 2011].

Seismic hazard maps purely produced from geological data rely on slip-rate data and fault geometry [Papanikolaou et al., 2013; Grützner et al., 2016b]. Obviously, well-founded measurements should pose inputs for throw-rate calculations and subsequent assessments. Rectified imaging of primary and secondary coseismic earthquake environmental/geological effects improve their quantification. Therefore, innovative techniques visualising the seismic landscape provide useful and well-founded information relevant for seismic hazard assessment.

1.3 Techniques to visualise the seismic landscape

During the past few decades many alternative techniques to drawings and sketches have been developed to obtain visible and non-visible information from earthquake geological effects, always aiming to identify their full extent and/or extracting slip rates, recurrence intervals and displacements per event (Fig. 1.6). Moreover, three main categories of methodologies and strategies used in palaeoseismology to visualise earthquake-related information can be distinguished: I) conventional methods including mapping and fault-trench investigations [e.g. Sieh, 1978; Hatheway and Leighton, 1979; Sieh, 1981; Rockwell et al., 2000; McCaolin, 2009; Baize et al., 2015], II) advanced techniques from diverse disciplines of remote sensing [e.g. Massonnet et al., 1993; Walters et al., 2011; Wilkinson et al., 2012; Bennis et al., 2014] and geophysics [e.g. Vanneste et al., 2008; Štěpančíková et al., 2011; Grützner et al., 2012] adding information to the conventional datasets, also including multidisciplinary studies that integrate earthquake evidence from different per-
Figure 1.6: Selected field techniques to obtain images of palaeoseismological evidence from the surface and the subsurface. Area size is classified as (+) small, (+++) medium, and (++++) large ground cover. UAV = unmanned aerial vehicle; TLS = terrestrial laser scanning (also t-LiDAR); Sfm = structure-from-motion photogrammetry; H/V = horizontal to vertical microtremor spectral ratio; ERT = electric resistivity tomography; GPR = ground-penetrating radar; SAR = synthetic aperture radar; VNIR = visible and near-infrared; SWIR = short-wave infrared; TIRS = thermal infrared sensor; UVN = ultra violet, visible, near infrared; NIR = near infrared.

spectives [e.g. Meghraoui et al., 2003; Reichert et al., 2003; Silva et al., 2009; Bubeck et al., 2015; Mason et al., 2016; Grützner et al., 2016b,a], and III) modelling geological and geomorphological processes based on observations and assumptions [e.g. Maniatis et al., 2009; Cowie et al., 2012; Turpeinen et al., 2015; van Balen et al., 2015].

For this study, mainly ground-penetrating radar (GPR) and terrestrial laser scanning (TLS) were used to visualise earthquake geological features of primary and secondary evidence. Numerical modelling of coastal uplift and sea-level marker development was applied to image the interplay of rapid coseismic coastal uplift and gradual sea-level change.
1. Introduction

1.3.1 Technological milestones

Thousands of years ago, when science was closely linked to philosophy, knowledge arose from experience since observations were the only data source available. A very basic technique to document observations and considerations is to draw a sketch showing descriptive characteristics. This was still the case in the early days of palaeoseismology when McGee [1892] described "a fossil earthquake" in North America. Furthermore, not until the 1970s modern palaeoseismology developed [Vittori et al., 1991]. The availability to access historical and instrumental data on seismicity became of primary importance to hazard studies and opened the opportunity to visualise spatial cluster of earthquake epicentres and localise active seismogenic structures [e.g. Allen et al., 1965; Wallace, 1981]. At that time researchers started mapping geological earthquake expressions for understanding earthquake hazard [e.g. Allen, 1975]. Additionally, steep development curves in the aerospace industry offered a complete new view on the earth’s surface and were gratefully accepted by earth scientists. For instance, palaeoseismologists started to use SPOT or Landsat satellite imagery to map fault traces and infer overall orientations [e.g. Ni and York, 1978; Papazachos et al., 1984; Lyon-Caen et al., 1988]. However, not only the viewing perspective introduced by satellite imagery was novel, but also the opportunity of monitoring the earth's surface and to visualise surface deformations in the fourth dimension; time [e.g. van Puymbroeck et al., 2000; Leprince et al., 2008]. In 1992 another milestone in earth observation and monitoring was reached. Active radiometric measurements (e.g. synthetic aperture radar - SAR) became available and thenceforward produce images of high spatial resolution, that, when collected at certain time intervals, can image co-seismic displacements at high precision [Massonnet et al., 1993]. Also in the beginning of the 1990s palaeoseismologists started using GPR systems in order to image progressive displacements in the shallow subsurface [e.g. Benson, 1995; Meschede et al., 1997]. However, technological revolution did not stop. Obtaining high-resolution morphological information from differential GPS measurements [e.g. Brasington et al., 2000] or digital photogrammetry [e.g. Lane et al., 1993; Chandler, 1999] were the next steps to produce images of surfaces at a new level of resolution. Technological developments and improvements still accelerate and so airborne and also terrestrial laser scanning have revolutionised the quality of digital elevation models, extending their spatial extent, resolution, and accuracy, what successively raised the number of applications in palaeoseismology [e.g. Lohani and Mason, 2001; McCaffrey et al., 2005; Rosser et al., 2005; Jones et al., 2007; Arrowsmith and Ziolk, 2009; McCalpin, 2009; Meigs, 2013; Bubeck et al., 2015; Wilkinson et al., 2015; Mason et al., 2016]. Parallel to that evolution photogrammetry enhanced. Image-based terrain extraction from overlapping stereo-pairs has become widely used for topographic modelling [e.g. Chandler, 1999; Hancock and Willgoose, 2001; Remondino and El-Hakim, 2006]. The latest advancement in image-based 3D modelling is 'structure-from-motion' (SfM) photogrammetry that operates under the same basic tenets as stereophotography and thus uses overlapping, but offset images to reconstruct 3D structures. However, a highly redundant, iterative bundle adjustment algorithm automatically extracts features from multiple overlapping images to specify the geometry of the scene [Westoby et al., 2012].
1.3.2 Techniques used to visualise the surface

Typically, the first approach in palaeo-seismological studies is locating and mapping surface deformation [McCalpin, 2009]. Aerial photographs and high-resolution satellite images provide visible information on surface lineaments. Modern earth-observation programs such as Landsat (NASA) or Copernicus (ESA) aim to map and monitor the earth-surface in order to improve environmental managements [e.g. Davis et al., 2016]. Multispectral images capture data of certain surfaces at specific frequencies across the electromagnetic spectrum and therefore allow extraction of additional information of surface material properties [e.g. Novak and Soulakellis, 2000]. Topographic information is valuable for evaluating the balance between tectonic movements and erosion/sedimentation, and therefore, helps to identify areas with tectonic activity [Burbank and Anderson, 2001; Donnellan et al., 2016]. State-of-the-Art techniques to obtain topographic information are high temporal and spatial resolution SAR (Synthetic Aperture Radar) data [e.g. Salvi et al., 2012], high-resolution airborne LiDAR surveys [e.g. Zielke et al., 2015], and ground-based and UAV (unmanned aerial vehicle)-based photogrammetry methods [e.g. Westoby et al., 2012; Bemis et al., 2014]. Such data pose input for the calculation of a number of geomorphic indices that provide quantitative measures on tectonic activity. The most common indices focus on drainage pattern (stream length gradient index, SL; concavity index, Ac), catchment geometries (valley floor width/height ratio, Vf; basin shape index, Bs, asymmetric factor, Af; hypsometric integral, HI), or mountain front geometry (mountain front sinuosity, Smf; terrain ruggedness index, TRI) [Hack, 1973; Bull and McFadden, 1977; Rockwell et al., 1984; Silva et al., 2003; Peters and van Balen, 2007; Viveen et al., 2012; Kale et al., 2014; Fountoulis et al., 2015; Zygouri et al., 2015]. Furthermore, high-resolution topographic data provides the potential to map uplifted marine terraces in a broad spatial extent [Jara-Muñoz et al., 2015].

Once geomorphological features of tectonic activity are located, these are commonly mapped in detail. Well preserved fault scarps may exhibit kinematic indicators which can be utilised to visualise the palaeostress regime [e.g. Doblas, 1998]. Slip rate trends are provided by topographic profiles measured perpendicular to the strike of the scarp illustrating the postglacial [Benedetti et al., 2002; Reicherter et al., 2011] vertical throw [Avouac, 1993; Michon and van Balen, 2005; Papanikolaou et al., 2005]. In order to replace arduous manual profiling and time consuming geodetic compass measurements, ground-based terrestrial laser scanning (TLS) has become a favoured technique that captures full 3D high-resolution models of mountain slopes [Gold et al., 2013; Bubeck et al., 2015; Wilkinson et al., 2015; Mason et al., 2016] and allows extracting detailed geometries and kinematics of fault scarp exposures [Wiatr et al., 2013].

1.3.3 Techniques used to visualise the subsurface

The only way to directly visualise earthquake geological effects in the subsurface is to excavate a trench. Fault-trenches expose stratigraphic expressions of palaeoearthquakes and may provide access to datable material. However, to create and preserve displaced strata and angular unconformities balanced conditions between tectonics and erosion/sedimentation rates are required. These primary evidence vary a lot along strike of a certain fault [Bubeck et al., 2015] and geophysical prospection surveys support trenching
site selection by visualising physical properties of sedimentary structures in the subsurface. Seismic methods contribute to palaeoseismological studies by detecting faults in greater depth and by characterising subsurface strata that have been offset, folded, or tilted by faulting [McCalpin, 2009]. However, even high-resolution seismic methods do not allow differentiating individual colluvial wedges of 1 to 2 m thickness against the main fault plane [e.g. Stephenson et al., 1995; Zilberman et al., 2005]. More suitable for detecting and assessing earthquake related sedimentary structures close to the surface [Neal, 2004] and therefore commonly used are electric resistivity tomography (ERT) and ground-penetrating radar (GPR) [Papanikolaou et al., 2015]. Geoelectrical surveys aim to detect lithological patterns by imaging near-surface electrical resistivity, whereas GPR investigations image buried structures by transmitting pulsed electromagnetic waves through the subsurface. Therefore, geophysical prospection has been proven suitable for estimating fault offsets and locating promising sites for trench excavations [e.g. Demanet et al., 2001; Štěpančíková et al., 2011; Grützner et al., 2012].

1.3.4 Modelling techniques to visualise and quantify palaeoseismological evolution

Numerical modelling of landscape evolution simplifies and idealises the real world in order to visualise and assess palaeoseismological arguments and consequences. The most common first order modelling approach in palaeoseismology is the concept of a seismic cycle which serves probabilistic-based seismic hazard assessment. Here, a simple linear function being the best fit regression through dated earthquake events, bounded by certain confidence intervals is representative for earthquake recurrence in both the past and the near future [e.g. Petersen et al., 1996; McCalpin, 2009; Papanikolaou et al., 2015; Rockwell, 2016]. Other common applications in geoscientific modelling are coulomb stress and stress transfer [e.g. Stein et al., 1994; Hubert et al., 1996] and scarp degradation [e.g. Colman and Watson, 1983; Mattson and Bruhn, 2001; Kokkalas and Koukouvelas, 2005]. As computers have increased in memory size and performance more complex models have become applicable, treating the real-world complexity by discretizing space and time. Numerical models allow understanding linkages among geological processes in a quantitative sense. One of the big advantages of modelling geological and tectonic processes by embedding knowledge of these processes themselves is the permission to explore where state-of-the-art understanding of these processes is weak and insufficient. Among others, generic numerical models provide visual images of landform generating processes and iteratively force to produce new data, for instance by field observations, which again pose new input elements in the models [Burbank and Anderson, 2001].

1.3.5 Ground-penetrating radar (GPR)

Ground-penetrating radar (GPR) has been used for several questions of archaeology and palaeoseismology, and has been predominantly proven useful for characterising palaeoseismic sites [e.g. Green et al., 2003; Salvi et al., 2003; Roberts et al., 2010; Bubeck et al., 2015]. Due to the emission of high-frequency electromagnetic pulses of energy into the ground structures and strata layering is visualised by detected dielectric discontinuities
Figure 1.7: 270 MHz GPR system used in heavy terrain for hanging-wall investigations at the Lastros fault in 2013 (photo by Alexander Woywode).

[Neal, 2004; Schrott and Sass, 2008]. A vertical resolution in the order of a few centimetres enables to image and trace faults in GPR profiles [Chow et al., 2001; Reichert, 2001; Reichert et al., 2003]. Furthermore, Reiss et al. [2003] have shown that GPR profiling is suitable to visualise hanging-wall sedimentary structures related to recent coseismic deformation, such as coarse-grained colluvial wedges and displaced strata. When GPR profiles are arranged adjacent to each other a detailed pseudo-3D site survey determines precise information on position and orientation of dip-slip faults [Vanneste et al., 2008]. The subjectivity and the need for interpreter experience for correlating GPR data to complex geological arrangements are reduced by applying coherence-based and texture-based algorithms from computer vision to the datasets. This does not only provide objective and vivid visualisations of different radar facies, but also significantly improves the efficiency and quality of 3D GPR interpretations [McClymont et al., 2008].

The SIR-3000 GPR system by GSSI with different antenna models (100, 270, 400
1. Introduction

 MHz) optionally with survey wheel have been used to visualise the hanging-wall architecture (Fig. 1.7) at Kaparelli (central Greece) and Sfaka (Crete). The applied antennae frequencies allow objects smaller than 0.1 m in diameter to be resolved. Penetration depths are controlled by the physical properties of the ground. The conductivity of the materials is a main controlling parameter [Neal, 2004] and is inversely related to the achievable penetration depth. Hence, the presence of moisture or ground-water limits the penetration depth as well as conductive materials, such as clayey layers.

After data acquisition, raw data requires processing incorporating gain adjustments and filtering to optimize imaging. For this study Reflex® software [Sandmeier, 2015] has been used to process raw GPR data.

1.3.6 Terrestrial laser scanning (TLS)

LiDAR (light detection and ranging) scanning and ground-based terrestrial laser scanning (TLS) in particular have become more and more attractive in palaeoseismological studies over the past two decades. Laser scanning technologies are the preferred source for obtaining high-resolution digital surface data [e.g. McCaffrey et al., 2005; Buckley et al., 2008; Barth et al., 2012; Wilkinson et al., 2012; Meigs, 2013; Wilkinson et al., 2015]. The fundamental principle of laser scanning is rapid time-measuring of one-dimensional distances using the wave of light in a very narrow interval of the electromagnetic spec-
A coherent laser beam of low-divergence and linearly polarised propagates in a well-defined direction and is thus relatively well localised even at long ranges [e.g. Smith, 2015]. Changing the direction per measurement results in rectified true 3D point cloud data representing investigated natural surfaces as a digital version.

The application of laser scanning is widespread and does not follow standardized workflows. Where airborne LiDAR data is suitable for generating large bare-earth digital elevation models (DEM) in metre-scale resolution, terrestrial laser scanning offers the opportunity to collect surface data at centimetre to millimetre point cloud density for regional to very microscale applications. As a result, visualising fault-traces even in low-relief regions (Fig. 1.8) is possible [e.g. Cunningham et al., 2006; Arrowsmith and Zielke, 2009; Zielke et al., 2015] as well as imaging fault zone geometries and kinematics [e.g. Sagy et al., 2007; Sagy and Brodsky, 2009; Gold et al., 2012, 2013; Wilkinson et al., 2015; Mason et al., 2016]. It is important to emphasise laser scanning as a measuring technique that offers many different types of surface visualisations. Among hillshaded surface models, inferred surface characteristics such as aspect, slope or curvature allow quantitative analyses of geomorphological features documenting the seismic potential of earthquake-prone fault zones [e.g. Jones et al., 2009; Zielke and Arrowsmith, 2012; Schneiderwind et al., 2017].

For logistical reasons two terrestrial laser scanning systems were applied in this study. An Optech ILRIS 3D laser ranging system and a Faro Focus 3D X330 system were used for close- to mid-range scans of palaeoseismic trench and coastal cliff exposures. For data processing CloudCompare software [Girardeau-Montaut, 2016] and several Matlab® toolboxes [The Mathworks, 2016] were applied.

1.4 Guide through this thesis

This thesis summarises the results from field campaigns in 2013, 2014, 2015, and 2016. The text is mainly based on one published article, one article in press, and one submitted manuscript:


1. Introduction

Figure 1.9: A roadmap of the thesis. The source section of an arrow should be read before the
destination chapter.
Figure 1.9 presents a graphical roadmap depicting the organization of the thesis. Figures and results were updated according to the latest research results. Main focus of the work is set on three main topics: multispectral imaging of geological earthquake evidence in the shallow subsurface, identifying tidal notch morphologies on cliff faces using computer vision, and numerical modelling of tidal notch sequences. Therefore, generalised summaries on techniques used in here and palaeo-sea-level marker in palaeoseismological studies are provided in chapter 1. In chapter 2 a brief summary on geodynamic settings in the Mediterranean Basin is provided followed by geological description of the study areas in the eastern Corinthian Gulf. Chapter 3 deals with multispectral imaging combining terrestrial laser scanning (TLS) and ground-penetrating radar (GPR) on palaeoseismological trench exposures, and is based on Schneiderwind, S., Mason, J., Wiatr, T., Papanikolaou, I., and Reicherter, K. [2016]: 3-D visualisation of palaeoseismic trench stratigraphy and trench logging using terrestrial remote sensing and GPR – a multiparametric interpretation. In Solid Earth 7 (2), pp. 323-340. DOI: 10.5194/se-7-323-2016. TLS is also discussed in Chapter 4 based on Schneiderwind, S., Boulton, S.J., Papanikolaou, I., and Reicherter, K. [2017]: Innovative tidal notch detection using TLS and fuzzy logic: Implications for palaeo-shorelines from compressional (Crete) and extensional (Gulf of Corinth) tectonic settings. Geomorphology (in press). DOI: 10.1016/j.geomorph.2017.01.028. Here high-resolution surface data is analysed using fuzzy logic towards an enhanced identification of palaeo-shorelines. A new view on tidal notches as palaeoseismological archive is then presented in Chapter 5. The text is mainly based on a manuscript that has been submitted to JGR: Schneiderwind, S., Boulton, S.J., Papanikolaou, I., Kázmér, M., and Reicherter, K. [submitted]: Numerical Modelling of Tidal Notch Sequences on Rocky Coasts of the Mediterranean Basin. Date of submission to the Journal of Geophysical Research - Earth Surfaces: 2016-11-03. The implications of all the work done are discussed in chapter 6. A conclusion and an outlook for future work are provided in chapter 7.
Der Langsamste, der sein Ziel nicht aus den Augen verliert, geht immer noch
geschwinder als der, der ziellos umherirrt

The slowest person who never loses sight of his goal always goes faster than one who
wanders around aimlessly
G. E. Lessing (1729-1781)

The earth doesn’t move backward (very much) when you walk only because it’s much
more massive than you are
K. C. Cole
CHAPTER 2

Active tectonic setting and seismicity of the Mediterranean

In his seminal work McKenzie [1970, 1972] investigates what happens when continents collide in the Mediterranean region, and presents the importance of extensional and strike-slip tectonics within the overall convergence. For the Mediterranean basin three main stages can be observed: I) simple and narrow deformation at the oceanic plate boundaries of the Atlantic [Serpelloni et al., 2007], II) subduction of African oceanic crust under the Aegean plate in the eastern Mediterranean resulting in the broad Alpine belt of seismicity and deformation comprising complex pattern of crustal stress and strain fields [Vannucci et al., 2005], and III) large scale thrusting between Arabia and Iran with absence of oceanic lithosphere [McKenzie, 1972]. Due to the collision at considerable rates of up to 30-40 mm/yr [Kahle and Muller, 1998], active tectonics and frequent seismicity formed out diffuse over vast areas and elongated belts, clustering especially in territories of Italy, Greece, and Turkey (Fig. 1.2). Frequent low-moderate magnitude events and occasionally large magnitude earthquakes (M>7) characterise the seismicity of the basin [e.g. McKenzie, 1972; Jackson and McKenzie, 1988; Serpelloni et al., 2007]. The majority of earthquake events is of shallow origin and widespread in the whole Mediterranean, whereas deeper events mainly accumulate at the Betic-Rif, in the Tyrrenhian Sea, and at the Hellenic Arc.

2.1 Western Mediterranean Region

Oblique convergence between Africa and Eurasia is displayed at the Moroccan and Algerian coastlines accommodating compressional and strike-slip mechanisms. This style of active and seismic deformation propagates eastward from the southern margin of the Alboran and Algerian basin (~ 8 Ma) to the Tyrrenhian Sea (~ 2 Ma) [Goes et al., 2004; Billi et al., 2007; Serpelloni et al., 2007; Faccenna et al., 2014]. Representative for compressional historic events in the western part of the Mediterranean are the 1980 Mw7.1 El Asnam, and the 2003 Mw6.8 Zemmouri earthquakes in Algeria [Meghraoui et al., 2004],
and the 2004 Mw6.3 Al Hoceima event in Morocco [Stich et al., 2005].

2.2 Central Mediterranean Region

The central part of the Mediterranean basin is characterised by varying styles of active tectonics. As the compressional domain terminates northeast of Sicily [Billi et al., 2007], the Calabrian Arc and central-southern Apennine region is dominated by almost east-west directed extension [e.g. Stewart et al., 1997; Serpelloni et al., 2007]. The Apenninic mountain chain hosts a segmented system of large normal faults responsible for numerous destructive earthquakes during historical times. For instance, in 1908 a Mw7.1 earthquake struck the Messina Strait region causing more than 60,000 casualties [Valensise and Pantosti, 1992]. In 2009 a disastrous earthquake event (Mw6.3) causing 308 deaths occurred in central Italy. The earthquake triggered two other events with M>5 [Papanikolaou et al., 2010]. Active compression normal to fold-thrust belts is observed along the eastern boundary of the Adriatic, for the northern Apennines and the south-vergent southern Alps. Representative seismic moment release of Mw7.2 was recorded in 1979 in Montenegro [Kuk et al., 2000; Bennett et al., 2008].

2.3 Eastern Mediterranean Region

In the eastern Mediterranean region tectonic activity is rather complex and characterises one of the most intense seismicity in the world [Facenna et al., 2014]. Here, the Arabia Plate forces northwards into Eurasia pushing and rotating the Anatolia Plate westward and southwestward. This is indicated by strike-slip movements along the sinistral Dead Sea Transform (DST) and East Anatolian fault (EAF) as well as the dextral North Anatolian fault (NAF) (Fig. 2.1). Therefore, most western Turkey as well as Greece are located on the overriding part of a north-dipping subduction zone consuming African oceanic crust [Yeats, 2012]. This zone follows the Hellenic Arc and Cyprian Arc which extend from the Ionian Sea in the West to the Levant coast of Syria and Turkey in the East. Related to the Hellenic Subduction arc-parallel and arc-normal crustal extension takes place ranging from the southern Aegean forearc on the island of Crete as far north as Bulgaria. A well-recognised region of rapid continental extension poses the Corinth Rift zone with up to 10-15 mm/yr of N-S extension [Roberts et al., 2009].

As reported in long historical records crustal earthquakes, particularly those related to strike-slip and normal faulting, caused severe damages in historical and recent times. Strong and frequent earthquakes events occur along strike of the DST. At the junction with the EAF the largest earthquake during the last five centuries occurred in 1822 and destroyed 60% of Aleppo [Ambraseys, 2009; Yeats, 2012]. Plate motion yields to stress accumulation along the NAF. As a result, a sequence of eight M>7 earthquake events occurred subsequently moving west along strike of the fault; the last event (M7.4) occurred in Izmit, 100 km east of Istanbul [Hubert-Ferrari et al., 2000]. Among others, the 1981 M6.4-6.7 Alkyonides earthquake sequence (eastern Gulf of Corinth) is representative for normal faulting seismicity. In February and March, a set of three subsequent main shocks produced significant surface faulting (up to 1.4 m) [Jackson et al., 1982] on both sides of
the basin-bounding fault system including hanging-wall subsidence along the shore of the Perachora Peninsula [Hubert et al., 1996; Collier et al., 1998].

2.4 Geology of the study area

Fieldwork was predominantly undertaken in the eastern part of the Corinthian Gulf. The seismic landscape is well developed and preserves numerous primary and secondary coseismic earthquake geological effects. Additionally, the 1981 earthquake sequence is well documented and accompanied imprints to the landscape are still obvious.

2.4.1 Geodynamical setting

The Gulf of Corinth in central Greece extends to 115 x 30 km and depicts one of the fastest extending regions in the world (Fig. 2.2). The basin has a maximum water depth of 800 m at its centre. For this active zone, geodetic crustal deformation rates range up to 20 mm/yr in the western part decreasing down to 4 mm/yr in the most eastern parts [e.g. Billiris et al., 1991; Briole et al., 2000]. Regional extension is oriented almost
N-S and widely accepted as resulting from far-field extrusion of Anatolia and Aegea. Presumably, narrowing of an initially wide rift province was triggered by localisation of intra-plate strain in the eastern Gulf at 2.2 Ma. Subsequently, a migration northwards of fault activity onto the faults which define the present-day southern rift margin took place [Armijo et al., 1996; Goldsworthy and Jackson, 2001; Leeder et al., 2008; Ford et al., 2013; Duffy et al., 2015].

A series of 5-20 km long, E-W to WNW-ESE striking normal faults is accompanied by N-S extension across the Corinth rift. The southern shore is principally formed by three major north-dipping fault segments, namely the Elwiki fault zone, the Xylocastro fault zone, and the South Alkyonides Fault Zone (SAFZ) in the eastern part that ruptured in 1981 [e.g. Armijo et al., 1996; Roberts, 1996; Collier et al., 1998]. Contrarily, south-dipping faults forming the northern shoreline are smaller and antithetically arranged [e.g. King et al., 1985; Sakellariou et al., 2007; Bell et al., 2009; Taylor et al., 2011]. However, the third major event during the Alkyonides earthquake sequence on 1981-03-04 occurred on the antithetic Kaparelli fault [e.g. Stiros et al., 2007]. Perachora Peninsula is situated...
Figure 2.3: Simplified geological map of the eastern Parts of the Corinthian Gulf [after Bornovas et al., 1984b,a]. Off-shore faults are from Sakellariou et al. [2007].
east of the Isthmus at the southern margin of the Alkyonides Gulf. The landscape is formed by step-faulting combining offshore NW to NE-dipping, coast parallel faults, SW-dipping faults at the southwestern shoreline, and almost N-dipping onshore faults being part of the SAFZ.

2.4.2 Geological setting

Today, extension occurs in an area where the continental crust was previously thickened as an alpine fold-thrust belt. Therefore, the geology of central Greece is characterised by NNW-SSE striking tectonic units (Fig. 2.2) that result from the complex west-vergent stacking of shallow water carbonates during the closure of the Tethys. In between these marine rocks pelagic sediments and volcano-sedimentary formations including mafic rocks occur [Papanikolaou, 2009]. In detail, nine distinct sedimentary facies belt ('tectonostratigraphic terranes') pose individual thrust sheets, which can be categorised in two main groups: I) continental terranes comprising pre-Alpine crustal basement lithology covered by (usually detached) shallow-water carbonate platforms, and II) oceanic terranes comprising basin sediments overlying ophiolites [Papanikolaou, 2013]. The western external Hellenides are mainly made of carbonate-series comprising Mesozoic to upper Eocene neritic limestones (Gavrovo-Tripolis Zone; external carbonate platform) overlain by Eocene to Oligocene flysch and Mesozoic pelagic sediments (Pindos Zone; Pindos/Cyclades oceanic basin). The lithology in the eastern Gulf of Corinth includes crystalline limestones, flysch units and Neogene-recent basin deposits (Fig. 2.3).

A description of the stratigraphy from old to young starts with lower-middle Triassic metamorphic schists which underlie all other units. Very thick sequence of marine nappes of the Boeotian zone (~800-1,000 m) is located above, consisting of deep sea flysch deposits, ultrabasic rocks in a schist-chert formation, and limestones of Triassic to Oligocene age [Bornovas et al., 1984a]. Miocene-Oligocene clastic sediments lie on top of the pre-rift stratigraphic block. The age of extension in the Gulf of Corinth is controversial, but certainly includes Pliocene-Pleistocene syn-rift sediments. This is indicated by Pleistocene and younger alluvium which are present in depressions and along the coasts.
The past is the key to the future

L. Serva
CHAPTER 3

Multiparametric interpretation of palaeoseismological trench exposures

Two normal faults on the Island of Crete and mainland Greece were studied to test an innovative workflow with the goal of obtaining a more objective palaeoseismic trench log, and a 3D view of the sedimentary architecture within the trench walls. Sedimentary feature geometries in palaeoseismic trenches are related to palaeoearthquake magnitudes which are used in seismic hazard assessments. If the geometry of these sedimentary features can be more representatively measured, seismic hazard assessments can be improved. In this study more representative measurements of sedimentary features are achieved by combining classical palaeoseismic trenching techniques with multispectral approaches. A conventional trench log was firstly compared to results of iso cluster analysis of a true colour photomosaic representing the spectrum of visible light. Photomosaic acquisition disadvantages (e.g. illumination) were addressed by complementing the dataset with active near-infrared backscatter signal image from TLS measurements. The multispectral analysis shows that distinct layers can be identified and it compares well with the conventional trench log. According to this, a distinction of adjacent stratigraphic units was enabled by their particular multispectral composition signature. Based on the trench log, a 3D-interpretation of attached 2D GPR profiles collected on the vertical trench wall was then possible. This is highly beneficial for measuring representative layer thicknesses, displacements and geometries at depth within the trench wall. Thus, misinterpretation due to cutting effects is minimised. This manuscript combines multiparametric approaches and shows: (i) how a 3D visualisation of palaeoseismic trench stratigraphy and logging can be accomplished by combining TLS and GRP techniques, and (ii) how a multispectral digital analysis can offer additional advantages to interpret palaeoseismic and stratigraphic data. The multispectral datasets are stored allowing unbiased input for future (re-)investigations.
3. Imaging a trench

3.1 Why visualise trench exposures

Seismic hazard assessment is still predominantly based on the instrumental and historical catalogues of seismicity. However, these catalogues are generally too short compared to the recurrence interval of particular faults [e.g. Wesnousky, 1986; Yeats and Prentice, 1996; Machette, 2000]. As a result, the sample from the statistical elaboration of the historical and instrumental data is incomplete and a large number of faults would have not ruptured during the period in which the historical record is considered complete [Stucchi et al., 2004; Woessner and Wiener, 2005; Guidoboni and Ebel, 2009; Grützner et al., 2013; Stucchi et al., 2013; Papanikolaou et al., 2015]. The need for fault specific studies and the extraction of recurrence intervals from palaeoseismological trenches was then initiated in the late 1970s [Sieh, 1978; McCalpin, 2009]. The goal is to extend the history of slip on a fault back many thousands of years, a time span that generally encompasses a large number of earthquake cycles [Yeats and Prentice, 1996].

Over the last few years fault specific studies and palaeoseismology have been further advanced and are now supported by new remote sensing tools that offer high spatial resolution (e.g. TLS) and geophysics that extend our data into the subsurface (Ground Penetration Radar (GPR), Electric Resistivity Tomography (ERT)) [Papanikolaou et al., 2015]. This chapter adds on such approaches and shows: (i) how a 3D visualisation of palaeoseismic trench stratigraphy and logging can be accomplished by combining TLS and GRP techniques, and (ii) how a multispectral digital analysis can offer additional advantages and a higher objectivity in trench data interpretation.

Palaeoseismological studies are often undertaken to identify earthquake recurrence intervals and maximum credible magnitudes of prehistoric earthquakes [McCalpin, 2009]. These parameters are needed for the accurate calculation of seismic hazard potential of active fault zones [Reichertz et al., 2009]. Evidence for palaeoearthquakes can be found within the sedimentary architecture of active faults where conditions are favourable for their preservation. Typical features caused by recurrent seismic events include: (i) progressive displacements [Keller and Rockwell, 1984], (ii) coluvial wedges, (iii) liquefaction, and (iv) fissure fills [Reichertz et al., 2003; Kokkalas et al., 2007; McCalpin, 2009] (see Fig. 3.1 f). The geometry and stratigraphic position of these features allow a retrodeformation of recurrent surface rupturing events, whereas carbon rich material can be used to date prehistoric earthquakes and determine recurrence intervals. To access these potential archives of seismic information trenches, which are often expensive, are excavated across deformation zones. Then, the classical approach is to document stratigraphy and structure by careful logging, either on paper and/or with photographs [e.g. Wallace, 1986; McCalpin, 2009]. The accuracy of the trench log is dependent on the logger’s experience and ability to define units of discrete deposits that have distinguishable lithological characteristics compared with adjacent deposits.

Palaeoseismic indicators are widely spread and their formation varies along fault strike [e.g. Bubeck et al., 2015]. For this reason, geophysical surveys undertaken prior to the trenching phase have become common practice over the last decade. For instance, ground-penetrating radar (GPR) measurements have been carried out to identify optimum trenching locations [e.g. Demanet et al., 2001; Alas et and Meghraoui, 2005; Grützner et al., 2012] and many studies have shown that earthquake related structures can be identified in the shallow subsurface with geophysics [e.g. Chow et al., 2001; Reiss et al., 2003; Bubeck
Figure 3.1: Guide to the study area. a) Map of Greece showing simplified large-scale tectonic structures (CG, Corinthian Gulf; CF, Cephalonia fault; NAF, North Anatolian fault; NAT, North Anatolian Trough; black lines with bars show active thrusts; black lines with marks show active faults) [after Kokkalas and Koukouvelas, 2005; Papanikolaou and Royden, 2007]. White boxes highlight study areas. b) Satellite image (Landsat 8, 2015) of the easternmost Gulf of Corinth. The Kaparelli fault is shown in red and the white box marks the position of the palaeoseismological trench of Kokkalas et al. [2007]. c) View of the Kaparelli trench. d) Satellite image (Landsat 8, 2015) of the study area at the Sfaka fault (red) in northeastern Crete; the white box shows the position of the road cut along strike. e) View of the Sfaka road cut. f) Sketch of a typical postglacial normal fault showing bedrock juxtaposed against Quaternary sediments which contain structures caused by recurrent earthquakes [modified after Reichert et al., 2003]. Colluvial wedges form at the base of the fault scarp from eroded material originating at the top of the scarp.
et al., 2015]. The excavated trench wall is then a 2D representation of the fault zone stratigraphy. It is assumed that the 2D geometry of the logged sedimentary features continues along strike either side of the trench; without widening the trench along strike, or excavating more trenches, we must assume that the 2D trench log is representative for this location along the fault. Hence, an interpretation of a 2D exposure of very local variations and/or accumulations of colluvial deposits yield results different from statistical significance which gets closer to the real world conditions. Trenches target predominantly palaeosols on either side of the fault, and then according to empirical relationships [Wells and Coppersmith, 1994] palaeomagnitudes can be estimated based on these co-seismic displacements. If no or only poorly expressed displaced palaeosols exist the geometry of sedimentary features within trenches is used to estimate previous earthquake displacements. As a "rule of thumb" colluvial wedge thickness equals half of the initial scarp height [e.g. Reichert, 2001; Reiss et al., 2003; McCalpin, 2009]. Such information are then used as input parameters for seismic hazard assessment. Therefore, tracing the geometry of these features is essential for the most accurate seismic hazard calculations. A better visualisation can improve the definition of separate unit boundaries and features, offering better interpretations and limiting uncertainties.

This chapter demonstrates how high-resolution TLS measurements and photomosaics can be used to assist in the interpretation of palaeoseismological exposures; we also show how an accurately arranged 2D GPR survey can assist to visualise sedimentary structures in 3D [e.g. Vanneste et al., 2008; Christie et al., 2009; Ercoli et al., 2013] within the trench wall. The TLS’s backscatter signal represents material reflectance of radiation in the near-infrared wavelength, and digital photo cameras collect information of the reflectance of visible light; therefore, a quasi-multispectral inspection of the exposures is possible. Ragona et al. [2006] developed a method using imaging spectroscopy on palaeoseismic exposures with hyperspectral and normal digital cameras. As an outcome they were able to enhance the visualisation of the sedimentary layers and other features that are not obvious or even not visible to the human eye. Another study undertaken by Wiatr et al. [2015] places emphasis on the use of the monochromatic laser beam’s backscattered signal to determine varying surface conditions. Using these techniques we assist experienced-based trench logging and obtain 3D spectral data to support the interpretation of palaeoseismological deposits. Two-dimensional GPR surveys, arranged for a pseudo-3D cube reconstruction, undertaken on top of the trench and on the vertical trench wall (Fig. 3.2) are used in combination with a high-resolution digital elevation model (DEM) from TLS scanning. This allows radar facies [Neal, 2004] to be distinguished and the sedimentological architecture at depth within the trench wall to be identified. Thus, the resulting 3D model from the GPR provides information on varying layer thicknesses and minimises misinterpretation due to cutting effects. The workflow comprising data acquisition, statistical analysis, interpretation and storage (Fig. 3.3) was calibrated on a road cut on the Island of Crete. Afterwards, the workflow is applied on a professionally excavated trench in mainland Greece.
Figure 3.2: A simplified model of investigated parts on footwall, scarp, hanging-wall and trench at both exposures of this study; visualisation shows the conditions at the Sfaka road cut. Dashed lines show the different workspaces: I) red, Overall workspace for a long-mid range TLS scan to retrieve the geometric relation of investigated components; II) blue, area of operations (log, photo, TLS, GPR) on the trench wall; III) green, workspace for GPR measurements (black arrows) on top of the colluvium.
3. Imaging a trench

![Diagram of data processing steps](image)

**Figure 3.3:** Flowchart of conventional trench logging, imaging spectroscopy and GPR survey; their comparison and combination. Palaeoseismological studies benefit from one or multiple datasets (grey) generated with this workflow.

### 3.2 Geological setting of the trenching sites

The study sites are both located in Greece, which is one of the most seismically active parts of the Mediterranean [McKenzie, 1972; Le Pichon and Angelier, 1979; Papazachos et al., 2000] due to the presence of the Hellenic Arc and Trench System. Crustal extension orientated both arc-parallel and arc-perpendicular [Mariolakos and Papanikolaou, 1981; Lyon-Caen et al., 1988] has led to the development of Quaternary carbonate bedrock fault scarps throughout both mainland Greece [Stewart and Hancock, 1991; Benedetti et al., 2002] and the island of Crete [e.g. Gaki-Papanastassiou et al., 2009]. These normal faults mainly consist of footwall Mesozoic carbonates juxtaposed against hanging-wall flysch and/or post-alpine sediments. Earthquake features such as colluvial wedges (a consequence of degradation of the scarp), fissure fills and displaced strata occur within the hanging-walls of these faults and datable material may be contained within buried palaeosols (see Fig. 3.1 f) [McCalpin, 2009]. To create those archives and preserve them over geological timescales, erosional processes must be lower than the rate of tectonic activity. These features therefore represent geological archives of palaeoearthquakes because they can record information about Holocene and Late Pleistocene earthquakes [e.g. McCalpin, 2009]. Ambraseys and Jackson [1990] estimate a maximum earthquake Magnitude of Ms 7.0 could occur on these normal faults using macroseismic and instrumental data, which coincides with fault segment lengths of 15 – 30 km [Wells and Coppersmith, 1994].
3.2.1 The Sfaka fault (NE Crete, Greece)

The island of Crete is the largest within the Greek territory and is directly adjacent to the subduction zone between Europe and Africa. The NNE-SSW trending Sfaka fault is located in northeastern Crete (Fig. 3.1 a) and forms the easternmost segment within the Ierapetra Fault Zone which is a major tectonic line of approximately 25 km cutting through the whole island [Gaki-Papanastassiou et al., 2009]. This northwest dipping normal fault is easy to recognize as a prominent fault scarp of up to 6 m. The scarp dips 70° towards the West and offsets smooth mountain slopes for approximately 5 km onshore (Fig. 3.1 d). Together with the opposing Lastros fault a 2 km wide graben structure is formed.

An outcrop in the form of a dirt road cut (located at 35°7’58.97"N, 25°54’26.01"E) exhibits the fault zone as a contact between footwall Mesozoic carbonates and hanging-wall colluvium (Fig. 3.1 e). The outcrop cuts the fault at an angle of approximately 75° from the fault strike.

3.2.2 The Kaparelli Fault (Gulf of Corinth, Greece)

The Kaparelli fault is located in the easternmost part of the Gulf of Corinth (see Fig. 3.1 a) which is associated with rapid extension oriented N-S [e.g. Papanikolaou and Royden, 2007]. The Kaparelli fault became well-known as it ruptured during the 1981 Corinthian Alkyonides earthquake sequence in February (24th, Ms6.7, depth: 10 km; 25th, Ms6.4, depth: 8 km) and March (4th, Ms6.4, depth: 8 km) [Jackson et al., 1982]. Many palaeoseismological studies using various approaches have been undertaken along this ca. 20 km long south dipping normal fault. For example Benedetti et al. [2003] used 36Cl cosmic ray exposure dating to determine the history of surface rupturing events on the 4-5 m high limestone scarp of the Kaparelli fault. Their results show evidence for seismic activity 20 ± 3 ka, 14.5 ± 0.5 ka and 10.5 ± 0.5 ka prior to the 1981 earthquake sequence. A palaeoseismological trenching study was conducted by Kokkalas et al. [2007]. The authors found evidence for at least three events in the past 10,000 years: 7370 ± 120 a, 7290 ± 140 a to 5640 ± 70 a, and 1225 ± 165 a. The excavations from Kokkalas et al. [2007] are still open; therefore, the already logged and interpreted structures within trench Kap-1 (Fig. 3.1 e) represent a perfect site to test remote sensing data acquisition.

3.3 Visualising the trench

The herein presented workflow comprises (i) a combination of conventional trench logging and remote sensing measurements, (ii) a comparison of common photographs and near-infrared images, and (iii) a GPR survey (Fig. 3.3). It combines palaeoseismic trenching techniques with TLS measurements to improve the accuracy of palaeoearthquake reconstruction. A multispectral analysis of TLS backscatter data and the luminescence of true colour photographs were compared to the manual trench log. A GPR survey was then conducted to obtain 3D information of layer continuation and thickness at depth within the trench wall (Fig. 3.2).
3. Imaging a trench

3.3.1 Conventional trench logging and photomosaic

A palaeoseismic trench is characterised by the subsurface exposure of fault zones and deformed stratigraphy. To accurately interpret these features, apparent dips and anthropogenic and/or exogenous influences must be excluded. Moreover, sketching lithological contents requires an exposure devoid of weathered and smeared parts that were caused by the excavation [McCalpin, 2009]. To simplify and prove the geometrical correctness of the trench log, a reference grid of one square metre was attached to the wall. The grid’s points of intersection also act as reference points for remote sensing applications [Reitman et al., 2015].

The trenches were logged in 1:10 scale. Thereby, discrete deposits that are composed of similar lithology considering consistent texture, sorting, bedding, fabric, and colour of individual layers are mapped. Photographs of every square metre were taken and later stitched together using an automatic panorama recognising tool including a manual editor of control points and straightening functions (Autopano Giga, Kolor). It must be noted that error values are already stored within image information due to differing luminous exposures; furthermore, holes and protruding boulders create shadows that partially change the reflection characteristics of certain sedimentological features. The Sfaka road cut faces north (see Fig. 3.1 d and e) and is surrounded by steep slopes. Since footwall and hanging-wall deformation structures are exposed the outcrop is a suitable palaeoseimological trench after manual levelling and deepening of most interesting parts. In Kaparelli the eastern trench wall (see Fig. 3.1 b and c) was investigated because it preserved the best stratigraphy and exhibits clear horizons of multiple faulting events [Kokkalas et al., 2007]. To avoid most of the differing luminous exposures, the photographs were either taken in the morning when the angle of sunlight was shallow and did not shine directly onto the investigated wall (Kaparelli trench) or in the afternoon when the sun disappeared behind the surrounding hills (Sfaka road cut).

The photomosaic of true colour images (RGB method) was converted into a grey-level image to eliminate hue and saturation information while retaining the luminance (0-255) using the rgb2gray function in MATLAB®. In a GIS the resulting image was georeferenced to a custom frame in order to make it comparable to all other datasets of this study.

3.3.2 TLS measurements

TLS is a remote sensing technique with high spatial and temporal resolution and is a very effective instrument for reconstructing morphology [Brochu and Lague, 2012; Wilkinson et al., 2015], geological settings and monitoring movements [Jones, 2006; Hu et al., 2012]. In seismic hazard assessment, this technology assists fault mapping [e.g. Arrowsmith and Zielke, 2009; Begg and Mouslopoulou, 2010] as well as providing a tool to trace palaeoevents based on changes in reflectivity and roughness on fault scarp [Wiatr et al., 2015]. A generated coherent laser beam with little divergence by stimulated emission is reflected off surfaces and the proportionate backscattered signal is detected, forming a non-contact and non-penetrative active and stationary recording system. Thus, from measuring the two-way-travel time (TWT) of a first pulse detection sequence, 3D surface data is acquired. The illuminated area is controlled by wavelength, beam divergence,
range between sensor and target, and also by the angle of incidence [Jörg et al., 2006; Wiatr et al., 2015]. In our study we used an ILRIS 3D laser ranging system (wavelength \( \lambda \) is 1,500 nm) from OPTECH Inc., Ontario, Canada.

The limitations of using TLS are high humidity [e.g. Lobell and Asner, 2002] and low target reflection with cumulative distance and shallow incident angle [e.g. Höfe and Pfeifer, 2007]. In order to assume constant soil moisture and to ensure the backscatter signal data quality, close range scans were done during the summer in dry conditions within a few hours. The scans were carried out almost perpendicular to the trench wall. Since the Kaparelli trench is too narrow for scans from inside, the data was collected from outside of the excavation. At the Sfaka fault road cut, scans were undertaken at 5 m distance.

Other benefits of applying TLS is its flexibility, the relatively quick availability of an actual dataset, and also its high spatial resolution with information about backscatter signal each referenced in x,y,z-coordinates. The result is an irregular but dense point cloud representing a highly detailed digital 3D surface model which can be easily implemented in geographical information systems (GIS) to generate accurate digital elevation models (DEM) or digital terrain models (DTM) [e.g. Wiatr et al., 2015].

For this study, the TLS scanning was undertaken at both close range and long-mid range to determine geometrical relationships between the footwall, hanging-wall, prolongation of the scarp, and trench wall (see Fig. 3.2). The backscatter signal of the TLS results from the reflection of transmitted waves of near-infrared light. In other words, each measurement is usually accompanied by a surface remission value, which quantifies the intensity of the reflected laser beam. The monochromatic backscatter signal values are stored as grayscale values from 0-255. The information on the monochromatic wavelength and the detected backscattered signal in the near-infrared reflects the surface properties which are invisible to the human eye. Thus, the backscatter signal was also used for the multispectral analysis. The raw-data was cleaned from isolated points and those that do not represent the area of interest. The mathematical and geometrical alignment of the different scan windows was then carried out. For project specific demands, the datasets were translated into a custom grid. The long-mid range data were used for the overall geometrical analysis creating high-resolution DEMs with a resolution smaller than 0.1 m. Data from the close range scan were processed for statistical calculations of the backscatter signal's spatial distribution. A detailed description on the applied workflow is given in Sect. 3.3.3.

3.3.3 Imaging Spectroscopy

Visualising an array of simultaneously acquired images that record separate wavelength intervals or bands is part of multispectral analyses. A common multispectral camera employs a range of film and filter combinations to acquire photographs that record narrow spectral bands of non-imaging data. Reflectance spectra map the percentage of incident energy (e.g. sunlight) that is reflected by a material as a function of energy wavelength. Absorption of incident energy is represented by downward excursions of a curve (absorption features). Upward excursions represent superior reflectance (reflectance peaks). These features are valuable clues for recognising and distinguishing certain materi-
3. Imaging a trench

als [Sabins, 1997]. Multispectral imaging, or imaging spectroscopy, has been used at many different scales for remote sensing. Probably the most prominent example for macro-scale investigations is the inspection of visible and near-infrared satellite imagery for mapping and monitoring vegetation [e.g. Tucker, 1979]. Ragona et al. [2006] introduced an application of high-resolution field imaging spectroscopy on palaeoseismic exposures using hyperspectral and common digital photo cameras. The authors conclude that imaging spectroscopy can be successfully applied to assist in the description and interpretation of palaeoseismic exposures because: (i) subtle or invisible features are displayed, (ii) quantitative analysis and comparisons of units using reflectance spectra can be undertaken, and (iii) unbiased data are stored for future access and analysis.

The limitations of multispectral approaches are, by their nature, closely connected to the application of photomosaics and TLS measurements. It is reasonable to re-emphasise the influence of moisture; where present it not only causes a darkening of the sediments (reduction in reflectance), but there is also a hard-to-quantify content variation across the exposure [Ragona et al., 2006]. Another error source appears due to morphological characteristics of a certain exposure, especially on surfaces that are not well prepared for palaeoseismic investigations and data collection. This means that the exposure must be flattened and cleaned to avoid changes in spectral amplitudes accompanying changes in illumination angle and distance. We assume that the moisture content was similar throughout the exposure and water absorptions should not affect the correlations because the spectral change is similar along the trench wall. Furthermore, the photos and TLS scans were taken almost perpendicular to the exposure so that optimal data quality can be expected.

The workflow contains geo-referencing and snapping the high-resolution raster data from the photomosaic and TLS backscatter signal to a coherent cell size (0.001 m) in GIS. Afterwards, an iso (iterative self-organising) cluster unsupervised classification was applied to a two-channel composition of both raster layers. Thereby, the number of classes was set to ten times the amount of included bands (photomosaic grey-level image and TLS backscatter signal image) as this provides sufficient statistics and enough cells to accurately represent a certain cluster. This type of clustering uses a process in which all samples are assigned to existing cluster centres during each iteration; new means are then recalculated for every class. The actual number of classes is usually unknown; therefore, it was initially set to 20 classes and the attribute distances between sequentially merged classes were analysed with the dendrogram method (hierarchical clustering). This reduces statistical misclassifications and provides information on distinct classes. Based on the outcome, classes which are statistically closest get merged and the dataset gets reclassified. Block statistics within a 3x3 cell environment are applied to erase noise by overwriting cell values to all of the cells in each block with the median value (Fig. 3.4). Moreover, resampling down to 0.02 m cells enhances visibility and allows a more general interpretation and comparison to the conventional log. This is because the scattered signal gets reduced so there is less influence by local variations. The threshold is chosen because average gridding and sketching inaccuracy is around 2% [McCAlpin, 2009].

48
Figure 3.4: An illustration showing how spatial information gets down sampled. The median value of surrounding cells provides the new cell \((x,y)\) value \((\lambda)\).
3. Imaging a trench

3.3.4 Ground-Penetrating Radar

GPR is a non-invasive and non-destructive geophysical technique that operates with high-frequency electromagnetic waves in the radio band to detect electrical discontinuities in the shallow subsurface up to approximately 50 m. Every GPR measurement contains a five-step process of: (i) generating, (ii) transmitting, (iii) propagating, (iv) reflecting, and (v) receiving electromagnetic pulses. The differing relative dielectric permittivities ($\varepsilon_r$) of varying materials control the transmitting velocity in relation to the speed of light ($c = 0.2998 \text{ m/ns}$) once the pulse is emitted from the antenna. Fractional reflections of the pulse on inhomogeneities and layer boundaries get received due to a dielectric contrast. In order to calculate depths of reflection the TWT (two way travel-time) is recorded in the order of nanoseconds. Depending on the frequency of the antenna, objects smaller than 0.1 m in diameter can be resolved. Common GPR systems perform at frequencies between 50 MHz and 1 GHz, where achievable resolution is a quarter of the wavelength. The relationship between penetration depth and spatial resolution is an inverse one; hence, a higher spatial resolution occurs at the expense of penetration depth and vice versa [Neal, 2004; Schrott and Sass, 2008].

Water is almost the only limiting parameter for the application of GPR because of its high relative dielectric permittivity. Moisture content dramatically decreases the electromagnetic wave velocity by stronger attenuation and leads to reduced penetration depths [Schrott and Sass, 2008]. Soil moisture differences often severely disrupt wave energy, which makes it even more difficult to interpret reflections. Dielectric contrasts are the main features of the GPR image interpretation, since any dielectric discontinuity is detected. Thus, targets can be classified according to their geometry and reflection facies.

GPR was carried out on the vertical trench wall and on the slope surface above the trench (see Fig. 3.2). At the Sfaka fault road cut three horizontal profiles were collected on the vertical exposure with 0.3 m spacing between profiles. 15 profiles were undertaken on top of the trench in a grid array to obtain a high resolution pseudo-3D cube. At the Kaparelli trench 20 profiles were collected on the vertical trench wall, and 14 on top of the hanging-wall. In order to make the GPR operationally effective, our survey provided efficient coupling of the antenna to the ground and a sufficiently large scattered electromagnetic signal for detection at or above the ground surface. A 400 MHz antenna together with a SIR-3000 control unit from Geophysical Survey Systems Inc. [GSSI, Salem, NH, USA] was used to obtain desired resolution and noise levels. The data processing was done using the software ReflexW® [Sandmeier, 2015] involving the following processing sequence: remove header gain, move start time, energy decay, 1D bandpass frequency, background removal, and average xy. Reflection hyperbolas of gravels were used to estimate wave velocity. Data migration was undertaken to correct angles, because dips are usually underestimated due to a complex 3D cone in which electromagnetic energy radiates [Neal, 2004].

Based on the layers distinguished in the trench log and taking into account the results of the multi-spectral analysis, GPR-data were then used to interpret the outcropping strata in 3D.
3. Results

3.4 Sfaka Fault, Crete

3.4.1 Trench Log

The trench was logged and ten distinct layers were recognised. These vary in colour, matrix, and geometrical alignments. The trench exposes the limestone footwall at its eastern end between 0 and 1 m. The limestone is heavily weathered and degraded, both within and above the trench. Adjacent to the bedrock fault plane is fault gouge which is approximately 1 m in thickness. However, true thickness is calculated to around 0.8 m when correcting for the trench’s 75° from fault strike. The western end of the gouge is the primary fault contact. Here, the clasts within the gouge are aligned vertically and there is an abrupt contact to the next units. These units are interpreted as fissures filled with palaeosols (Fig. 3.5 a). Palaeosol 1 comprises light brown to reddish brown very gravelly silty clay with occasional cobbles and containing roots and rootlets, and Palaeosol 2 comprises light brown to brown gravelly clay containing rare cobbles. Both these palaeosols have high clay contents and there is a sharp contact with the colluvial layers further to the west. The remaining sediments within the trench are colluvial deposits C1 to C6. C1 is cemented colluvium located at the western end of the trench. C2 to C6 are individual colluvial layers which can be traced from the cemented colluvium to the fissure fills. These colluvial layers are offset by a number of small displacement secondary faults. These minor faults are typical of extension in unconsolidated sediments.

The trench is not dominated by scarp derived colluvial wedges formed after rupturing events. Instead earthquake evidence comes in the form of fissure fills which have developed within the hanging-wall adjacent to the fault gouge (Fig. 3.5 b). These fissure fills are filled with palaeosols and are faulted against colluvial material which is partly scarp derived and partly hanging-wall derived. Due to the nature of the sloping hanging-wall and the location of both trenches, we believe that the main source of colluvial layers C2 to C6 is hanging-wall colluvium from the south at higher elevations. This is also evidenced by the alluvial/colluvial fan located 85 m to the west of the trench. Two displacement events can be inferred based on fissure fill and colluvial stratigraphy. Dip slip faulting causes the hanging-wall to be downthrown and tilted; due to a slightly concave fault plane below the trench site, a tectonic fissure then opens up between the fault gouge and colluvial layers, and tilting is taken up on the small displacement antithetic faults within the colluvial layers. The fissure was then filled with scarp derived and local hanging-wall material. The slope surface then stabilises allowing gravelly topsoil to accumulate. The second displacement event then occurs and the above described process is repeated.

3.4.1.2 Imaging Spectroscopy Analysis

The greyscale photomosaic stores visual impressions in a way similar to the human eye and represents a weighted sum value of luminance within the range of visible light per pixel. Luminance at 1,500 nm detected by TLS significantly differs in some parts of the trench wall (Fig. 3.5 b+c). As shown in Table 3.1, the light fault gouge material is highly reflective in both photomosaic and high resolution digital backscatter model (HRDBSM).
3. Imaging a trench

Figure 3.5: See facing page.
Table 3.1: Median greyscale values of photomosaic, high-resolution digital backscatter model (HRDBSM) and 2-component composition per stratigraphic unit from the trench log. The composition is the result of allocation of photomosaic and HRDBSM in equal parts, to visualise certainties and their variation within given zones. Error is given by single standard deviation.

<table>
<thead>
<tr>
<th>Layer</th>
<th>Photomosaic</th>
<th>HRDBSM</th>
<th>Composition</th>
</tr>
</thead>
<tbody>
<tr>
<td>recent topsoil</td>
<td>132 ± 21</td>
<td>197 ± 12</td>
<td>138 ± 23</td>
</tr>
<tr>
<td>fault gouge</td>
<td>221 ± 19</td>
<td>239 ± 12</td>
<td>224 ± 23</td>
</tr>
<tr>
<td>palaeosol I</td>
<td>165 ± 19</td>
<td>170 ± 13</td>
<td>87 ± 23</td>
</tr>
<tr>
<td>palaeosol II</td>
<td>156 ± 19</td>
<td>192 ± 10</td>
<td>128 ± 21</td>
</tr>
<tr>
<td>C6</td>
<td>131 ± 19</td>
<td>178 ± 10</td>
<td>99 ± 21</td>
</tr>
<tr>
<td>C5</td>
<td>152 ± 21</td>
<td>199 ± 11</td>
<td>142 ± 21</td>
</tr>
<tr>
<td>C4</td>
<td>143 ± 22</td>
<td>198 ± 10</td>
<td>140 ± 21</td>
</tr>
<tr>
<td>C3</td>
<td>171 ± 19</td>
<td>200 ± 12</td>
<td>144 ± 24</td>
</tr>
<tr>
<td>C2</td>
<td>147 ± 22</td>
<td>186 ± 13</td>
<td>116 ± 25</td>
</tr>
<tr>
<td>C1</td>
<td>144 ± 18</td>
<td>199 ± 15</td>
<td>144 ± 30</td>
</tr>
<tr>
<td>Boulder</td>
<td>168 ± 26</td>
<td>177 ± 15</td>
<td>102 ± 26</td>
</tr>
</tbody>
</table>

The homogeneous silty layer (fault gouge) contains only a few voids due to excavation works that influence reflectance value range. Resultant colorimetric shift expressed by the 2-component composition almost solely depicts the highest value ranges for this part of the trench wall (Fig. 3.5 d). In contrast, the cemented colluvium to the west is highly irregular in the sense of reflectance. Both the photomosaic and HRDBSM show a heterogeneous greyscale value distribution that is even more embodied by high-grade contrasts in the two-channel composition. Similar observations occur for larger boulders that protrude out of the trench wall (Fig. 3.5 a-d).

Colluvial layers C2-C5 are distinctively different in their reflectance characteristics. Where transition between both units is indeed visible in the photomosaic, a sharp contrast in reflectance characteristics of near-infrared is recognisable. Moreover, the named colluvial deposits do not only appear as a collection of diffuse values but show evidence of alignments. An upward oriented structure of approximately 0.5 m thickness is obvious in the HRDBSM and false colour composition. The structure follows a secondary fault within the colluvial strata.

Figure 3.5: Compilation of analytical input and outcome at the Sfaka fault road cut. a) Trench log produced in the field and corrected with b) Photomosaic in the office. c) High-resolution digital backscatter model (HRDBSM) from TLS measurements. d) two-channel composition from b) and c). Note, green and red are 100 percent different. (PM = photomosaic). e) Visualisation of spatial distribution of seven classes from the unsupervised classification. White dashed lines indicate coinciding arrangements and some influence from daylight.
Figure 3.6 visualises percentages of seven classes, estimated from the unsupervised classification on individual identified layers within the trench log. Either the majority of a certain layer is fulfilled by one single class or by a certain composition of two or three classes. Where Table 3.1 shows the dominance of high values within the fault gouge layer, the illustration of unsupervised classification proves this layer to be almost completely (70%) represented by one single class (7). Although class 4 covers almost the same value range as class 7, fault gouge exposure is only covered by 8% by class 4 (see Fig. 3.6 a).

In the unsupervised classification, the fault gouge is the only layer in this trench wall where the majority is covered by one single class. Palaeosol I and palaeosol II have a similar ratio of effecting classes but class 6 is not present in the palaeosol II signature, allowing them to be differentiated. By visualising the spatial arrangement of influencing classes the differentiation between these two layers is even better (Fig. 3.5 e). While Figure 3.6b only shows percentage significance of class ratios per layer, the spatial distribution promotes the reconstruction of a certain layer. The accumulation of class 5 especially in the lower part of palaeosol I is obviously different from any other cluster in palaeosol II, although quantitative statistics conclude a similar composition of classes.

Except for C6, which is well represented to around 80% by class 1 (31%) and 2 (52%), and C1, which appears as an unsorted collection of classified responses, the remaining colluvial lithologies appear with similar ratios, especially classes 1, 2, and 5. In a quantitative way no distinction can be recognised. Also, large scale clustering of classes within the layers is absent. However, arrangements, especially of class 7, are obvious and coincide with coarse-grained gravels within the colluvium. Within C3, a micro-cluster of approximately 25 pixels are arranged along a slightly bent line dipping about 50° towards the footwall. A similar arrangement of class 7 with an even smaller cluster (3x3 pixels) and wider spread is indicated in C5, dipping 15° towards the footwall. Furthermore, the surrounding matrix is slightly more expressed by class 5 in C5, whereas C3 has subjectively no preferred matrix content (Fig. 3.5). Alterations are expected to decrease with increasing depth. Dependent on rock composition and mean annual precipitation, the formation of new minerals is commonly related to depth from surface. C4 does not show any spectroscopic attribute except for a complete absence of class 6 and low range greyscale values (see Fig. 3.6 a). Clasts or large boulders protruding out of the trench wall are represented by intermediate value range class 5 on top and wide value range class 1 at the bottom (Fig. 3.5 c).

3.4.1.3 GPR Data Interpretation

Using the trench log and multispectral information enables radar facies to be distinguished. Figure 3.7 confirms the distinction of individual layers by comparison of reflected electromagnetic signal intensity. Reflections of visible and near-infrared light within certain zones that fit with trace increment and dimensions of the GPR system (30 x 2 cm) were sampled and correlated with the radar’s first arrival. As the vertical resolution is a quarter of the wavelength $\lambda$ (here: 30 – 40 cm), reflection amplitudes were averaged over 9 cm into depth per trace.

A good correlation between backscattered signals of both passive and active methods is obvious in some parts. A significant contrast in all GPR images is traced by the abrupt transition from fault gouge to palaeosol I (see Fig. 3.5 c and Fig. 3.7). Where
Figure 3.6: Statistical analysis of two-channel composition image. a) Seven distinct classes were estimated from the unsupervised classification. b) Histograms of representative classes per identified layer. Either the majority of a mapped layer is filled up by one class (e.g., fault gouge) or by a certain composition of 2 or 3 classes (e.g., Palaeosol II and C6).
3. Imaging a trench

![Figure 3.7: Varying reflectance of electromagnetic waves along the trench wall. Transitions between individual layers are depicted by drastic changing shapes of reflectance spectra. The error range is given by the standard deviation of each sample.](image)

reflections of visible and near-infrared light are intense on the surface of the fault gouge exposure, they rapidly decrease in signal strength on the palaeosol surface. The opposite reflectance behaviour is observed for radar reflections in the very shallow subsurface; the first lithological transition is characterised by the change of low to moderate reflection amplitudes in the fine-grained homogeneous fault gouge to higher reflection intensities from heterogeneous palaeosol I.

Moderate reflectance with intermediate variance designates the exposure of palaeosol I. A slightly decreasing trend is obvious within this section just before an abrupt rise in both visible and near-infrared light reflection values. This changeover is not obvious from GPR mean values. However, the value range given by the standard deviation of each sample has a wider reach than in the previous section (Fig. 3.7). Moreover, there is little distinction between individual colluvial deposits from GPR reflection amplitudes.

As previously stated, the HRDBSM shows an unrecognised feature in the middle of the trench exposure. A change is proven by a drastic drop in reflections from the GPR signal approximately 3 m from the fault plane. At the same position there is also a minor photomosaic and HRDBSM value decline. Thus, a conspicuous progression similar to a Gaussian bell shape curve in the middle of a dataset is obvious.

Layer C1 is not individually considered since the coupling of the antenna on heavily weathered cemented material with rugged surface relief was not sufficient. However, other transitions recognised in trench log and imaging spectroscopy can be traced in GPR images. This then leads to a 3D model of coseismic features within the hanging-wall (Fig. 3.8). Seven out of the ten (boulders are not included as an individual layer) previously mapped units plus the limestone fault plane to the West and the adjacent loose material to the East can be traced at depth using GPR.
Figure 3.8: Three-dimensional reconstruction of differing layers within the outcrop from GPR image interpretation. Partial reflection of radar waves on layer contacts leads to significant backscatter signals at depths down to approximately 3 m.
The 3D interpretation from GPR images visualises the continuation of distinct layers observed from multispectral analysis into depth. The limestone fault plane and fault gouge clearly differ in GPR images. Also, the cemented colluvium C1 is characterised by continuous and high amplitude reflections. Coarse grained components within other colluvial layers are represented as signal scattering hyperbola within a homogeneous matrix facies. However, a distinction between C4, C5 and, C6 could not be done with these data. The two palaeosols differ in the recorded intensity of the reflected electromagnetic waves. Where palaeosol I is characterised by high amplitude reflections, palaeosol II contains only minor reflection hyperbola caused by small clasts within the homogeneous matrix.

3.4.2 The Kaparelli fault-trench, Gulf of Corinth

The description of the Kaparelli fault trench follows the lithological designations of Kokkalas et al. [2007]. The hanging-wall and footwall of the Kaparelli fault are clearly separated by a 70–80° south dipping fault zone (Fig. 3.1 b). This zone is characterised by a chaotic assemblage of sheared deposits and material from surrounding or overlying units that has fallen into cracks and fissures. The footwall consists of multi-coloured pebbly-cobbly gravel deposits with a wide range of coarse-grained sub-angular to well-rounded clasts in a silty cemented matrix. The hanging-wall block comprises thick deposits of sandy silt (loess deposits) with many steeply dipping fissure fills, some cutting the entire trench wall and others only partly. The fissure width ranges from around 10 cm to over 80 cm and are filled with sub-angular to rounded gravel deposits in a silty matrix (Fig. 3.9 a).

The trench log, calibrated using the results from Kokkalas et al. [2007], correlates well with the results from imaging spectroscopy (Fig. 3.9 a). Coarse grained parts of the exposure to the northern end exhibit a widespread range of greyscale values in both the photomosaic and HRDBSM. Due to a grain size in the order of tens of centimetres and the resulting rough relief, shadows are generated in such a way that significantly influences the colour texture of the photomosaic and the backscattered signal. However, a distinct transition to a silty-sand unit, which prior to this study was described as the fault zone of the 1981 rupture event [Kokkalas et al., 2007], is very clear. Few and much smaller clasts in this unit (diameter is about 1 cm, < 15 %) and a homogeneous matrix have led to a uniform display in the false colour image. This composition of concurrent greyscale values in the photomosaic and HRDBSM occurs three times in constant offsets along the trench exposure. Pure silt underlies the silty sand. A fissure fill structure of pebbly gravel, dipping about 70° to the south, separates the two blocks of silty-sand and silt layers. Again, a rougher relief leads to a large range of backscattered signal values from

Figure 3.9: Results from the Kaparelli fault trench site from Kokkalas et al. [2007]. a) 2-channel composition from multispectral approach. Red and green are 100 % different whereas yellow colouring represents intermediate correspondence of both channels. The trench log (black lines) fits with the multispectral cluster of a certain composition. b) Three-dimensional reconstruction of the trench exposure. Recorded thickness of the colluvial wedge from 1981 is about 0.6 m.
3. Imaging a trench

Figure 3.9: See facing page.
both active and passive systems. However, sharp delimitations of juxtaposed lithological units based on their spectroscopic appearance are clear and discernible. A buried soil horizon and a colluvial wedge resulting from the 1981 surface rupturing event [Kokkalas et al., 2007] are visible and clearly textured by a certain composition of greyscale values.

In Figure 3.9 b, a three-dimensional reconstruction from GPR images of the trench wall shows that exposed structures do not only occur on the surface but are also traceable into the hanging-wall. Using layer differentiation from imaging spectroscopy helps to recognise certain radar facies even when there are only subtle distinctions. Major components of the trench wall are identified in individual GPR images. Their three-dimensional extension information is assembled by interpolating between multiple overlaying GPR images. Hence, information on continuation into depth as well as the varying thicknesses of individual layers is gathered. For instance, the colluvial wedge has only a minor variation in its thickness to 2 m penetration depth. The estimated average height for this unit is 0.6 m. This correlates to palaeoevent magnitudes of M6.5 [Reiss et al., 2003] which is comparable to previous ruptures [Kokkalas et al., 2007]. Adjacent units that differ by huge grain size contrasts, like sand and silt next to gravel units, are easy to recognise. Coarse components produce chaotic reflections, while fine grained units of homogenous material appear with even and quasi-parallel reflections. Thus, the very fine grained silty clay parts produce fewer reflections than those of pure sand. The unit of debris-element association contains poorly sorted coarse-sized gravels that are expressed by wavy reflection pattern that do not appear in the hanging-wall in the South.

3.5 Discussion

Trenching investigations have been one of the established methods in palaeoseismic research for the last decades. However, the outcome is highly dependent on the ability of the trench logger to define mappable units and the influence of sunlight since only visual appearance is used to make decisions on individual layer distinction. Furthermore, producing an accurate log and interpretation requires experience and excellent sketching skills. This process can be enhanced using the outcome of a numerically and a multispectral view of the palaeoseismic exposure, which allows quantitative information (reflectivity of electromagnetic waves at different spectra at certain materials) to be assigned to mapped units within the trench wall.

There are some significant disadvantages of passive data collection imaging techniques. These are mainly due to differing angles of illumination because the trench exposure is not a perfectly even surface at all scales; at larger scales surface undulations dramatically increase. Thus, the lightest parts in the photomosaic, visualised for the Sfaka roadcut as class 7 with an average value of 221, mainly represent a high matrix luminance and the top (bright) sides of boulders and clasts. Rectification and parallax effects yield an additional error in the order of a few centimetres. However, those effects can be used in structure from motion applications to speed up data collection and improve photomosaic quality as shown by Reitman et al. [2015]. High-resolution 3D images and the near infrared backscatter signal from TLS provide information on the physical properties of materials. Colour, matrix, surface roughness and orientation, and varying water content influence the TLS backscatter signal. A multispectral approach, using unsupervised clustering on both
spectra supports the results from the trench log and complements the findings. Thereby, a
distinct layer signature given by particular compositions of effecting classes allows adjacent
stratigraphic units to be differentiated. Some areas within the multispectral image lack
evidence for distinct spectroscopical characteristics. However, these areas can still be
defined when they are adjacent to areas with static characteristics; the boundary between
two areas is clearly defined as long as one area can be classified using the unsupervised
clustering. Therefore, a spectroscopically inconspicuous and completely heterogeneous
area surrounded by regions with static characteristics is still sufficiently confined. Within a
given error range due to manual gridding on the trench wall, georectification and blending
pixels of the photomosaic data, the results show many resemblances to the manually drawn
trench log.

The results of the imaging spectroscopy verified the lithology of the trench wall and
the resulting image from the unsupervised classification serves as a calibration factor for
GPR measurements. Due to the GPR’s resolution being about 0.1 m, the calibration
is necessary to recognise and interpret minor differences in sedimentological compositions.
This method allows more accurate calculations of mean geometric layer thicknesses to be
made, which are needed to correlate the amount of vertical offset caused by a specific
surface rupturing event. Information on the average height of a colluvial wedge can be
estimated from the in-depth data and then be used to estimate palaeomagnitudes [e.g.
Reiss et al., 2003]. The quality of the 3D GPR image and its interpretation depends
on well-structured data acquisition and processing, as well as on the experience of the
operator. The coupling of the antenna to the surface is decreased on bumpy surfaces,
which leads to lower quality data. Moreover, the reference grid on the surface poses
a source for stumbling. However, the grid is needed to fuse the geophysical data with
remotely collected data and to locate the GPR images in three-dimensional space. An
alternative to a grid made of string is colour spray to mark locations for orientation; but,
these would have a significant impact on the results from imaging spectroscopy. When
the survey is accurately planned and organised good results can be obtained which allow
a 3D interpretation of sedimentary features to between 2 and 3 m depth within the trench
wall.

The biggest disadvantage of the presented workflow is by far the effect of sediment
moisture content on reflectance, both in the multispectral analysis and GPR survey. For
the multispectral analysis there is not only darkening of the sediments, which leads to
an overall reduction of reflectance, but significant partial absorptions at wavelengths near
1.4 and 1.9 µm is also common [Lobell and Asner, 2002; Ragona et al., 2006]. Moreover,
water content in a given medium leads to distortion effects and high attenuations of elec-
 tromagnetic waves [Neal, 2004; Schrott and Sass, 2008]. However, for conditions when the
moisture content is similar throughout the trench wall, water absorptions should not affect
the correlations because reflectance along the wall should be affected uniformly. Ragona
et al. [2006] have shown that identifying stratigraphy with samples that maintain high
amounts of their original moisture content is possible; however, the authors’ suggestion
to consider necessary approaches to minimise changing reflectance is endorsed. Indeed,
the herein presented workflow was successfully tested on normal faults vertically displace-
ing carbonatic bedrock from mostly post-glacial colluvial sediments. Therefore, it can
be suggested that this technique can only be applied in semi-arid to arid regions such as
the Mediterranean or the western USA, where the sediment moisture content is relatively
low, at least for a couple of months per year. However, the multispectral analysis and classification does not incorporate layer orientations and can therefore be applied not only on bedrock normal faults but on any kind of layer discontinuity separating two individual electromagnetically responding facies. Furthermore, the presented technique is robust in identifying distinct sediments (see Figs. 3.5 and 3.9). Exposures in humid climates most likely maintain much more water. However, Holocene surface ruptures preserved in the shallow subsurface are capable to clearly show progressive displacements which can also be detected by differing electromagnetic responses [Grützner et al., 2016a].

Other potential error sources using this technique are dependent on the characteristics of the individual trenching sites and the equipment used. Some sites are hard to access because of steepness, height and/or width of the excavation. Extremely steep or narrow trenches make the installation of the scanning equipment difficult. Exposure heights exceeding usual body heights generate problems for the GPR survey; these can be overcome using ropes and wooden tools to ensure good coupling. Scaffolding usually consists of metal which may lead to interferences in the GPR image. If the trench wall is not properly prepared in terms of cleaning, or the embedded sediments produce a rough surface because of coarser grain sizes, spectral amplitudes will change because of varying illumination and incident angles. Therefore, the spectroscopic interpretation must take these accompanying effects into account. Moreover, extremely complex sedimentological architectures may cause complicated multi-pathing effects on the radar waves. The presented workflow has basic requirements concerning computing capacities; the collected high-resolution data from conventional photo cameras, TLS scanning and GPR measurements engage substantial disk space and random access memory.

One major benefit from this workflow is the storage and future use of the raw data. The majority of palaeoseismic trenches are designed to be closed after field investigations are completed. This means that not only is there no future access to these exposures, but the sedimentological environment of the excavated site is also destroyed. If a trench is left open after field investigations, the trench walls will get degraded and altered by weathering effects. TLS and GPR measurements provide and store information on the visual appearance of the trench and the reflection properties of different electromagnetic wavebands. The reflectance spectrum at each pixel of an image provides unbiased compositional information. This saved data can always be used for future (re-)analyses. Another benefit is the ability to record trench data in hazardous exposures without extensive, time consuming and costly safety precautions. Also, as trenches are often only open for limited durations, the logger might not have enough time to accurately sketch and measure components, or he may rush to finish. In these cases, capturing and recording the outcrop in a multidimensional manner (x,y,z coordinates of each data point plus reflectance values of visual and near-infrared light and pseudo-3D information within the hanging-wall) enables efficient productivity and forms a complementary approach.

### 3.6 Conclusion

Identifying and mapping individual lithological units along a palaeoseismological exposure in accordance with colour and matrix specifications as well as sedimentary structures and soil formations are core competencies of palaeoseismic trenching studies. However, the
accuracy and quality of the log and interpretation is highly dependent on the experience of the trench logger, and is thus subjectively influenced. Hence, minor differences in lithological description from expert to expert are expected, especially if one logger has access to no more than a photomosaic. In order to prove whether conventional trench logging methods used to map coseismic features in a palaeoseismic trench wall can be objectively enhanced, we created an accurate digital version of the exposure and its physical properties. This was done by combining routine logging with vertical GPR measurements and imaging spectroscopic approaches from normalised photomosaics and high resolution TLS backscatter models. Both the studied palaeoseismic exposures, on Crete and mainland Greece, exhibit sedimentary structures whose constituent parts and shape are essential information for a palaeoseismic reconstruction.

After the conventional trench logging was completed, TLS scans were undertaken at close range. The near-infrared backscattered signal was combined with a luminance bearing photomosaic of the same trench wall. Statistical and classification techniques reproduce an objective digital copy of a palaeoseismic trench log. In order to define distinct units, four options to characterise and differentiate individual layers by imaging spectroscopy can be registered:

- Significant dominance of a certain class within a distinct layer
- Certain composition with spatial clustering
- Certain composition with certain arrangements
- Distinct borders between individual layers although one or both are not determined by applied statistics

Subtle or invisible features are enhanced and become part of a quantitative analysis, and comparisons of units using their reflectance on certain wavelengths [see also Ragona et al., 2006] can be carried out. The results show that based on distinct layers in the trench log, in combination with the outcome of imaging spectroscopy, a 3D-interpretation of GPR-data carried out vertically on the trench wall is possible. Hence, the spatial extent of palaeoseismic features can be traced within the trench wall. The resulting 3D model from the GPR provides information on representative layer thicknesses, displacements, and geometries. This is highly beneficial since it minimises misinterpretation due to cutting effects.

To extract such fault specific information is not only crucial for identification and mapping active faults but also depicts complementary input for seismic hazard assessment by extracting more accurate magnitudes of palaeoearthquakes [Papanikolaou et al., 2015]. The use of TLS became a major tool to obtain such data. So far, this modern technology was used for fault mapping at regional to very scale coverage with up to millimetre resolution [e.g. Arrowsmith and Zielke, 2009; Begg and Mouslopoulou, 2010; Wilkinson et al., 2010; Bubeck et al., 2015; Wilkinson et al., 2015]. Further, the visualisation of bare-earth topography in regional scale [Cunningham et al., 2006] as well as detection of roughness changes along fault scarps [Wiatr et al., 2015] are scopes of application. Here another approach of the use of TLS in palaeoseismology is presented. Recording and measuring the backscattered signal in the near-infrared band enables the visualisation of
usually non-visible electromagnetic waves. The spectral response represents material specific properties and gives evidence for differing lithology along the exposure. For seismic hazard assessments, accurate and justified decisions on the interpretation of such data are needed. To further assist, high resolution GPR profiling visualises the associated sedimentary architecture within the hanging-wall and quantifies and qualifies event horizons to estimate palaeomagnitudes and slip rates on active normal faults [Reiss et al., 2003].

The presented workflow does not form an alternative to conventional trench logging since this approach only records complementary data. Information on detailed grain-size distribution along the exposure or the orientation of certain components is not addressed by the workflow. Even photomosaic methods cannot offer required pixel resolution. However, if logistics are difficult and/or trench wall are hazardous, a TLS scan and photographs can be applied from outside of the exposure and be used to quickly provide high resolution data. This forms an alternative data collection method when the opening time is short or when operators cannot stay safely in the trench. The provided data visualises features that are usually not visible, allows decisions on interpreting the seismic history of the fault to be justified, and the spectrum reflectance data provides unbiased measurements that can be (re-)processed any time after the trench has been backfilled.

Reconstructing the palaeoseismological history of both trench exposures is not an integral part of this chapter. However, the objective of improving individual event horizon recognition using multispectral viewing and 3D visualisation of GPR images was successfully undertaken. This method can therefore contribute to the accuracy of seismic hazard assessment.
Science, my lad, is made up of mistakes, but they are mistakes which it is useful to make, because they lead little by little to the truth

J. Verne (1828-1905)
Tidal notches are a generally accepted sea-level marker and maintain particular interest for palaeoseismic studies since coastal seismic activity potentially displaces them from their genetic position. The result of subsequent seismic events is a notch sequence reflecting the cumulative coastal uplift. In order to evaluate preserved notch sequences, an innovative and interdisciplinary workflow (Fig. 4.1) is presented that accurately highlights evidence for palaeo-sea-level markers. The workflow uses data from terrestrial laser scanning and iteratively combines high-resolution curvature analysis, high performance edge detection, and feature extraction. Based on the assumptions that remnants, such as the roof of tidal notches, form convex patterns, edge detection is performed on principal curvature images. In addition, a standard algorithm is compared to edge detection results from a custom Fuzzy logic approach. The results pass through a Hough transform in order to extract continuous line features of an almost horizontal orientation. The workflow was initially developed on a single, distinct, and sheltered exposure in southern Crete and afterwards successfully tested on laser scans of different coastal cliffs from the Perachora Peninsula. This approach allows a detailed examination of otherwise inaccessible locations and the evaluation of lateral and 3D geometries, thus evidence for previously unrecognised sea-level markers can be identified even when poorly developed. High resolution laser scans of entire cliff exposures allow local variations to be quantified. Edge detection aims to reduce information on the surface curvature and Hough transform limits the results towards orientation and continuity. Thus, the presented objective methodology enhances the recognition of tidal notches and supports palaeoseismic studies by contributing spatial information and accurate measurements of horizontal movements, beyond that recognized during traditional surveys. This is especially useful for the identification of palaeo-shorelines in extensional tectonic environments where coseismic footwall uplift (only 1/2 to 1/4 of net slip per event) is unlikely to raise an entire notch above the tidal range.
4. Tidal notch detection

![Diagram of tidal notches, curvature analysis, and edge detection]

Figure 4.1: A simplified scheme illustrating the high-performance workflow for tidal notch detection.

4.1 Tidal notches in palaeoseismology

In microtidal seas, such as the Mediterranean, tidal notches can be used to derive and quantify relative coastal movements during the Holocene [Pirazzoli, 1991]. To develop these prominent strandlines, ranging from a few centimetres to several metres deep, the sustained action of physical, chemical, and biological erosion within the tidal range is necessary. Therefore, exposure to wave action, lithologic resistance to quarrying, and the strength of the rock able to support the weight of the overburden are key parameters effecting the shape of resultant notches [Trenhaile, 2015]. In tectonically active regions, these distinct ecological and morphological features define the modern shoreline, and when equivalent older features are different from the present-day sea-level coseismic activity can be inferred (Fig. 4.2) [e.g., Boulton and Stewart, 2015]. However, a direct correlation of individual sea-level markers to palaeoearthquake parameters is an outstanding challenge especially in extensional tectonic settings. For example, the shoreline of western Crete was uplifted by up to 9 m during the compressional M 8.5 Hellenic earthquake in 365 A.D., forming a classic example for a lifted prominent strandline as a consequence of rapid emergence [Shaw et al., 2008]. This distinct palaeoshoreline is well-preserved and has not been affected by wave attack or midlittoral erosion. By contrast, shorelines that experienced rapid emergence due to extensional tectonic movements, such as those from Perachora Peninsula in the Gulf of Corinth, are not likely to preserve fully developed tidal notches. In these settings, the amount of coseismic displacement is usually up to an order of magnitude lower than in megathrust events, and moreover the uplift component is estimated to be only 1/4 to 1/2 of the net slip per earthquake [e.g., Armijo et al., 1996; McNeill et al., 2005; Papanikolaou et al., 2010] and thus not likely to exceed the tidal range of ~ 0.4 m in the Mediterranean Sea [Evelpidou et al., 2012a]. Therefore, it is suggested that apparent notches reflect the cumulative effect of multiple seismic events and individual notch levels cannot usually be attributed to specific earthquakes in regions of tectonic extension [e.g., Stewart and Vita-Finzi, 1996; Cooper et al., 2007; Boulton and Stewart, 2015].
4. Tidal notch detection

Figure 4.2: Collage of raised shorelines on Crete and Central Greece and associated notch profiles extracted from TLS data. The tidal notch at Agios Pavlos (a) was raised by the 365 A.D. earthquake and forms the reference for notch detection [Shaw et al., 2008]. Exposures at the coast of Perachora Peninsula (Gulf of Corinth) are known from literature - [Kershaw and Guo, 2001] and [Pirazzoli et al., 1994a] - and pose testing targets in this study: b) Mylokoopy Bay; c) Heraion Harbour, and d) Heraion Lighthouse.
4. Tidal notch detection

The identification of a palaeoshoreline is, among bioerosional remnants or consolidated beach deposits, based on the recognition of distinct erosional marks of former midlittoral zones [Pirazzoli et al., 1994a]. Typically, the notch position is mapped on a 1:5000-scale map [Cooper et al., 2007] and measurements are made to create morphometric profiles. Profiles are usually collected by tape measure [e.g. Kershaw and Guo, 2001] and include the average vertical extent of a notch and the maximum indentation [e.g. Antonioli et al., 2015]. Vertical sheltered coasts are preferred for precise notch measurements [Pirazzoli, 1986], yet often these cliffs are inaccessible, and for that reason mid-range profiling using a handheld laser distance meter allowing evaluation of inaccessible and dangerous cliffs has also been employed [Kázmér and Taboroši, 2012]. To address morphometric variations, a structure-from-motion (SFM) approach is also presented by Bini et al. [2014], which produces high resolution 3D models from a surface using a series of overlapping photographs.

The problem of lateral profile heterogeneity is extensively discussed by Kershaw and Guo [2001], demonstrating that active fault segments crossing cliffs, local variations of different wave and surf regime, and/or bedrock heterogeneity result in different notch profiles even in nearby sites [see also Evelpidou et al., 2012a]. Furthermore, collecting multiple profiles manually is time consuming and contains potential error sources. For instance, the correlation of different extracted levels from morphometric profiles is challenging and requires a constant reference datum over the time period of profile collection. We suggest that terrestrial laser scanning (TLS) provides all requirements for palaeoseismological studies on emerging coasts. The data are of high precision and resolution, and enables the analysis of the surface curvature of a whole cliff in a reasonable amount of time.

This paper aims to present an interdisciplinary study of computer vision and palaeoseismology. High resolution data from TLS is investigated utilising multiscale image analysis and semi-automatic edge detection. Conventional gradient analysis is compared to modern modelling from Fuzzy logic methodology. Afterwards, feature extraction by Hough transformation gives spatial evidence for the existence of tidal notches within an entire sequence of palaeo-strandlines on a cliff.

In their comprehensive analysis of tidal notches in the Mediterranean, Antonioli et al. [2015] concluded that notch formation processes have not changed during the last 125 kyrs. Similar widths of both last interglacial and modern notches suggest equivalent tidal ranges as zones of notch formation. Hence, the retreat zone of a tidal notch representing mean sea-level can be inferred by knowing the local tidal amplitude and the position of either roof or floor. Particularly in the Mediterranean, the use of tidal notches as palaeo-sea-level markers to determine rates of tectonic activity is widespread, since potential errors are limited by low tidal ranges and the lack of strong waves [Pirazzoli and Evelpidou, 2013]. Therefore, the coastline at Perachora Peninsula in the eastern Gulf of Corinth provides suitable conditions to apply an innovative method improving tidal notch identification and comparison on local and regional scales. In order to verify and calibrate the method, which focusses on changing curvature at the roof or bottom of a notch, a distinct tidal notch in southwestern Crete ~ 1 m above recent sea-level uplifted by the 365 A.D. earthquake [Shaw et al., 2008] is investigated as reference model.
4.2 Investigated sites

4.2.1 Agios Pavlos, SW Crete

The island of Crete is directly adjacent to the Hellenic subduction zone between Europe and Africa (Fig. 4.3) and comprises a complex geological and tectonic structure that results from successive thrusting of alpine geotectonic units and the activity of major detachment faults. Crustal extension orientated both arc-parallel and arc-perpendicular has led to the development of Quaternary carbonate bedrock fault scarps throughout the island [Caputo et al., 2010]. These normal faults mainly juxtapose Mesozoic carbonates of the Pindos unit in their footwall against hanging-wall flysch and/or post-alpine sediments. Vertical tectonic movements along the western part of the island are associated with both fault populations, causing earthquakes along the nearby Hellenic trench and on normal faults onshore. As a result, clearly visible emerged shorelines are developed on the limestone cliffs. The 365 A.D. earthquake rapidly uplifted the well indented strandline by ~1 m at Agios Pavlos, located approximately 70 km eastwards from the activated structure and evidences the recent regional uplift phase [Stiros, 2010]. Crete has experienced ~2.5 km of uplift since the Early Tortonian (Miocene) in several different phases [Meulenkamp et al., 1994]. The most recent phase of uplift, as evidenced by uplifted Messinian deposits [Krijgsman, 1996], began at around 6 Ma and continues to the present day. The study location is located inside a 200 m wide bay and is protected from rough seas in accordance with official nautical cartographies and data from oceanographic buoys [http://utmea.enea.it/energialmare/].

4.2.2 Perachora Peninsula, eastern Gulf of Corinth

North-South directed extension with rates of 10–15 mm/yr makes the Gulf of Corinth one of the most rapidly extending areas on Earth. Along the southern shore of the graben are active north-dipping normal faults uplifting coastal regions in the footwall. Rates of fault motion lie in the range of 1–10 mm/yr and are evidenced by Quaternary and Holocene palaeoshorelines [Armijo et al., 1996; Morewood and Roberts, 1999; Cowie and Roberts, 2001; McNeill and Collier, 2004; Leeder et al., 2003; Cooper et al., 2007; Roberts et al., 2009]. Leeder et al. [2005] estimate slip rates of ~2.5 mm/yr for normal faulting structures in the Alkyonides Gulf and the Perachora Peninsula over a period of 0.6 Myrs (Fig. 4.3). However, the authors also postulate that onshore faults (Schinos and Pisia) are more active than parallel offshore structures.

The coastline of the Perachora Peninsula is predominantly comprised of Mesozoic and Pleistocene carbonates. In some parts of the southwestern part of the peninsula, a thin composite volcanosedimentary series of basic rocks occurs. Occasionally, marine deposits of Tyrrenhian age comprising conglomerates crop out along northern coastlines [Bornovas et al., 1984b].

The Heraion archaeological site is located at the northwestern tip of the Perachora Peninsula (Fig. 4.3b). The tidal notches at this site are described by several authors. Pirazzoli et al. [1994a] identified four raised notches at the lighthouse between +1.1 and +3.2 m and dated them to 4.4–4.3 kyrs BP (+3.2 m), 2.4–2.2 kyrs BP (+2.6 m), and 0.4–0.2 kyrs BP (+1.1 m) (see Fig. 4.2). Kershaw and Guo [2001] tried to correlate these
4. Tidal notch detection

Figure 4.3: Overview map of studied sites. a) Map of Greece showing simplified large-scale tectonic structures (CG, Corinthian Gulf; CF, Cephalonia Fault; NAF, North Anatolian Fault; NAT, North Aegean Trough; black lines with bars show active thrusts; black lines with marks show active faults) [after Papanikolaou and Royden, 2007; Shaw et al., 2008]. Red boxes highlight study areas. b) DEM (from 10m contour lines) of the Perachora Peninsula. Red lines with marks indicate normal faults that have been activated during the 1981 earthquake sequence [Bornovas et al., 1984b]. LV, Lake Vouliagmeni. c) DEM (SRTM-1) of the southwestern coast of Crete. The morphology indicates tectonic structures (black line with marks) that potentially down-throw coastal areas [Bonneau, 1985].
notches to exposures at the harbour of Heraion only a few hundreds of metres to the east (+0.75 and +2.05 m). The authors conclude that differential uplift on cross-cutting faults causes dislocations of former strandlines and prevents a correlation between the two sites.

Another site mentioned by both studies is located along the northern shore of the peninsula. The Mylokopy beach actually consists of two small bays, separated by a tombolo. At the tip of the tombolo a massive limestone block contains up to five notch generations, which vary in height from the surrounding cliffs because of fault activity. In addition, three different notch morphometric profiles (identified notches at +0.4, +1.2, +2.0, and +2.6 m) can be extracted due to varying exposure to the sea and abrasional components [Kershaw and Guo, 2001].

4.3 Methodology for visualising tidal notch morphologies

The methods presented include data acquisition from TLS and processing for semi-automated edge detection based on the surface curvature of a cliff. One scan from the distinct shoreline at Agios Pavlos operates as a reference for a unique tidal notch at this particular cliff, since the 365 A.D. thrust event raised the strandline > 1 m from the erosional zone. Thus, we assume this exposure is not affected by ongoing erosion. Consequently, the method is developed from this exposure and then tested on sites from the Perachora Peninsula.

4.3.1 Theoretical assumptions

The term tidal notch refers to a horizontal erosion feature formed at sea-level due to coeval action [Antonioli et al., 2015] of chemical, physical, and biological factors [Pirazzoli, 1986]. However, the predominant agent is commonly assumed to be bioerosion [Evelpidou et al., 2012a], which is restricted to carbonate rocks. Well-defined vegetational belts are the result of different grazing or boring organisms each living in individual horizontal galleries. Therefore, Pirazzoli [1986] suggested a vertical zonation (Fig. 4.4a) for notches, which also indicates maximum erosional potential at mean sea-level (Fig. 4.4b). Moreover, the classical symmetrical notch profile [e.g. Laborel et al., 1999; Trenhaile, 2015] is formed of three main sections (Fig. 4.4): I) A floor or base which extends to the limit of permanent immersion at tidal low stand; II) a retreat zone of maximum concavity exhibiting the inflection point near mean sea-level, and III) a roof near high tide level.

In an area of extensional tectonics, such as the Gulf of Corinth, the ratio of footwall uplift to hanging-wall subsidence is estimated to 1/4 to 1/2 where the total net slip is not likely to exceed ~2 m, since normal faulting structures usually do not produce earthquakes > M 7.0 [e.g. Jackson et al., 1982; Stewart and Vita-Finzi, 1996; Papanikolaou et al., 2010]. Offshore, but close to the coast, normal faulting seismic activity causes rapid emergence of coastal cliffs; however, coseismic uplift exceeding the tidal range of ~0.4 m is unlikely since it would require minimum mean displacements of 1.6±0.4 m (based on Wells and Coppersmith [1994]; for M 6.5–7.0 empirical maximum displacements range from 0.8 to 2.1 m) which are unrealistic values of surface faulting for the vast majority of normal faulting
Figure 4.4: Theoretic assumptions. a) Zonation of a simplified tidal notch (R, roof; F, floor; IP, inflection point) following suggestions of Pirazzoli [1986]. b) Evenly distributed erosional potential pointing at mean sea-level causes a symmetrical shape of a tidal notch (I). When the erosional zone gets offset by an earthquake (II-IV) the level-based erosional potential attacks the prior to this created cliff morphology (III). The resulting shape comprising two notch generations (1. and 2.) exhibits patterns of convex or concave curvature (c). d) Visualises the estimate of the normal vector ($\hat{N}$) at any point (P) along a normal section from principal curvatures $k_1$ and $k_2$. 

4. Tidal notch detection
earthquakes. Thus, the former and new erosional zone along the cliff would overlap, overprinting the earlier notch (Fig. 4.4b). Pirazzoli [1986] labels features of this origin as 'ripple notches'. However, depending on the time and vertical displacement, the resulting shape is tantamount to a widened single notch; due to the tidal range variation. Only at close range minor variations will be detectable on the surface curvature and normal to the orientation of the roof (Fig. 4.4c).

4.3.2 TLS

Terrestrial laser scanning (TLS) is a commonly used remote sensing technique with a high spatial and temporal resolution and is highly effective for reconstructing morphology [Wilkinson et al., 2015], interpreting trenches and outcrops [Schneiderwind et al., 2016], monitoring movements [Rosser et al., 2013], extracting slip vectors [Jones et al., 2009], and recording smoothness along fault planes [Wiatr et al., 2015].

The fundamental principle underlying TLS is rapid measurement of one-dimensional distances using a model-specific wavelength within the electromagnetic spectrum. A coherent laser beam with little divergence propagates dominantly in a well-defined direction and is reflected off surfaces, forming a non-contact and non-penetrative active and stationary recording system. Most common are systems that make use of the time-of-flight principle, where the instrument measures the time delay between emission, reflection and receiving the laser pulse. Phase-based TLS bypass the requirement of a high-precision clock by modulating the power of the laser beam and measuring the phase difference between the emitted and received waveforms [Smith, 2015]. The result is an irregular but dense point cloud (x,y,z coordinates) representing a highly detailed digital 3D surface model. In both systems, the data quality is controlled by the range between sensor and target, surface properties (e.g. moisture, roughness), and also the angle of incidence.

In this study we used a time-of-flight mode operating Optech ILRIS 3D system for scans collected on Crete and a Faro Focus 3D system (phase-based mode) during the survey in central Greece due to logistical constraints. All scans were undertaken during calm sea conditions and from close-range to mid-range (max. 100 m). In order to correlate the data from multiple sites at the Perachora Peninsula, hourly tide gauge data from the Posidonia station (Hellenic Navy Hydrographic Service) was applied to the individual point clouds referenced to mean sea-level (Fig. 4.5).

Once the point clouds are corrected for their individual spatial information, principal curvature analysis is performed. In general, curvature is the second derivative of a function f(x) and describes the amount by which a geometric object differs from being flat. Depending on the sign, the object is either convex or concave at any point P, and the surface normal \( \vec{N} \) is oriented perpendicular to the surface towards maximum curvature. The magnitude k of difference from a flat object is quantitatively described by:

\[
k = \frac{f''(x)}{[1 + (f'(x))^2]^{3/2}}
\] (4.1)

The mean curvature at a point on a third dimension uses both the maximum and minimum normal curvatures. These principal curvatures are orientated mutually perpendicular with \( k_1 > k_2 \) (Fig. 4.4d). However, since tidal notches are a horizontal sea-level
4. Tidal notch detection

**Figure 4.5:** Data acquisition and processing: a) Close- to mid-range laser scanning; b) Tide gauge data provided by the Hellenic Navy Hydrographic Service ($x$ = mean sea-level, $\sigma$ = standard deviation, $r$ = tidal range) from the moment of scanning (red dots). c) High resolution point cloud data adjusted to mean sea-level using the tide gauge data as a reference datum. d) Segments are then prepared for surface curvature analysis. Extraction of two-dimensional information about the surface curvature reduces error sources from interpreting 3D surfaces.
marker, only the vertical principal curvature is respected for the analysis. Moreover, the minimum curvature \(k_2\) highlights exclusively convex patterns corresponding to features such as the roof of a tidal notch. This automatically excludes sources of misinterpretation (e.g. joints or cracks) and focuses on horizontal differences (Fig. 4.5d).

To calculate the surface curvature, TLS data provides surface information with \(x, y, z\) coordinates, where the \(z\)-coordinate describes the lateral indentation value. To sharpen the principal curvature information, standard averaging and 2D median filtering are applied.

### 4.3.3 Edge detection

The curvature defines a parameter essential for curve sketching. However, this value does not have a primary link to neighbourhood relationships. Indeed, the curvature at any point is calculated from the adjacent points but it does not quantify geometric alignments, such as straight edges, and the curvature of neighbouring pixels is not compared. Therefore, methods of edge detection are applied which aim to identify points where abrupt changes and discontinuities in the surface curvature occur. Furthermore, the process reduces the curvature plot to its significant details that appear as convex objects.

#### 4.3.3.1 Canny Method

Edge detection is an integral part of many computer vision systems and multiscale image analysis. The method results in a dramatic reduction of processed data, while preserving structural information about object boundaries [Canny, 1986]. In general, an image contains edges where the gradients along the \(x\)- or \(y\)-axis show rapid changes in image intensity. For instance, the transition from black to white (which equals the values of 0 and 255 in an 8-bit array) within just two pixel cells depicts a sharp edge with the highest possible gradient. Ideally, the result is a binary image that only contains information about edges within the initial intensity image. To decide whether an edge is located at a certain part of the image, one of the following criteria has to be fulfilled:

- The first derivative of the intensity is larger in magnitude than a given threshold; or
- The second derivative of the intensity has a zero-crossing (i.e. where the intensity of the image changes rapidly or the first derivative changes sign).

The built-in Matlab\textsuperscript{TM} edge function provides several estimators that implement these rules. Furthermore, sensitivity for horizontal over vertical edges can be applied. The Canny edge detector has become standard in edge detection by defining two thresholds for strong and weak edges, respectively. Technically, the algorithm applies a Gaussian noise reduction and a non-maximum suppression to eliminate multiple responses. Edges classified as weak only persist in the resulting binary image when these are connected to strong edges. Therefore, the three criteria of edge detection (good detection, good localization, and low spurious response) are addressed [Canny, 1986; Bao et al., 2005].
4. Tidal notch detection

4.3.3.2 Fuzzy Logic

Zadeh [1965] described a fuzzy set as a class of objects without a precisely defined criterion of membership. Within a fuzzy set each object is assigned to a grade of membership ranging between zero and one. Hence, approaches for decision-making [Bellman and Zadeh, 1970] and cluster analysis [Bezdek and Harris, 1978] were developed. Translated to edge detection from surface curvature the Fuzzy logic approach allows the use of membership functions to define the degree at which a pixel belongs to a convex edge or a different region. This is also the essential statement defining the membership function. Therefrom, other than from the Canny edge detector, the result is an intensity image and not a binary type. Consequently, edge detection and recognition still belongs to the user and is not the result of any blackbox approach securing transparency in the process.

Edge detection using Fuzzy logic comprises three steps. Firstly, directional gradients ($G_x$, $G_y$) and gradient magnitudes ($Mag_x$, $Mag_y$) serve as input information for a fuzzy set and have to be obtained from the curvature plot using the Prewitt gradient operator (Fig. 4.6). The Prewitt operator is a standard edge detection algorithm that accurately highlights vertical or horizontal alignments [Zhang et al., 2013] (Fig. 4.6b). Secondly, a fuzzy inference system (FIS) specifies a zero-mean Gaussian membership function for each input where the range of directional magnitudes depicts the limiting range values (Fig. 4.6c). If the gradient value is zero the pixel belongs to the zero membership function of grade 1. The grade along function quantifies the degree of membership of a certain element to the fuzzy set. In order to adjust the sensitivity of edge detection, multiples of standard deviation ($s_x$, $s_y$) of both zero membership inputs control the edge detector performance. Because of the high resolution of TLS data and dense point cloud, those values should be $\geq 1$ to decrease sensitivity for areas of minor interest (e.g. small cracks or joints). Furthermore, defining $s_x \gg 1$ encompasses the majority of plan curvature within the zero-membership function and thus excludes those from analysis. Therefore, a triangular membership function is specified for the output intensity image. Start, peak, and end of the triangles influence the intensity of the detected edges and can be adjusted as required to improve edge detection performance.

The third step of edge detection from Fuzzy logic includes rule specification and evaluation of the FIS. For classification of the intensity map, two rules are necessary which access three simple principles of set theory (If-then, AND, OR):

- If $Mag_x$ is zero and $Mag_y$ is zero then intensity is white
- If $Mag_x$ is not zero or $Mag_y$ is not zero then intensity is black

By this formulation a pixel of gradient different from zero depicts black and belongs to an edge (Fig. 4.7). Furthermore, the gradient is defined to be zero by Gaussian membership functions and forms the input for the applied FIS.

4.3.4 Hough Transform

The Hough transform is a popular tool for feature detection due to its robustness to noise [Fernandes and Oliveira, 2008]. The technique aims to find imperfect instances of objects
Figure 4.6: Fuzzy set edge detection. Edge detection is performed on principal curvature images (a). Two-dimensional gradients (b) are individually addressed in defined membership functions (c). The intensity map (d) shows subsets of different memberships. White pixels belong to a uniform region; only very dark pixels represent detected edges (Fig. 4.7c).
Figure 4.7: Comparison of applied analyses. a) Principal curvature depicting a high resolution image of the cliff morphology. b) Edge detection after Canny. It is successful in notch detection but also highlights small edges of minor interest. c) Edge detection from Fuzzy logic highlighting rapidly changing gradients in a horizontal manner.
representing line features by a voting procedure. For this procedure image objects are compared to the parametric term of a straight line. For some technical reasons, it is proposed to use its Hesse normal form since vertical lines would give rise to unbounded values of the slope [Duda and Hart, 1972]:

\[ \rho = x \times \cos \theta + y \times \sin \theta \]  \hspace{1cm} (4.2)

where the variable \( \rho \) is the distance from the origin \((0,0)\) to the line along a vector perpendicular to the line, and \( \theta \) is the angle between the \(x\)-axis and this vector with a range of \(-90^\circ < \theta < 90^\circ\). Thus, the gradient of a line feature is the tangent of \(90 - \theta\). The result of the Hough transformation is a parameter space matrix comprising \( \rho \) and \( \theta \) vectors for each pixel \((x, y)\), where the algorithm determines evidence of a straight line with respect to neighbouring pixels. Furthermore, it depicts a voting map \([0 \ 1]\) representing the discretised parameter space of detected objects [Fernandes and Oliveira, 2008]. Local maxima (peaks) in this map represent parameters \((\rho, \theta)\) of the most likely lines that can be extracted.

Since the Matlab\textsuperscript{TM} Hough function requires a binary image input, the intensity map from Fuzzy Logic edge detection is converted using a global image threshold [Otsu, 1979]. Besides that, line segment extraction from the Hough transform follows the same workflow for both data sets from edge detection (Canny Method and Fuzzy Logic) (Fig. 4.8). After the Hough transform is computed, peak values in the voting map are identified, where the user specifies the number of peaks to identify and thus, controls the influence of minor objects.

4.4 Results

4.4.1 Developing methods, Agios Pavlos, SW Crete

The workflow was developed utilising curvature analysis, edge detection using a Canny algorithm, a Fuzzy logic approach, and Hough line extraction on laser scan data from Agios Pavlos. The principal curvature analysis clearly highlights convex patterns (Fig. 4.7a) as expected from theoretical assumptions (Fig. 4.4). The prominent strandline is obviously defined by an evenly convex roof and floor. However, not only horizontal asperities resulting from erosion at sea-level are registered. In order to reduce the image information and to focus on almost horizontal and continuous features, two individual edge detection approaches were applied. The conventional Canny edge detector predicts sharp changes in surface curvature suitable for the roof and the floor of the notch. Furthermore, minor morphological irregularities are ignored and not interpreted as a discontinuity. However, the algorithm does not sufficiently exclude information from plan curvature and consequently omits edges from features of minor interest, such as joints, cracks or weathering aspects (Fig. 4.7b). The Canny edge detector returns a bivalent set of uniform areas and edges and thus, does not differ for gradual irregularities within the subset "edge". Consequently, only the predominant horizontal orientation of edges detected at the extents of the notch is evidence for its existence. The membership functions of the Fuzzy logic approach allow outputs of quasi-probabilistic edge occurrence (Fig. 4.7c).
Figure 4.8: Hough Transform from detected edges. Dashed areas indicate potential line features with absolute $\theta > 80^\circ$. Due to its sensitivity, extracted line segments from Canny edge detector (a) are more spread and randomly orientated than from Fuzzy Logic edge detection (b). Peaks in the normalised voting map (squares) represent parameters for most likely lines. Zoom indicates a peak cluster of almost horizontal oriented line features corresponding to elevations of the notch’s roof and floor, respectively.
This means detected edges, which are almost the same as from the Canny detector, are ranked towards the grade of conformance with formulated rules. Furthermore, focus is complied with horizontal features reducing image information once more towards the recognition of sea-level marker.

The Hough transform returns a matrix of a discretised parameter space displayed as a graph of line feature distance from the origin ($\rho$) against line feature deviation from vertical ($\theta$). Fig. 4.8 contrasts the resulting matrices from Canny edges with edges determined from Fuzzy logic. It is obvious that peaks and hot spots representing accumulations of $\rho$, $\theta$-pairs are wider spread when Canny edges determine the input for Hough transform. Especially from $\theta$-values distributed edge orientations are confirmed (see also Fig. 4.7b). However, for almost horizontal line features the corresponding absolute $\theta$-value should be $> 80^\circ$, since it represents the normal vector orientation. When edges determined from the Fuzzy logic approach are input for the Hough transform the resulting peaks are clustered at highest $\theta$-values. Furthermore, hot spots are clearly separated from each other and enable correlation to corresponding heights in the laser scan (Fig. 4.8b). Peaks located at minimum or maximum $\rho$-values correspond to the upper or lower image extent. The laser scan at Agios Pavlos shows some minor wave action resulting in a lack of data in the lower part of the cliff section and causing detected edges and determined line features in this region (Fig. 4.9).

When comparing the results of line feature determination from different inputs, it

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**Figure 4.9:** Feature extraction from scan of the cliff at Agios Pavlos. a) Overall result. b) Extracted line objects from Canny edges. c) Objects from Fuzzy logic edge detection representing the sea-level marker more concentrated along the notch extent line.
is conspicuous how spread peaks in the parameter space influence the focus on distinct morphological features. Line structures extracted from Canny edges do not represent the roof and floor of the notch exclusively. Lines following edges from generic irregularities, such as those from weathering in the lower parts, are also extracted. Indeed, features with $\theta$-values $<80^\circ$ can be suppressed in the plot (see Fig. 4.9b) but this still does not provide a threshold for distinct features. Due to the membership functions of the Fuzzy logic approach gradual distinction of edge detection enables adjustment of such thresholds. As a result, only the notch at $\sim 1.2$ m is highlighted (Fig. 4.9c). Therefore, it seems the identification of tidal notch morphologies on coastal cliffs is possible.

4. Tidal notch detection

4.4.2 Testing methods, Perachora Peninsula, E Gulf of Corinth

The entire workflow was tested at different sites along the coast of the Perachora Peninsula. This setting has been extensively studied due to the 1981 earthquake sequence that attracted several research groups, and Holocene tidal notches have been described. Kershaw and Guo [2001] recognised five different notch generations ($\sim 2.7$, $\sim 2$, $\sim 1.2$, $\sim 0.4$, and 0 m) at Mylokipy Bay (see Fig. 4.2b). Laser scan data, covering an area of almost $6.5 \times 3.8$ m of the cliff, was processed for curvature analysis. Line feature extraction from Hough transformation confirmed evidence for all five levels (Fig. 4.10a). Obviously, edges from the Canny detector result in many more line features across the scan than from the Fuzzy logic approach. Canny edges produce line structures almost evenly spread from $\sim +1$ m up to the top of the scan window. Only insignificant line features are determined for the lower most part of the scan data. A confirmation of published indentations is only possible because of their known extents. Furthermore, the recently developing notch is only evidenced by Fuzzy logic edges. Thus, Canny edges indicate remnants of tidal notches but are accompanied by noise which is the result of morphological structures of minor significance. Due to the significant number of extracted lines from Canny edges it is hard to identify distinct levels. Also, line orientation is predominantly not horizontal but showing slight inclined trends although only features of $\theta > 8^\circ$ are considered. Structures from Fuzzy logic edges appear much more horizontal. It is noticeable that even Fuzzy logic edge detection method produces line features of considerable length that do not belong to any of the published notch morphologies, yet are located between two published notch levels ($\pm 2.4$ m).

Similar results can be noticed for both sites at Heraion. Kershaw and Guo [2001] identified two notches at the southern part of the Harbour and correlated them to four notches determined by Pirazzoli et al. [1994a]. The output for both sites supports the potential of tidal notch detection from Fuzzy logic edges. Lower parts at the Heraion

Figure 4.10: Results of testing methods along coast of the Perachora Peninsula. The locations of investigated data in each scan are indicated colourising the height levels. Published notches are provided and correlated to the results of line extraction at Mylokipy bay (a), Heraion harbour site (b), and Heraion Lighthouse site (c). Red arrows indicate the position of the roof of known notches. Black arrows point at morphological characteristics that could correspond to new notches.
4. Tidal notch detection

Figure 4.10: See facing page.
harbour site are significantly rougher than from the rest of the scan and produce line structures without significant cluster levels. A horizontal and convex morphology ~2 m a.s.l. evidences the remnant of a notch roof (Fig. 4.10b). Its remains are poorly preserved and only a few line features are extracted from Fuzzy logic edges. However, a conventional 2D profile supports its existence. A notch at +1.8 m in between both published notches might represent a so far unrecognised earthquake event.

At the cliff beneath the lighthouse, Canny edges only produce poor results (Fig. 4.10c). There is only one evidence for a notch provided as a line feature at ~1 m. This feature matches to a convex edge that might represent the roof of a so far unpublished notch just below the lowermost notch identified by Pirazzoli et al. [1994a] at +1.1 m (see also Fig. 4.2d). It is worth noting that the roof corresponding to the notch at +1.1 m was missed by the Canny algorithm. Contrastingly, Fuzzy logic edges provide evidence for the roofs (+3.5, +3.0, +2.0, and +1.3 m) of all four published notches at corresponding heights (inferred mean sea-levels at ~3.2, ~2.6, ~1.7, and ~1.1 m). However, there is also evidence for a further notch roof at ~2.4 m in between two recognised notch horizons. This evidence is supported by both edge inputs following same parameters in the Hough transform and the 2D profile (Fig. 4.10c).

4.5 Discussion and Concluding remarks on tidal notch detection

TLS is a commonly used technique for morphological purposes [e.g. Rosser et al., 2013; Wilkinson et al., 2015]. Due to its flexibility, quality, and accuracy, the resulting data highlights even minor evidence of spatial peculiarities. In this study, the detailed examination of tidal notches preserved owing to tectonic activity and coastal uplift has been undertaken. Thereby, uplift values in the order of a few decimetres are expected in extensional settings [Papanikolaou et al., 2010] and therefore, a high spatial resolution is required and this is offered by the TLS. Furthermore, a mesoscale downward widening of pre-existing tidal notches is likely. The former notch floor as well as biological markers, such as Lithophaga agents, could be overprinted by the newer tidal notch generation. Thus minor but horizontally consistent changes in the surfaces’ curvature might be evidence for sea-level indicators that were eroded along their lower extent over time, or did not have enough time to develop because of short recurrence intervals between uplift events. Thereby, the local tidal amplitude (here: 0.2 m) forms the resolution limit. Traditional profiling with tape measures or laser distance meter [Kázmér and Taborosi, 2012] aims to identify tidal notches from a digital copy of the vertical cliff topography. When corrected for sea-level datum, information about elevation and notch dimensions can be inferred. This includes both horizontal and vertical extent per feature [Pirazzoli, 1986]. Multiple profiles can only be correlated when referring to the same datum. However, spatial variations in cliff topography of closely positioned sites are hard to verify from horizontally stacked 2D profiles, as a consequence of bedrock heterogeneity, local variations of wave action, and/or fault movements [Kershaw and Guo, 2001]. Utilising TLS measurements in notch studies presents the opportunity to collect high resolution spatial data from exposures (even from distance) in a rectified manner, which is not pos-
sible using conventional tape measurement or photogrammetry and SfM approaches [Bini et al., 2014]. Even submerged notches down to 0.8 m are not excluded from TSL surveys when using systems operating at the green-wavelength [Smith, 2015].

The presented workflow aims to detect the roof and/or floor of raised tidal notches by reducing spatial information and focusing on horizontal continuities. Convex patterns, pointing towards the sea, pose evidence for remnants of tidal notches (see Fig. 4.4). The principal curvature analysis highlights such patterns but does not link those to the attributes of two-dimensional orientation or continuity. However, the magnitude of curvature can be utilised to describe significant morphological changes. Such information is input data for edge detection analysis. Herein, two methods of edge detection were tested in order to reduce spatial information towards its varying significance. In computer vision and image processing, the Canny edge detector algorithm depicts a standard operator [Bao et al., 2005] for tracking ridges in gradient magnitude images [Canny, 1986]. A disadvantage of this method is that all extracted edges appear to have the same significance (see Fig. 4.7b). Thus, edges in areas of minor interest and oriented both vertically and horizontally, appear the same as those of relevance for tidal notch detection. Therefore, a Fuzzy logic sequence was constructed comprising of membership functions that enable exclusive focus on significant horizontal changes in surface curvature. Even if the input information is incomplete or imprecise, the approach outputs predominantly continuous and horizontally oriented structures. Instead of crisp boundaries between two classes (e.g. edge or uniform), the membership functions are defined to give probabilistic information on edge existence (see Fig. 4.7c). However, resulting edge information from both algorithms were individually used as input data for the final Hough transform, which intends to extract continuous line features. Missing points on the desired curves as well as spatial variations between the ideal line and the noise edge points are the result of imperfections in either the image data or the applied edge detection algorithm. The Hough transform produces discrete parameter space matrices of the spatial data in which voting peaks indicate a continuous line object. Furthermore, minor restrictions to the objects orientation yield in spatial matching of identified lines and tidal notch extents (see Fig. 4.9). The ability to adjust the edge detection algorithm for individual requirements, using a Fuzzy logic approach appears to be more reliable for highlighting notch morphologies than the Canny edge detection. Due to the possibility of excluding plan changing curvature and defining membership grades, the line objects extracted from Fuzzy logic edge detection is most suitable.

As mentioned above edge detection and line object extraction target remnants of raised notches, such as their roof and/or floor. This should not be confused with the aims of traditional cliff profiling. Here, the depth of a notch is not analysed and thus the outcome does not allow any conclusion on the developing period as a function of the erosion rate. Only the vertical extent is measurable if the notch is completely preserved. In Agios Pavlos, it is possible to obtain estimates of the tidal range ($\sim 0.35 \pm 0.05$ m) which are consistent with estimates from Evelpidou et al. [2012a]. However, assuming a constant local tidal range throughout the Holocene allows the projection of the historic mean sea-levels with half the erosive zone beneath the detected roof and half above the detected floor, respectively. Hence, historic sea-levels can be reconstructed although the majority of their morphological footprints in a coastal cliff are no longer existent. Furthermore, data collection via TLS enables the extraction of multiple traditional profiles easily for
conventional analyses as well (see profiles in Fig. 4.2) and adds coherent information on the third dimension to address local heterogeneities. Therefore, traditional and presented approaches validate and complete each other from the same data base.

Palaeoseismological studies are frequently assisted by tidal notch investigations in areas of coastal tectonic activity [e.g. Kershaw and Guo, 2001]. In particular, in extensional tectonic settings the footwall coastal uplift is not likely to exceed several decimetres [e.g. Papanikolaou et al., 2010]. However, Pirazzoli et al. [1994a] identified a series of four tidal notches of Holocene age at Heraion (Fig. 4.2d), each displaced by repeated uplifts of about 0.8 ± 0.3 m. Assuming a ratio of 1/4 net slip per event, this would equate to 4 m total offset in an area where Jackson et al. [1982] reported just minor coseismic uplift of 0.2 during the Alkyonides earthquake sequence (M 6.4–6.7) in February and March 1981. If evidence for remnants of tidal notches in between more distinct features are detected by using high resolution data in high performance algorithms, palaeomagnitude estimates get more realistic. For instance, both Canny and Fuzzy logic edges provided evidence for notch roofs at +1.0 and +2.4 m at the cliff beneath the lighthouse, respectively. These positions fit in the idea of regular displacements during earthquakes and reduce mean notch offset yielding reliable values of coseismic uplift (0.5±0.2 m per event). A second example is obtained at Mylokokpy. Including additional notch roofs (−0.6, −1.3, and −2.25 m) would result in repeated uplifts of about 0.4 ± 0.18 m corresponding to magnitudes of M 6.7 ± 0.1 in accordance with Wells and Coppersmith [1994]. The results help to reconcile the discrepancy between the palaeoseismic record and the direct observations of co-seismic displacements provided by Jackson et al. [1982]. Minor but horizontally continuous remnants revealed by dense point cloud data are usually not validated in single 2D profiles. However, the identification of new notch levels would (partially) solve the paradox between large tectonic uplift values and plausible palaeomagnitudes.

The results show the possibility of tidal notch detection by curvature analysis and subsequent edge detection and line feature extraction. It is shown that morphologies accepted as tidal notches can be detected by reducing high resolution point cloud data towards the principal curvature pointing at the roof or the floor of a notch, respectively (see Figs. 4.9c and 4.10b). Even evidence for previously unidentified structures are extracted from the data. As a consequence more realistic uplift values would result if these features get proven as remnants of tidal notches. The workflow enables the objective validation of observations along coastline by evaluating coastal cliffs in three dimensions. Therefore, reliable statements on coast uplifting earthquake events are possible. The variability of conventionally collected tidal notch profiles [Kershaw and Guo, 2001] is circumvented by instant 3D data collection in high resolution and applied spatial analytics. Furthermore, the semi-automated workflow provides fast results once adjusted for individual needs. The benefits are as follows:

- Enhanced objectivity in recognising tidal notch morphologies on cliff faces.
- More insights from high-resolution 3D TLS by recognising undiscovered notches or features corresponding to multiple notches.
- Valuable information on morphological characteristics even of only minor distinction and their spatial distribution especially in extensional tectonic settings, where coseismic uplift is much less than in compressional environments.
However, data quality and thus the reliability of the outcome remain dependent on the preservation of individual tidal notches on a coastal cliff. Sheltered sites in microtidal seas provide perfect conditions for tidal notch preservation after emergence whereas inhomogeneous and disturbed cliffs exposed to the open sea [Pirazzoli, 1986] are not likely to be good archives of Holocene earthquake events. Furthermore, varying bedrock consistency or the presence of bedding planes may yield in the formation of minor structural notches. Especially when the bedding is horizontally oriented, misinterpretation by remote morphological analysis cannot be neglected [Kershaw and Guo, 2001]. This implies that along coast a natural variance of tidal notches masked by surf processes and inhomogeneities yields different results of tidal notch identification. Therefore, careful site selection for palaeo-shoreline identification should consider constraints of marine attacks, tectonic influences on- and offshore and coastal geology. In order to consider such local lateral variations, 3D data acquisition helps to reduce sources of misinterpretation. Therefore, we show that TLS combined with up to date post-processing edge analyses can form a rigorous and useful approach to the interpretation of palaeoseismic records from Holocene tidal notches.
As far as the laws of mathematics refer to reality, they are not certain, and as far as they are certain, they do not refer to reality

A. Einstein (1879-1955)
CHAPTER 5

Numerical modelling of tidal notch sequences on rocky coasts

Tidal notches have had the potential to form at sea-level from ~6.5 kyr BP in the Mediterranean basin and preserve a symmetrical shape comparable to a quadric polynomial. Continuous erosion, predominantly by biological agents, affects a limestone cliff face from low-to high-tide level at <1 mm/yr. Statically determined, the roots of a quadric polynomial are defined by the tidal range representing the limits of effective erosion. However, gradual variations of eustatic sea-level rise (slow) and coseismic uplift/subsidence (fast) in tectonically active regions contribute to vertical shifts in the erosional base at coastlines. As a consequence, the cliff morphology gets modified through time resulting in widening, deepening and separation of notches and possible overprinting of older features. In order to investigate successive modifications of coastal cliff morphology, we developed a numerical model that considers the erosion rate, the erosion zone relative to sea-level, the regional sea-level curve, and tectonic motion. The results show how slow and rapid sea-level change bias the modern cliff face, and highlights that the present-day notch sequence from top descending to sea-level is not inevitably of decreasing age. Furthermore, the initiation of notch formation is not necessarily linked to the date of a certain seismic event. Especially in extensional tectonic settings where coseismic uplift is low and coastal morphological marks are not as distinct, knowledge about coastal evolution is beneficial for paleoseismological research.

5.1 Why model coastal cliff evolution?

Tidal notches are a generally accepted sea-level marker [e.g. Pirazzoli et al., 1982, 1989, 1991; Laborel et al., 1999; Kershaw and Guo, 2001; Evelpidou et al., 2011a,b, 2012b,a; Boulton and Stewart, 2015; Antonioli et al., 2015]. Ongoing horizontal erosion of chemical, physical, and biological agents [e.g. Furlani et al., 2011; Antonioli et al., 2015; Evelpidou and Pirazzoli, 2016] contributes to notch formation at mean sea-level. As a result, obvious ecological and morphological topographies that range from a few centimetres up to several
metres deep occur predominantly on limestone coastlines [Pirazzoli, 1986]. It is generally assumed, when these features are raised or submerged from present-day sea-level, that a paleo-historic tectonic activity can be inferred from obtained sequences. In particular, tidal notches along coasts of the Mediterranean Sea have been an important marker of coastal tectonism determining rates of Holocene tectonic uplift [e.g. Pirazzoli et al., 1982, 1989, 1991, 1994a; Stewart and Vita-Finzi, 1996; Rust and Kershaw, 2000; Kershaw and Guo, 2001; Evelpidou et al., 2012a; Antonioli et al., 2015; Goodman-Tchernov and Katz, 2016] (Fig. 5.1). However, it remains unclear as to what present morphologies can reveal regarding the paleomagnitudes and coseismic uplift of historic earthquakes. It is generally assumed that tidal notches form during relative sea-level stagnation; when vertical land movements and eustatic trends are unison. The database of Boulton and Stewart [2015] demonstrated that the formation of tidal notches is not linked to periods of stable or unstable climates in the past, rather it is likely that tectonic activity and earthquake clustering control the spatial and temporal distribution of tidal notches. Only rapid offset between the strandline and erosional base can form a new notch generation. Thus, the distinction between surface displacing potential in compressional and extensional tectonic settings is absolutely essential [Schneiderwind et al., 2017]. Yet, coseismic offsets on normal faults are at least an order of magnitude smaller than those from thrusting events. However, the misconception that multiple and stacked notches are evidence for meter-scale coseismic events produced by normal faulting still persists [e.g. Stewart and Vita-Finzi, 1996; Cooper et al., 2007; Boulton and Stewart, 2015].

Only a few attempts to model tidal notch formation have been undertaken. Conceptual notch formation is understood as a reflection of normal distributed erosional potential resulting in a symmetrical shape with the retreat zone of maximum convexity at mean sea-level. Pirazzoli [1986] developed the generally accepted idea of a symmetrical V-/U-shaped notch profile on a sheltered cliff where the floor extends to the limit of permanent immersion at tidal low stand, and the roof marks the upper limit of frequent high tides. The maximum retreat point is located near mean sea-level. Gradual relative sea-level

Figure 5.1: Study areas throughout the Mediterranean. a) Plate tectonic setting of the Mediterranean [modified after Faccenna et al., 2014]. Inset shows vertical motion velocity derived from continuous GPS stations provided by Serpelloni et al. [2013]. Obvious coherent patterns of uplift (e.g. Alps) and subsidence (e.g. Spain) demonstrate the tremendous diversity of the Mediterranean geodetic field. Boulton and Stewart [2015] provided a database on tidal notches (rectangles) in the eastern Mediterranean basin. Estimates on regional and local uplift rates range across from -1.09–2.4 mm/yr [Collier et al., 1992; Westaway, 1993; Stewart and Vita-Finzi, 1996; Stewart et al., 1997; Zazo et al., 1999; Stiros et al., 2000; Sivan et al., 2001; Leeder et al., 2003; Zazo et al., 2003; Lambeck et al., 2004; McNeill and Collier, 2004; Westaway et al., 2004; Antonioli et al., 2006; Cooper et al., 2007; Ferranti et al., 2007; Carcaillet et al., 2009; Roberts et al., 2009; Cundy et al., 2010; Schildgen et al., 2012; Roberts et al., 2013; Harrison et al., 2013]. Stars indicate test regions: (1) western Gulf of Corinth, (2) eastern Gulf of Corinth, (3) eastern Sicily and Calabria, (4) southern margin of the central Anatolian Plateau, (5) Samos Island, (6) Tuscan coast, (7) Carmel Coast, and (8) northern Adriatic. (continued)
Figure 5.1: (continued from facing page) Earthquake (EQ) data covering < 500 events (1905-2015) with a maximum focal depth of 20 km is provided by the USGS. AA = Alpine Arc, DA = Dinaric Alps, CA = Carpathian Arc, BC = Betic Cordilleras, HSZ = Hellenic Subduction Zone, NA = Northern Adriatic, NAF = North Anatolian Fault, RG = Rhine Graben. Red box indicates extents of b. b) Geodynamics of central Greece. Faults were compiled from Koukouvelas et al. [1996], Papanikolaou and Papanikolaou [2007], Papanikolaou and Royden [2007], Sakellariou et al. [2007], Roberts et al. [2009], and Grützner et al. [2016b]. Strain rates are from Hollenstein et al. [2008]. Stars indicate the location of the test sites for model comparison with actual cliff faces from the Eliki fault (E) and Cape Heraion (H).
change may produce a variety of tidal notch profiles. Evelpidou et al. [2011a,b] provided a set of graphic schemes of tidal notch profiles resulting from different combinations of relative sea-level changes. In general, the authors pointed out, that relative sea-level stability deepens the notch whereas gradual sea-level change widens the morphological incision. Furthermore, rapid sea-level changes can be divided into two categories. Firstly, where rapid relative movements greater than the tidal range result in notch formation while the former notch remains preserved, and secondly rapid displacements smaller than the tidal range that produce notch profiles with two closely located vertices separated by a small undulation in between. In other words, the pre-existing morphology gets modified due to overlapping erosional zones prior to and after the displacement [see also Pirazzoli, 1986; Evelpidou et al., 2012a]. Notch profile modification is also a product of increasing exposure to wave action. Other than bioerosive agents, cliff quarrying by wave action is generally not considered in tidal notch development. It is generally assumed that quarrying is insignificant for sheltered exposures [Pirazzoli, 1986; Antonioli et al., 2015]. However, Larson et al. [2011] introduced an analytical, yet physically based, model that considers wave impacts on coastal dunes and cliffs from laboratory experiments. Their results show complex feedbacks in cliff notch evolution when nearby beaches provide sediments that increase the erosive capacity of impacting waves. A third approach is presented by Trenhaile [2014] focusing on notch formation by tidal wetting and drying cycles and salt weathering. Here, notch profiles were produced within the 3,000-6,000 year period of constant relative sea-level. As a result of ongoing erosion affecting the same cliff section, several iterative cliff collapses were generated. Wetting and drying cycles as well as salt weathering attain importance especially when saline water penetrates into structural discontinuities of the bedrock. Evaporation processes and subsequent cumulative deposition of salt crystals trigger fragmentation of the rock and result in geomorphic modifications. A similar process occurs owing to frost weathering in cold climates [Trenhaile and Mercan, 1984]. By applying a gridded mathematical model Trenhaile [2016] suggests for limestone notch profiles in the Mediterranean that notch morphology is the product of a variety of local- (e.g. cliff slope and bed resistance to erosion) and regional-scale (e.g. varying erosional efficacy) factors. By adding a wide range of different variables (e.g. variable slope gradient and notch collapse on a local scale and general influence of sea-level changes on a regional scale) a theoretical approach is provided suggesting that similar profiles can be produced by different combinations of applied parameters.

All previous notch models do not address actual changing glacio-hydro-isostatic conditions during the Holocene. Although previous studies generally consider changing sea-levels, both rapid and gradual, unlocking the temporal interplay between sea-level change causative factors has not yet been deeply investigated. However, considering actual and region-specific parameters enhances the understanding of the development of palaeoshorelines and their deformation by active tectonics particularly for palaeoseismological studies. Therefore, previous models are herein described as static (theoretical) models. In order to visualise the development of notch sequences incorporating eustatic and isostatic balances, erosion rates, coseismic uplift, and cliff steepness, we present a simple numerical model that simulates the migration of the erosional base through the Holocene. Furthermore, local sea-level curves and coastal uplift rates for eight regions across the Mediterranean Basin act as input parameters in order to verify potentials of notch formation and associated theoretical palaeoseismological significance when earthquake activity is introduced as
well. Both slow and rapid relative landmass displacements interplay through time causing overprinting and modification of pre-existing notch generations. As the first application in this manner, the time-sliced visualization enables researchers to have an enhanced understanding of tidal notch sequence evolution, and thus better interpretations of co-seismic sequences on tectonic coasts.

5.2 Contributors to notch sequencing

The term tidal notch refers to a horizontal erosion feature at sea-level [Kelletat, 2005b] due to the coeval action [Antonioli et al., 2015] of biological, chemical, and physical factors Pirazzoli [1986]. Pirazzoli and Evelpidou [2013] consider only tidal notches that exclusively formed by bioerosional processes in sheltered places, whereas other publications endorse that the biological component at least dominates notch forming erosion potentials [Evelpidou et al., 2012a; Antonioli et al., 2015]. Frequently submerged by periodic tides, horizontal galleries of endolithic bivalves are most active in the midlittoral zone that extends across the tidal range [e.g. Pirazzoli, 1986; Evelpidou et al., 2012b]. Other organisms that contribute to bioerosion within the midlittoral zone are eroding cyanobacteria, limpets and chitons [Laborel and Laborel-Deguen, 2005]. Thereby, sheltered and vertical exposures are promising locations for the preservation of symmetrical sea-level markers.

Antonioli et al. [2015] point out, that salt weathering, wetting and drying cycles, the potential of karst dissolution, and wave action also play important roles in notch formation. The occurrence of a spray zone in more exposed sites introduces a physiochemical erosion component in terms of salt weathering, where the deposition of salt crystals and hydration will modify the notch shape. Porter et al. [2010] demonstrated that intertidal wetting and drying and salt weathering is also possible. Dependent on the frequency and duration of tidal immersion and exposure intervals periods for salt crystallization within cracks and fissures are formed supporting this type of haloclastic weathering. Chemical erosion through the dissolution of carbonates is not a common effect of seawater exposure, which is (over-) saturated with CaCO$_3$ [Kelletat, 2005b]. The content of calcium carbonate may be lowered only in very localized coastal sections next to springs that show evidence of solution by effluent groundwater [Evelpidou et al., 2012a]. Indeed, nearby freshwater sources support karst dissolution and therefore increase the erosion rate [Evelpidou et al., 2015; Evelpidou and Pirazzoli, 2016]. The vulnerability to different types of physical erosion on coastal cliffs is influenced by the resistance of the rock to wave attack, which is a function of lithology and structural discontinuities, such as cracks, fissures, joints, bedding planes and faults [e.g. Kershaw and Guo, 2001; Trenhaile, 2014, 2015]. The rock is even more affected when turbulent water contains air that gets compressed when smashed against the rock and causes cavitation pitting [Antonioli et al., 2015]. However, cliff collapses are rare for Mediterranean limestone coastlines [Trenhaile, 2016]. Thick-bedded neritic limestones support the overburden and hence the preservation of decimeter-scale deep incisions (Fig. 5.2). Furthermore, most of these Mesozoic limestones are often not deformed by tectonics; e.g. the massive Parnassos and Gavrovo-Tripolis Units crossing the Corinthian Gulf in central Greece comprise of 1.5-3 km thick neritic mostly undeformed limestones [e.g. Papanikolaou, 1984; Papanikolaou and Papanikolaou, 2007].

As a function of the erosion rate, the period of balanced eustatic sea-level rise and
Figure 5.2: Examples for deeply incised notches withstanding cliff collapse in the Mediterranean. Thick-bedded Triassic – Lower Jurassic neritic limestones belonging to the Boetia Unit along the coast of the Perachora Peninsula (eastern Corinthian Gulf, Greece) can support the overburden even when incised up to ~2 m. Profiles are indicated by white dashed lines. a) and d) show raised shorelines closely located to Cape Heraion. b) A sequence of different raise shorelines also at Cape Heraion. c) Incised Cliff at Sterna (see Fig. 5.1b).
isostatic regional uplift controls how deep an indentation develops. However, eustasy, isostasy, and vertical tectonic movements exhibit considerable spatial and temporal variability throughout the Holocene [Lambeck et al., 2004] (Fig. 5.1). Boulton and Stewart [2015] compared local sea-level curves with associated regional uplift estimates and concluded that the highest elevation tidal notch on uplifting coasts should date to ~6,000 yrs BP. Not until that time did the rate of eustatic sea-level rise decrease to ~1 mm/yr and reach gravitational equilibrium with the continental lithosphere [Carminati et al., 2003; Stocchi et al., 2005]. In his modelling approach Trenhaile [2016] concludes, that notches develop as long as sea-level change is no greater than 5.6 mm/yr. For the Mediterranean, sea-level rise decreased to this rate ~6,800 years ago. Subsequently, slow relative sea-level changes have caused gradual changes of the erosional base at emerging coastlines.

By contrast, discrete notch levels record abrupt shoreline changes caused by local seismic displacements. In order to preserve the shape and fragile inter-tidal fauna, rapid removal from the tidal zone and lift beyond the reach of waves is needed [Boulton and Stewart, 2015]. A distinctive example for a prominent and well-developed strandline, that was rapidly offset by up to 9 m asl, is carved into coastal outcrops of western Crete and attributed to the huge 365 A.D. M8.5 Hellenic earthquake [Pirazzoli, 1986; Shaw et al., 2008; Stiros, 2010]. However, in rifting regions shallow normal faulting events of $M\leq7$ commonly produce coseismic uplift limited to a few decimeters along the footwall of the causative fault. Along such faults the uplift/subsidence-ratio is estimated to be 1/2 to 1/4 of net slip per event [e.g. Stewart and Vita-Finzi, 1996; Armijo et al., 1996; McNeill et al., 2005; Papanikolaou et al., 2010]. Even in microtidal environments, such as the Mediterranean Sea, rapid displacements due to coseismic uplift most likely do not exceed the tidal range.

Therefore, as a consequence of both, slow and rapid variations in the position of the erosional base, notch shape modification occurs. To distinguish between notch widening and new notch development is challenging (Fig. 5.3). It has to be expected, that the time period for notch formation might be short and the resulting indentation is only of minor scale, and that massive overprinting and degradation of older features has occurred since ~6,000 years BP.

In order to evaluate stagnation and shifting of the erosional base projected on a present-day cliff face, the long-term geodetic motion should be considered. However, the vertical component of the Mediterranean geodetic field varies dramatically. Continuous GPS stations all over Europe highlight the presence of spatially coherent patterns of uplift and subsidence (Fig. 5.1). Serpelloni et al. [2013] presented up to 14 years of vertical GPS ground motion rates for the Mediterranean region. Their results show that the fastest subsidence of ~3 mm/yr is located in southern Spain, while general uplift (~2 mm/yr) is obtained for the Alps. Furthermore, the dataset indicates landmass uplift of ~1 mm/yr towards the eastern part of the Mediterranean Basin, such as for the island of Crete and the Cyclades. However, the network density here is significantly lower than in central Europe, thus the vertical deformation is less well constrained for the eastern Mediterranean. In addition, the precision of vertical positions determined by most GPS station is ~1 mm/yr [Serpelloni et al., 2013; Facenna et al., 2014] and observation periods are small in comparison to geological timescales [Papanikolaou et al., 2005].

Long-term Quaternary activity is generally reflected in coastal geomorphology, includ-
Figure 5.3: Logic tree for tidal notch sequence evolution. The static notch formation model incorporates only the erosion rate (ER) to estimate the notch depth. The dynamic model considers gradual sea-level (SL) changes due to unbalanced eustasy (E) and isostasy (I), and coseismic land displacements. Resulting cliff shapes contain widened notches (a), emerged notches (b), or a combination of all that (c).
ing uplifted Pleistocene marine terraces and notches of Holocene age in steep calcareous cliffs. Benefits of dating such features are that they represent approximations of cumulative rates over multiple seismic cycles [McNeill and Collier, 2004]. Nevertheless, variations in vertical movements across the Mediterranean region are also presented by several studies (Fig. 5.1). For the western Mediterranean only very minor uplift rates are obtained [e.g. Zazo et al., 1999, 2003]. In the central region, predominantly concentrated at the coastlines of Sicily and southern Italy, several studies have been undertaken calculating uplift rates ranging from 1.0 – 2.4 mm/yr [e.g. Westaway, 1993; Stewart et al., 1997; Antonioli et al., 2006]. The rapidly extending Corinthian Gulf produces Holocene uplift rates of 0.3 mm/yr in the eastern parts up to 1.5 mm/yr in the most western parts [e.g. Stewart and Vita-Finzi, 1996; Leeder et al., 2003; Cooper et al., 2007]. While close to the Hellenic Subduction Zone (HSZ) Roberts et al. [2013] calculated an average uplift rate of 1.0-1.2 mm/yr derived from marine terraces at the island of Crete. For the southern margin of the central Anatolian Plateau (CAP) estimates for Holocene uplift range from 0.6 to 0.7 mm/yr [Schildgen et al., 2012]. By contrast, the Levantine coastline is assumed to be tectonically stable for the past 10 ka [e.g. Sivan et al., 2001, 2004; Goodman-Tchernov and Katz, 2016].

5.3 Dynamic notch formation

In the first instance, the rate of relative sea-level change determines whether a tidal notch will develop or not. For the Mediterranean, estimates of limestone erosion rates range from 0.2-1.0 mm/yr [Pirazzoli and Evelpidou, 2013; Evelpidou and Pirazzoli, 2016]. Thus, balanced conditions between eustasy and isostasy have to persist for about 200 years to develop a significant 20 cm deep notch. The notch height would approximately equal the tidal range for which estimates range from 0.3 – 0.4 m for the majority of the Mediterranean [Evelpidou et al., 2012a; Evelpidou and Pirazzoli, 2016], although exceptional tides of up to 1.8 m may occur in the northern Adriatic [Trenhaile, 2016]. Shape modification is given by exposure, and/or organic accretions [Pirazzoli, 1986; Antonioli et al., 2015]. Increasing exposure yields an upwards shifted roof while the base remains at low tide level. Thus, the height of a notch is primarily controlled by exposure to wave action. Biological accretions are located at the base of a notch. If present the lower part of the notch is no longer a mirrored copy of the upper part but reduced in size. However, a simplified description is supported by sheltered conditions and the absence of biological accretions. Then, the coefficients of a quadric polynomial can cover the requirements to describe such symmetrical shapes (Fig. 5.4). As in a conceptual static model, notch depth is specified by erosion rate [mm/yr] × time [yr] of a constant erosional base.

The dynamic model (considering actual relative sea-level change; Fig. 5.3) calculates the parabolic erosion for every year considering both rapid and slow relative sea-level changes and computes the cumulative sum of erosional impacts. Using a local sea-level curve and information regarding ongoing isostatic and dated coseismic uplift as inputs to control the migration of the erosional base enables us to describe the vertical cliff morphology at a given moment.
Figure 5.4: Assumptions leading to a quadric polynomial to cover requirements of tidal notch shape description: a) actual tide gauge data from April 2016 for the eastern Gulf of Corinth provided by the Hellenic Navy Hydrographic Service. \( \rho \) is the tide period. b) the diurnal tide equals a sine function in a long-term average. Plotted into an unit cycle the moment \( \rho \) depicts an angle pointing to an associated height within the tidal range. Each height depicts a bin in the histogram plot; c) a quadric polynomial describing the depth \( f(z) \) along a symmetrical notch profile. Floor (\( F \)) and roof (\( R \)) depict the roots separated by the tidal range (\( TR \)) along the \( x \)-axis. Note, the \( x \)-axis is labeled as ‘\( z \)’ for a better understanding since it represents the vertical orientation of the cliff face. The erosion rate (\( ER \)) corresponds to \( c \) and determines the depth of the notch after one year.
5.4 Methodology to model a tidal notch profile

The modelling algorithm developed here incorporates eustatic and isostatic balances, erosion rates, cliff steepness, and coseismic uplift. Therefore, the input parameters that have to be specified for the model are: I) the tidal range [m]; II) the erosion rate [m/yr]; III) a long-term coastal uplift rate [mm/yr] and eustatic information [e.g. Lambeck and Purcell, 2005], and IV) the average steepness of the cliff in degrees. Values of coseismic uplifts [m] from one or more events are optional, but have to be linked to a number of years BP if given.

The initial conditions in the model are characterized by an infinite cliff. The influences of the cliff slope are widely discussed in Pirazzoli [1986, Fig. 3] yielding in asymmetrical notch shapes when different from a straight 90-degree. In order to run predictions from the most favorable conditions [Pirazzoli, 1986] the initial setting corresponds to situations where the cliff is vertical.

Following the static model idea, notch development occurs at sea-level and does not change through time. The depth \( d(z) \) of the notch at different stages is expressed as a quadratic polynomial as follows:

\[
d(z) = az^2 + bz + ER \times dt
\]  

(5.1)

where the coefficients \( a \), \( b \), and \( ER \) define the co-domain and shape of the graph in accordance to the erosional base. The erosion rate \( ER \) provokes a translation in \( y \)-direction and represents the eroded depth after one year \( (dt) \) at mean sea-level. The parameter \( b \) is the gradient at the \( y \)-axis. However, when mean sea-level is set to zero the inflection point is located at the \( y \)-axis. Hence, the gradient \( b \) is zero. The curvature at the inflection point is defined by \( a \). Thereby, the roots at floor and roof, respectively, act as targets the graph has to pass. Since these are given by the extents of the tidal range, they can be utilized to transcribe \( a \) as a function of the erosion rate as follows:

\[
a = \frac{f(z)}{(eb - z_{s1}) \times (eb - z_{s2})}
\]  

(5.2)

Here, \( f(z) \) is the determined depth in succession of one year of penetrating erosion. \( eb \) depicts mean sea-level and is set to zero. The roots \( z_{s1} \) and \( z_{s2} \) represent the floor and the roof of the notch and hence obtain values of half the tidal range with different sign.

In order to consider the change of the erosional base through the Holocene, the difference between a given sea-level curve and applied coastal uplift is included. When correcting a sea-level curve for the isostatic trend, the time of notch formation is indicated for periods where the gradient is \(~0\) (Fig. 5.5a). Afterwards, the Cartesian grid is translated for every year, so that the origin always equals the erosional base. Thereby, the grid translation circumvents complex recalculations of \( a \). Indeed, the result is an examination per year. Therefore, a matrix (size: elevation, years BP) is generated incorporating the cumulative sums of each simulated profile. Consequently, the last array of the matrix represents the modelled notch sequence covering the entire vertical extent of affected parts during the simulation. Furthermore, by calculating the erosional base it is also demonstrated that the erosive zone does not project below -3 m (Fig. 5.6).
5. Tidal notch modelling

![Image](image.png)

**Figure 5.5:** Applied sea-level curve and coastal uplift (inset graph) resulting in tidal notches: a) sea-level curve [e.g. Lambeck and Purcell, 2005] corrected for coastal uplift; b) resulting notch sequence.

assume that this is shallow enough to exclude larger scale morphologies such as always submerged shore platforms from the model.

In order to simplify the model, we refrain from considering influences by wetting and drying cycles and salt weathering, predicting vertical cliff profiles only for sites with high potentials of tidal notch development and preservation. Only those shorelines are commonly used to infer seismic history. Following the suggestions from Pirazzoli [1986], cliffs sheltered from wave action located in a microtidal environment pose ideal sites for tidal notch development with a minimum of shape modification and best preconditions for tidal notch preservation. For the central and eastern Mediterranean low-moderate wave energy potentials with mean values around 6-7 kW/m are presented in Besio et al. [2016]. Altimeter significant wave height measurements suggest mean values of ~1 m for the entire Mediterranean Basin [Queffeleu and Bentamy, 2007]. In calm and semi-enclosed sub-basins within the Mediterranean such as the Tyrrhenian coast, the northern Adriatic, the Ionian Aegean Sea, and the Levantine coast, wave heights are in the order of up to 0.6 m with wave periods of around 1-5 s [e.g. Ayat, 2013; Liberti et al., 2013]. Furthermore, we are referring to tidal notches which do not have to be confused with other marine...
notches formed by sediment abrasion. Closely located sediment sources such as beaches and strong currents may support the development of such abrasional notches, which do not necessarily correspond to the tidal range. Moreover the amount of bioerosion is minimized in such grinding environments [Kelletat, 2005b]. An important condition for tidal notch preservation is the bedrock lithology. Databases on tidal notches in the Mediterranean [e.g. Boulton and Stewart, 2015] show that they mostly occur in neritic thick-bedded or even massive limestones. Besides that cliff collapse is uncommon in the Mediterranean [Trenhaile, 2016] due to lithological conditions (see Fig. 5.2), considering cliff failure is irrelevant for decoding the cliff evolution of an actual cliff where preserved paleostrandlines can be observed.

5.5 Results

In order to demonstrate how sensitive the algorithm is for differing input parameters and how diverse notch development occurs through time we applied local conditions of eight different regions across the Mediterranean to the algorithm (Tab. 5.1). Furthermore, coseismic activity is included in the model for two specific sites (Fig. 5.1b, Tab. 5.2).

5.5.1 Uplifting coastal regions: Western/Eastern Gulf of Corinth and eastern Sicily and Calabria

The Mid-Holocene sea-level curve for the Peloponnese coast (Greece) from Lambeck and Purcell [2005] shows a monotonically increasing sea-level and does not contain character-
istics such as a mid-Holocene highstand or punctuated parts (Fig. 5.6a,b). At ~7,000 yr BP the rate of sea-level change decreases considerably and potentially forms conditions for relative sea-level stagnation at ~6,000 yrs BP, when applying an average coastal uplift of 1.2 mm/yr [de Martini et al., 2004]. After correcting the curve for the uplift trend, the resulting gradient allows the timing of notch development to be described (see Figs. 5.3, 5.5a). The result of 7,000 years of a vertically shifting erosional base is shown in figure 5.7. Almost 6.8 kyr BP notch formation begins and corresponds to a 15 cm deep notch ~1.4 m above present-day sea-level. A minor variation and sea-level rise ~6.1 kyr BP yields an upward shift of the erosional base. Not until 5.9 kyr BP is the next equilibrium is reached. During the period in between an upward grazing occurs which indicates that the rate of sea-level change is still slow enough to significantly erode the limestone. From 5.9 – 2.6 kyr BP, a period of almost no sea-level change occurs at a corresponding height of ~2 m resulting in an indentation of almost 1.5 m depth and 0.4 m height, at an erosion rate of 0.5 mm/yr. The subsequent gradual and slow lowering of the erosional base until present-day sea-level produces and overprints the first stage of notch formation. From a present-day view a notch appears at ~1.5 m which is actually the result of two erosional phases 6.5 kyr and 2.5 kyr BP, respectively. It should be noted that the first period yields in a 15 cm deep notch that gets heavily overprinted by the second phase. To conclude cliff morphology evolution for the western Gulf of Corinth, an entire sequence of three notches at ~2 m, ~1.5 m and present-day sea-level can develop without any rapid vertical motion of the strandline.

The same sea-level curve forms the input for the eastern Gulf of Corinth simulation. The highest extensional rates of up to 15 mm/yr are estimated for the western part of the Gulf. Long-term vertical movements towards the Alkyonides Gulf are lower; where, a net uplift rate of 0.7 mm/yr is applied following estimates of Stewart and Vita-Finzi [1996] [see also Roberts et al., 2009]. The resultant modelled cliff section is markedly different to that predicted for the western Gulf of Corinth. The trend corrected sea-level curve does not reach its stagnation phase, where the gradient is almost zero, until ~3 kyr BP (Fig. 5.6b). However, at ~6.8 kyr BP the rate of change is small enough so that the erosive potential penetrates almost the same area over a considerable time period at a corresponding height of ~1.9 m. Gradual vertical shifts during the period between ~6.2 and 6 kyr BP graze the rock no deeper than 0.1 m but across ~1 m in height. Hereafter, the rate of sea-level change decreases again increasing the penetration time and supporting the development notch of a small notch. At ~5 kyr BP a third decrease in sea-level gradient occurs, but which is still faster than the 0.7 mm/yr uplift. Corresponding to todays’ sea-level the erosional zone is located ~0.5 m during that stage (Fig. 5.7b) and shifts to the present datum ~3.8 kyr BP. Subsequently, a fourth stage of a lowered sea-level change gradient causes an increasing indentation at a corresponding height of ~0.2 m. However, very minor gradual changes widen the developing notch. The stagnation is then reached ~3

Figure 5.6: Sea-level curves and uplift trend corrected erosional bases corresponding to todays’ sea-level datum. Regions in Greece and Italy experience > 1 mm/yr coastal uplift (a-c). (d-f and h) These regions represent areas of minor uplift. The northern Adriatic coast has continuously subsided since the last glacial maximum (g).
5. Tidal notch modelling

Figure 5.6: See facing page.
kyr BP. This phase lasts for \(\sim 2,000\) years ending up in a final stage of gentle lowering the erosional base. The last 1,000 years are dominated by forming the present-day tidal notch of \(> 1\) m depth that overwrites pre-existing erosive structures.

The sea-level curve for the north-western part of the Ionian Sea also has two main periods of rising sea-level. The transition between both is not as clear as for the Gulf of Corinth but \(\sim 6.5\) kyr BP sea-level rise stopped outpacing an uplift of \(\sim 1.0\) mm/yr [Westaway, 1993; Tortorici et al., 1995; Stewart et al., 1997; Ferranti et al., 2007]. When the sea-level curve is corrected for this uplift trend, the erosional base has two minor rising steps at \(\sim 5\) kyr and \(3.5\) kyr BP, respectively (Fig. 5.6c). The highest erosional level of \(\sim 2\) m dates to \(\sim 3\) kyr BP before it decreases to present-day sea-level. A more detailed view is given by corresponding time slices in figure 5.8. Incision into the cliff began 6.4 kyr BP when the rate of sea-level rise decreased significantly. However, not until 5.5 kyr BP is the gradient low enough to form a distinct indentation at a corresponding height of \(\sim 0.4\) m. The penetration period lasts \(\sim 600\) years when a gradual upward shift of the erosive base for \(\sim 0.5\) m occurs, resulting in a widening of the indentation. Around 4.8 kyr BP the erosion occurs at a corresponding height of \(\sim 0.9\) m and forms a notch of \(\sim 0.5\) m depth at 0.8 m asl. Subsequently, another upward shift \(\sim 3.3\) kyr BP causes a third distinct indentation at \(\sim 1.2\) m asl. An equilibrium state for \(> 1,000\) years yields not only in a resulting notch depth of up to 0.8 m but also overwrites the roof topography of the underlying notch. During the period between 1.8 kyr BP and 1 kyr BP successive lowering of the erosional base causes a downward widening of the latest indentation and erosion of former notch topographies. The subsequent stage is dominated by grazing the cliff downwards and again of overprinting structures that formed \(\sim 4,500\) years earlier. Today, the erosive sequence shows three to four notches (\(+1.2, +1.0, +0.4\), and currently forming) which match up with elongated periods of relative sea-level stagnation but their depths are heavily altered during grazing phases.

5.5.2 Moderately uplifting coastal regions: southern Turkey and eastern Aegean Sea

For the southern margin of the central Anatolian Plateau (CAP, southern Turkey) uplift at 0.7 mm/yr was estimated by Schildgen et al. [2012]. This rate slightly outpaces the local sea-level curve within the period of the last 6,000 years. However, the rising rate of sea-level change is characterized by very minor variations so that the corresponding erosional base evenly decreases from \(\sim 1.4\) m to the present datum (Figs. 5.6e and 5.9a). The development of the cliff morphology is characterized by distinct notch formation for about 1,500 years from 6-4.5 kyr BP, subsequent lowering of the erosional base resulting in extensive downward grazing, and again focused notch formation at \(\sim 0.7\) m from 4-2.5 kyr BP, removing the floor of the earlier feature. A subsequent minor shift of \(\sim 0.1\) m causes a third indentation just below the last. The present-day notch is the result of gradual down-shifting the erosional base.

By contrast, a distinct knickpoint forms at \(\sim 6\) kyr BP along the coast of Samos Island (Fig. 5.6h) where the net uplift is estimated at 0.6 mm/yr [Stiros et al., 2000], which is comparatively low but matches the overall rate of sea-level rise for the last 6,000 years. Consequently, notch development occurs along a very narrow horizon for the Mid-Holocene.
5. Tidal notch modelling

Figure 5.7: Time slices of tidal notch development on rapidly emerging coastlines such as the western Gulf of Corinth (a) and its eastern part accompanied by moderate emergence (b). Blue lines indicate sea-level at that time.
Figure 5.8: Simulated notch formation time slices for eastern Sicily and Calabria which pose examples for moderately emerging coasts. Blue lines indicate sea-level at that time.

period (Fig. 5.9b). Ongoing erosive penetration against the cliff appears at \(~5.8\) kyr BP resulting in a notch of \(~0.7\) m depth at a corresponding height of \(0.7\) m. A minor shift of \(~0.2\) m towards today’s datum causes distinct notch formation from \(3.5\)–\(1.5\) kyr BP. The present day tidal notch forms during the last 1,000 years, resulting in a single composite notch.

5.5.3 Tectonically stable regions: Tuscan coast and Carmel coast

A short period (\(7.5\)–\(7\) kyr BP) of lower rates of sea-level rise followed by again steep rising before sea-level change adjusts at a moderate slope \(~6.5\) kyr BP is the most characteristic part of the Tuscan coast Mid-Holocene sea-level curve (Fig. 5.6d). Lambeck et al. [2004] stated that the shorelines along the northern and central Thyrrenian coasts are largely free from vertical tectonic movements and uplift is only at \(0.2\) mm/yr in the Holocene interval. Therefore, trend correction to estimate the erosional base has only a minor influence on rates of sea-level change. Potentially, tidal notch development initiates \(2.9\) kyr BP without further noteworthy vertical changes. From the time slices (Fig. 5.10) it
Figure 5.9: Time slices from tidal notch simulation in southern Turkey (a) and in the eastern Aegean Sea (b). Both regions are representative for coastlines emerging at significantly less than 1 mm/yr. Blue lines indicate sea-level at that time.
5. Tidal notch modelling

is obvious that considerable grazing of the cliff began \(\sim 6.7\) kyr BP. Approximately 1,000 years later the rate of sea level change lowers again and causes deeper grazing up to \(\sim 0.1\) m. However, \(\sim 2.9\) kyr BP a distinct notch begins to form, at around \(-0.8\) m. Subsequent lowering the rate sea-level rise causes enhanced penetration per level while still distinct widening can be observed. A distinct notch is finally formed at \(\sim 0.7\) kyr BP with the inflection point just below present-day sea-level (Fig. 5.10).

The sea-level at Carmel Coast (Israel) has risen by not more than \(4\) m since \(\sim 6.8\) kyr BP; with only \(\sim 0.5\) m increase during the last 2,800 years. This means that from \(6.8-2.8\) kyr BP sea-level rises in average at \(\sim 0.9\) mm/yr, and afterwards at \(\sim 0.2\) mm/yr. The latter rate coincides with uplift rates estimates by Sivan et al. [2001] (Fig. 5.6f), which is why notch formation is expected to occur during the latest times of the Holocene; and furthermore not higher than at present datum. However, the results in figure 5.10 show that the potential of considerable erosion is already given \(5.8\) kyr BP. The erosional base evenly rises until \(\sim 3.9\) kyr BP and causes a \(1\) m wide band along the vertical cliff with an average period of water contact of \(\sim 300\) years; the resulting depth is \(\sim 0.15\) m. A subsequent period of \(\sim 1,100\) years prolongates the contact time to \(\sim 400\) years due to a lowered rate of sea-level rise. At an erosion rate of \(0.5\) mm/yr the cliff gets deepened by \(0.2\) m during that period (Fig. 5.10b). Since the given tidal range of \(0.3\) m already extends to the corresponding datum at that time, the floor of the most recent notch develops. A complete overlap of both rates, sea-level rise and net uplift, occurs for the last 900 years resulting in a notch of \(>1\) m depth at present-day sea-level.

It is obvious that both regions experience similar evolution of tidal notch development. For both regions the same net uplift values are applied. The differently shaped sea-level curves determine the calculated depth and significance of modelled notches. However, overall evolution and resulting cliff morphology bear resemblance at both parts of the Mediterranean.

5.5.4 Subsiding coasts: NE Adriatic Sea

Coastal subsidence \([-0.35\) mm/yr; Lambeck et al., 2004\] causes steeper gradients within the evolution of erosional levels than coastal uplift (Fig. 5.6g). The trend corrected sea-level curve for the NE Adriatic has average rates of \(\sim 1.2\) mm/yr from \(6-2.7\) kyr BP and \(\sim 0.4\) mm/yr since \(2.7\) kyr. Therefore, indentations are expected to be not as deep as in uplifting regions. Modelled time slices in figure 5.11 reveal that erosion begins \(\sim 6\) kyr BP carving the cliff down by \(\sim 0.1\) m. This means, already by that time the erosional zone affects the same cliff section for \(\sim 200\) years at a modelled erosion rate of \(0.5\) mm/yr. Except for a few minor vertical undulations the resulting cliff is carved by the same rate for the subsequent 3,300 years. At approximately \(2.7\) kyr BP the lowered gradient of erosional base migration causes deeper incision at \(>1\) m asl on the modern cliff (Fig. 5.11). The erosion rate exceeds the absolute vertical motion value which results

Figure 5.10: Time slices of Mid-Holocene notch development in tectonically stable regions. (a) northern Thyrrenian Sea, coast of Tuscany. (b) Carmel Coast, Isreal. Blue lines indicate sea-level at that time.
Figure 5.10: See facing page.
in predominant notching rather than widening of the indentation. However, the recently developing notch at ∼0.2 m asl is 0.3 m deep and is the result of a gradual uplift shift of the erosive tidal zone.

5.5.5 Introducing cliff slope and coseismic activity to the model

Pirazzoli [1986] already pointed out the influence of sloping cliff faces. In particular, cliffs dipping gentler than 90° require more time to develop distinct notch morphologies, and the resulting notch shape will be asymmetrical with short roofs and long floors. However, this is the result of cumulative erosion over several years. The erosional potential remains normally distributed within the tidal range, which is why only the dip of the cliff face has to be adjusted in the model input. Here, a moderate-high angle cliff slope of 80° is chosen causing asymmetrical notch shapes while preserving the ability to form distinct features.

A famous cliff face exhibiting tidal notch morphologies that are repeatedly associated with coseismic events is located at the north-western tip of Perachora Peninsula in the
Alkyonides Gulf [e.g. Kershaw and Guo, 2001; Pirazzoli and Evelpidou, 2013]. Pirazzoli et al. [1994a] identified four shorelines each offset by ≈0.8±0.3 m with estimated recurrence intervals of 1,600 years from dating organic material of three notches at Heraion (tab. 5.2). The strandline that was not dated (+1.7 m asl) by the authors is here calculated in accordance to the estimated recurrence interval and average displacement.

Introducing coseismic uplift events creates a stepwise modification to the emerging coast function in our model (Fig. 5.12a). Dependent on the rate of relative sea-level change the adjustments to the function representing coastal movements cause a prolongation of notch formation or result in the development of a new notch generation. Considerable erosion begins ≈6.6 kyr BP and thus forms no difference to the model without coseismic uplifting events. However, a much higher position (≈+0.6 m) on the modern cliff face than without a correction for relative vertical land movements (≈−2.4 m; see Fig. 5.7b) is obvious. A significant shift occurs until ≈5 kyr BP ending up in notch formation at ≈+2.5 m (asl.) prior to coseismic event 1. The first event lowers the erosional base by 0.6 m. As a consequence the former strandline is lifted above the tidal range and a new notch generation develops at ≈+2.2 m (asl.). At that time (≈4.4−2.4 kyr BP) the rate of sea-level rise is still significantly different from the applied uplift rate causing upward carving and erosion of pre-existing erosional features. Prior to event 2 the erosional base is located at ≈+2.8 m (asl.) forming a second notch generation during relative stagnation of ≈300 years. The accompanying 0.9 m coseismic uplift of event 2 throws the erosive zone back to a corresponding level at ≈+2.0 m. As the modelled scenario without coseismic uplift events already showed (see Fig. 5.7b) the last ≈2,000 years are dominated by stagnation of the erosional base. Thus, in periods between events 2 and 3, and between events 3 and 4 distinct shorelines form at ≈+2.0 m and ≈+1.3 m (asl). The sloping cliff morphology causes an asymmetrical appearance of both notch generations as predicted by Pirazzoli [1986]. Subsequently, event 4 displaces the pre-existing prominent strandline and results in the development of a new notch generation at modern sea-level. The projection of the erosional base on the modern cliff face shows slight shifting prior to ≈2.5 kyr BP. As a result, vertical carving instead of horizontal deepening is the dominant erosive factor. Excluding the recently developing tidal notch the cliff face exhibits three to four distinct indentations whose individual position on a modern cliff are further associated with coseismic events.

A second scenario including coseismic uplift events is modelled for the western Gulf of Corinth. Estimates on coseismic footwall uplift values per event and dated shorelines from Stewart and Vita-Finzi [1996] serve as input parameter additional to those already introduced in section 5.1.1. The authors dated two raised shorelines (1.8 m and 3.7 m asl.) along the Eliki Fault to 1680 ± 130 yr BP for the lower strandline, and 2290 ± 115 yr BP and 2600 ± 265 yr BP for the upper notch. Furthermore, they found no expression which can be attributed to a surface rupturing event that happened 1861 AD. However, from these dates an earthquake recurrence interval of ≈700 years is assumed here. Earthquake events of M6.0-7.0 as found from normal faulting events elsewhere tend to produce coseismic footwall uplifts of 0.2-0.3 m [e.g. Jackson et al., 1982; Papanikolaou et al., 2010]. Therefore, the scenario includes 11 earthquakes accompanied by normal distributed coseismic uplifts ranging from 0.2-0.3 m within a period of ≈8,000 years (tab. 5.2). Events 1 and 2 produced surface displacements during periods of unstable sea-level conditions. Therefore, no offset erosional expressions result from these two events. The
5. Tidal notch modelling

Figure 5.12: Results for the Heraion site in the eastern Gulf of Corinth including coseismic activity. a) Coseismically modified progression of the erosional base. Arrows refer to events listed in table 5.2. b) Modelled notch sequence evolution including coseismic uplifting events dated by or inferred from Pirazzoli et al. [1994a]. Triangles indicate erosional features at corresponding mean sea-level on the modern cliff in between two coseismic uplift events. Blue lines indicate sea-level at that time.
5. Tidal notch modelling

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<td>3</td>
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first notch begins to form ~6.7 kyr BP at ~+3.6 m but widens up to a level of ~+4.1 m because of minor sea-level changes outpacing the coastal uplift. Event 3 lowers the erosional base to a corresponding height of ~3.9 m. The subsequent 700 year period is dominated by stagnation of the relative erosional base forming a distinct notch of ~0.3 m depth (Fig. 5.13b IV). The ensuing period of ~1,400 years contains two events (4 and 5) that stepwise lower the relative erosional base which stays at a gradient of almost zero apart from that. The resulting notches (V and VI) comprise erosional contributions from 3,000 years before. A distinct notch (VII) forms from ~3.8-3.0 kyr BP without noteworthy vertical shifting. During the period between event 7 and 8 notch VIII develops at ~+2.8 m but gets widened due to a small component of gradual lowering the relative erosional base. The same process happens to notch IX that develops at ~2.0 kyr BP. Even more obvious is the widening effect for the last two notches (X and XI), which are still separated from each other due to the successive uplifts from events 9 and 10. However, a sea-level rise slower than modelled uplift causes a gradual lowering of the relative erosional base in the latest Holocene period. The predicted modern cliff morphology shows nine different historical sea-levels ranging from ~+0.7 m to ~+4.0 m. While uppermost notch generations (III - VI) appear stacked and concentrated, younger notch generations (IX - XI) are more spread along the vertical cliff. Presumably, the uppermost notch generations (III-VI) would not be differentiable in the field since their vertical spread is only 0.5 m and thus just exceeds the applied tidal range.
5. Tidal notch modelling

**Figure 5.13:** Results for the Eliki site in the western Gulf of Corinth including the coseismic activity of 11 events. a) showing the local sea-level curve and landmass evolution exhibiting coseismic uplifts. b) modelled time slices of cliff face evolution. Triangles indicate erosional features at corresponding mean sea-level on the modern cliff in between two coseismic uplift events. Blue lines indicate sea-level at that time.
5. Tidal notch modelling

5.6 Discussion

The focus of this study has been to visualize tidal notch formation during the late Holocene, incorporating sea-level change, coastal uplift/subsidence, erosion rates, coseismic activity, and cliff steepness. We have been able to show how coastal cliff morphologies develop within a migrating tidal range [see also Pirazzoli, 1986; Evelpidou et al., 2011a,b; Trenhaile, 2016] using actual late-Holocene sea-level curves for the Mediterranean. The tidal range shifts along a vertical cliff due to gradual relative sea-level changes optionally accompanied by coastal tectonic activity. In the following discussion, the contributions of slow and rapid vertical shifts of the erosional base are discussed with regards to the applied modelling parameters.

5.6.1 Modelling parameters and inputs

The modelling algorithm we present here deals with ideal conditions. Therefore, the results do not represent actual and naturally existing cliff faces.

Firstly, these ideal conditions incorporate a perfectly sheltered site, where tidal range is low and constant [Pirazzoli, 1986]. Assuming the absence of spray enables us to ignore notch roof modification by haloclastic processes above high tide level. In addition, carbonate lithologies sheltered from strong wave action enable grazing or coring organisms to settle and contribute to the erosion in vegetational bands within the midlittoral zone [Torunski, 1979]. Moreover, Pirazzoli [1986] points out that only in sheltered sites does the midlittoral zone equals the tidal range. Since the model calculates erosive effects only for the applied tidal range of 0.3 m [Evelpidou et al., 2012a; Evelpidou and Pirazzoli, 2016], predicted profiles correspond to natural sites without strong wave action. If a different tidal range is used, then the affected parts of the cliff vary accordingly. Hence, a smaller tidal range yields narrower indentations whereas a wider tidal range increases width of a tidal notch [see also Trenhaile, 2014]. When the erosional base varies through time narrow notches appear to have greater separation than wider features.

Secondly, the algorithm intends to model tidal notches formed within the tidal range without significant contribution of wave quarrying, sediment abrasion or chemical weathering. Mechanical wave erosion is of a second order when unbroken waves are reflected from steep cliffs and when a source for abrasive material is absent. Moreover, wave quarrying is most common in storm wave regions and where coastal cliffs are comprised of rocks with structural discontinuities [Trenhaile, 2015]. The majority of carbonate coastlines throughout the Mediterranean do not have plunging cliffs and/or are located next to beaches, or are fronted by wave breaking foreshores. However, steep cliff faces of massive limestones located far from abrasion material such as sands and pebbles exist. Indeed, carbonate cliffs tend to develop karst formations which may be accompanied by freshwater exchange. If so, the sea-water is locally diluted and thus chemical dissolution of calcium carbonate contributes little to the overall erosion rate. Moreover, the solubility of calcium carbonate is higher at low temperatures. This leads to the assumption that during the glacial era lower water temperatures in general and higher precipitation resulted in higher amounts of limestone dissolution. However, Evelpidou et al. [2011b] suggest using the term 'visor' for notch profiles where the base is missing due to chemical dissolution. For such localized phenomena notch developing effective erosion is not limited to the tidal
range and hence those morphologies are not suitable sea-level markers. Herein, model predictions aim to illustrate the evolution of paleostrandline sequences as they are commonly used to infer coastal coseismic activity [e.g. Pirazzoli et al., 1982, 1989, 1991, 1994a; Stewart and Vita-Finzi, 1996; Kershaw and Guo, 2001]. This includes focusing on low energy sites, where bioerosion dominates tidal notch development. Furthermore, late Holocene tidal notches form out in the hot and semiarid environment of the Mediterranean. This circumstance decreases the ability of limestone dissolution, in general.

Thirdly, vertical lithological inhomogeneities are not considered. Varying bedrock consistency and accompanying erosion rates are not modelled. Consequently, the profiles represent homogeneous limestone cliff morphologies resulting from even erosion at 0.5 mm/yr [e.g. Evclidou et al., 2012a; Furlani and Cucchi, 2013; Pirazzoli and Evclidou, 2013]. Dating tidal notches and deriving erosion rates is a challenging task since radiocarbon bearing dating material is of very sensitive organisms that can easily be eroded by various agents after their displacement [e.g. Evclidou and Pirazzoli, 2016]. However, some efforts have been undertaken to derive estimates for erosion rates across the Mediterranean. The applied erosion rate is in good correlation with recently derived estimates from a well-dated fossil tidal notch in Greece (0.64 mm/yr) [Evclidou and Pirazzoli, 2016] and from micro-erosion meter measurements in the northern Adriatic (0.31 mm/yr) [Furlani and Cucchi, 2013]. In general, it should be noted that varying the applied erosion rate in different model runs modifies the predicted indentation depth but not its position [see also Trenhaile, 2014].

Fourthly, modelled notches are not constructed to collapse. Overburden that cannot be supported by the lithology results in cliff collapse, which is basically controlled by the depth of a notch. Trenhaile [2014] includes the ability of cliff collapse in his modeling approach and concludes that failure is mainly dependent on the maximum notch depth. For collapse scenarios the author uses maximum notch depths of 2 m. However, the notch profiles predicted herein do not exceed 1.5 m in depth (see Fig. 5.7a). Moreover, if cliff collapse occurs environmental conditions change likely resulting in significantly different wave action and constitute a possible sediment origin [Trenhaile, 2015]. This would clearly contradict other model assumptions at a certain point.

Fifthly, horizontal differences are not displayed by a two dimensional notch profile. Fault movement leading to differential uplift, local variations of wave and surf regimes, and horizontal bedrock heterogeneity are reasons for differing notch profiles on a local scale [e.g. Kershaw and Guo, 2001].

The five listed caveats imply significant simplifications to the model. Each of above mentioned points has considerable influence on the shape of a notch profile. Furthermore, the combination of all of them is considered to modify a tidal notch for each investigated region individually. However, considering such assumptions enables the development of a mathematical model of a symmetrically effecting erosion potential within the tidal range per year. Then, cumulative erosion depicts the base for the static notch developing model that only distinguishes between continuous erosion per level or not.

Predicting actual scenarios using local sea-level curves and regional landmass movements is a novelty to the assessment of tidal notches as earthquake geological effects. Conceptual models [e.g. Pirazzoli, 1986; Evclidou et al., 2011a,b] indicate shape modification by sea-level change and even already distinguish slow and rapid changes of the
erosional base. Trenhaile [2016] modelled notch formation considering linear and sudden sea-level changes as well. In contrast to the model presented in here, the author also considers changing erosional efficacy, cliff collapse [see also Trenhaile, 2014], and varying rock resistance. However, by simplifying the model assumptions and orienting to specific regions we were able to achieve similar conclusions on the influence of local and regional-scale factors. Furthermore, we prove for the conclusion that similar profiles can be produced by different combinations of incorporating factors [see Trenhaile, 2016]. One difference between both models is the applied distribution of the erosional potential. Where Trenhaile [2016] uses a linear function we consider a distribution following a quadratic polynomial based on repeated immersion of cliff parts due to tides following a sine function in a long-term average (Fig. 5.4). Erosion rates measured at different heights (vertical resolution is 0.25 m) over a 3-year period on a limestone slab in the northern Adriatic indicate that the mean downwearing rate follows a symmetrical shape [Furlani and Cucchi, 2013]. Long-term measurements will show what is the best fit function describing the erosional potential distribution. However, since both models yield to similar results the fitting appears to be of a second order. As a major difference between both models Trenhaile [2016] constructed a more theoretical approach, while we clearly orient at region-specific conditions. Hence, our model provides the opportunity to compare natural occurrences of tidal notch sequences with derived scenario interpretations. Existing investigations and interpretations on coseismic tectonic history might have to be reassessed due to so far unknown consequences concerning submerged notches and/or timing and magnitude of coastal coseismic activity.

Therefore, the most dynamic component in the model is the applied sea-level curve. The shape, rates, punctual characteristics, and the overall richness of details of a sea-level curve form fundamental input for the dynamic model (Figs. 5.6 and 5.10). Commonly, a variety of sea-level indicators are used to reconstruct sea-level history. Typically, these indicators are of biological, sedimentological, erosional, and archaeological remnants [Lambeck et al., 2004; Kellett et al., 2005a]. However, the spatial distribution across the Mediterranean, concentration of certain markers in some places, and differential tectonic activity cause gaps in the availability of local sea-level curves (e.g., Spain, North Africa) [Pirazzoli, 1991] and vary the quality. By contrast, many sea-level curves have been published for shores at southern France, the Aegean, the Levant, and the Adriatic in the past decades. All histories applied here have a significantly changing rate of sea-level rise plotted in the period between 7-6 kyr BP. A global meltwater pulse caused rapidly changing sea-levels (10-20 mm/yr) until 7 kyr BP (Fig. 5.5a), before slow-moderate (0.2-2 mm/yr) rise adjusts in the Mid-Holocene. Lambeck et al. [2014] point out that in the past 6.7 kyr BP only 4 m of global sea-level rise took place, which equals an average rate of ~0.6 mm/yr. Moreover, the authors predict actually two stages of sea level rise, the first going from 6.7-4.2 kyr BP, and the second covering the last Holocene period. Following their estimates, 75 % of Mid-Holocene sea-level rise took place during the first stage (1.2 mm/yr). Sea-level changing at ~0.2 mm/yr during the second stage is broadly consistent with other studies [e.g. Pirazzoli, 1991]. For the Mediterranean Sea a third trend of 1.7 mm/yr covering the past century is predicted by Wöppelmann and Marcos [2012]. However, the last is not considered in this study since the applied erosion rate of 0.5 mm/yr produces indentation of 5 cm only in 100 years of relative stand-still. Furthermore, this high-rate change still does not submerge the cliff from the tidal range of 0.3-0.4 m [Lambeck et al.,
5. Tidal notch modelling

2004; Evelpidou et al., 2012a; Antonioli et al., 2015] within this short period. The local sea-level curves applied to test the algorithm are consistent with overall characteristics described above. However, the relative sea-level at the knick-point around ~6.5 kyr BP varies as well as the timing for the second rate lowering period. While the relative sea-level was ~6-7 m below present datum (Fig. 5.6a-d) in the central Mediterranean since the Mid-Holocene, a ~3-4 m rise occurred in the eastern parts of the basin (Fig. 5.6e,f,h). The second, more minor change generally appears between 3-2 kyr BP and thus later as predicted from the global sea-level curve [Lambeck et al., 2014]. Furthermore, in the central basin this change tends to occur at ~3.0-2.5 kyr BP while the eastern Sea reaches this point ~500 years later [e.g. Sivan et al., 2001].

5.6.2 Solving the issue with submerged notches

In regions of significant tectonic activity reconstructing the sea-level history is problematic since most sea-level indicators refer to some specific part of the tidal range and their displacement by fault activity requires accurate adjustments to resultant vertical motion. Roberts et al. [2009] demonstrated the variability in uplift even over short distances along fault strike. Yet tectonic activity is essential for estimates of long-term landmass uplift/subsidence. Hence, for regions such as the seismically high-active Gulf of Corinth it is unlikely that representative estimates for both, sea-level history and paleo-tectonic rates, will be found. In such cases assumptions and spatial generalizations have to be made; for instance, the sea-level curve for the shore of Peloponnese is representative at least for the western part of the Corinthian Gulf.

Interestingly, Boulton and Stewart [2015] hypothesized that in order to initiate notch formation uplift rates needed to equal rates of sea-level rise. This statement presumes that subsiding coasts will not experience a relative stagnation under conditions of steadily rising sea-levels and that subsiding coasts are not suitable for tidal notch development without more complex tectonic movements involved during the late Holocene.

The difference between relative sea-level and corresponding landmass position forms the relative erosional base projected on a modern cliff face. Different uplift rates applied to the same sea-level curve have a huge impact on the shape of erosional base evolution (Fig. 5.6a,b). As a result, tidal notches form at different periods and appear on different corresponding levels (Fig. 5.7 & 5.14a). Furthermore, a combination of low coastal uplift rates (< 1 mm/yr) and significant sea-level change since the Mid-Holocene yields tidal notches to appear below present-day sea-level. Coasts that are considered to provide stable conditions potentially exhibit submarine notches (Fig. 5.10). Here, the model is confirmed by observations made at the southern Levantine coast by Goodman-Tchernov and Katz [2016]. The authors concluded sea-level history provides a period of relative stagnation, followed by drowning. At a coast that is generally considered to be not tectonically affected, only eustatic characteristics provide potential for notch development.

The results modelled from subsiding coastal conditions show that relative stagnation does not mean sea-level rise and vertical landmass motion have to occur in unison. In detail, modelled cliffs get significantly carved when the erosional base shifts at <1.1 mm/yr using an overall erosion rate of 0.5 mm/yr [see also Trenhaile, 2016]. Moreover, horizontal deepening dominates vertical carving when the difference between vertical land motion
5. Tidal notch modelling

and sea-level rise is $< 0.5$ mm/yr. This implies, a notch in a limestone cliff (erosion rate: 0.5 mm/yr) that develops while sea-level rise or landmass motion dominates by 0.5 mm/yr for about 200 years is $\sim 0.1$ m deep and $\sim 0.4$ m high (including 0.3 m tidal range). When introducing the effect of varying tides, spray, and weathering the interpretation of such an expression would most likely conclude a "relative stagnation" to form it. Even the occurrence of a notch located on a subsiding coast becomes plausible if vertical relative land motion does not exceed the absolute threshold value. Benac et al. [2004] described submerged notches in the northern Adriatic. Their results show well expressed but asymmetric tidal notches always submerged by at least 0.2 m indented between 0.18 - 1.50 m. These values range in the same order as our modelled notch estimates.

5.6.3 The role of coseismic displacement

The modelled cliff sections show the significant impact of the sea-level curve shape on time and duration of notch formation in accordance to a given constant coastal up-lift/subsidence. Dependent on the shape of the sea-level curve the cliff morphology results from grazing, incising, and overwriting only from gradual climatically driven sea-level changes. When introducing coseismic activity to the model even more dynamics are addressed in the system. The abrupt migration of the erosional base potentially yields to the development of an entirely new notch generation. However, in combination with an arcuate shaped erosional base height curve (Fig. 5.6) the migration is not oriented purely in one direction. Furthermore, repeated coseismic activity results not necessarily of the same migration stepsize since rates of a mean relative sea-level change vary from ascending ($\sim 6$ kyr BP) to flat ($\sim 4$ kyr BP), and also descending ($\sim 2$ kyr BP). As a result, lower sections of a modern cliff face potentially formed the strandline at least two times since 7 kyr BP (Figs. 5.6a, b, c, and 5.12a).

In particular, modelled results for the southern Italian coastline show how repeated erosion modifies the developing cliff face (Fig. 5.8). When introducing coseismic displacements to the model, modification is even more apparent. Both scenarios where coseismic uplift was included indicate that coseismic offset possibly results in one of two options: I) rapid displacement of the erosional zone causing the development of an entirely new notch generation, which likely overprints older features to a greater or lesser degree (Fig. 5.13a), or II) prolonging or re-entering the erosive phase at a certain level in periods of gradual sea-level change (Fig. 5.12a). Furthermore, the scenario modelled for Heraion illustrates that a notch sequence on a modern cliff from top descending to sea-level is not inevitably of decreasing age caused by coseismic uplift events in periods of slightly uplift outpacing sea-level rise. Moreover, when rapid coseismically induced displacement prolongs the erosive phase the modern expression cannot be used to infer information about the specific event since it only causes deepening of the pre-existing notch.

5.6.4 Model versus reality

The reliability of both scenarios modelled for the western and eastern part of the Gulf of Corinth should not be overvalued due to idealized assumptions and generalizations according the applied sea-level curve. The two different scenarios of seismological history
are applied since they pose results from different views. The inputs for the Heraion model are inferred from a study that aimed to directly investigate episodic uplift deduced from Holocene shorelines [Pirazzoli et al., 1994a]. The differences between model and natural cliff are most likely the result of the inherent model assumptions, a sea-level curve not specifically estimated for that region, and onshore tectonic activities which are not considered in the model (Fig. 5.14b). However, modelled tidal notches and minor indentation can be correlated to actual observations. For instance, notches modelled to +1.4 m and +2.8 m might coincide with observed expressions at +1.7 m and +3.1 m. Even depth relations between modelled and natural notches resemble each other in appearance. On the natural cliff the notch at +2.6 m forms the deepest indentation of the natural sequence which matches with the notch modelled to +2.0 m. In fact, no significant indentation is modelled in between these notches giving evidence for cumulative offsetting contributed by both, coseismic uplift and gradual relative sea-level change. However, balancing smaller coseismic uplift values from off-shore origin and down-throwing contributions from active on-shore faults potentially result in more paleo-strandlines than observed so far [Schneiderwind et al., 2017].

The Eliki Fault scenario is based on the assumption of a regular seismic cycle with multiple coseismic uplift events each of 0.2-0.3 m [Stewart and Vita-Finzi, 1996]. Uplift values of this range and recurrence interval (not exceeding 1,000 years) are plausible and consistent with paleoseismological principles in extensional tectonic settings [Jackson et al., 1982; Papanikolaou et al., 2010]. The results for this scenario are more consistent with reports of the natural cliff face. Stewart and Vita-Finzi [1996] described prominent notch levels at +1.8 m dating to 1680 ± 130 years BP, and at +3.7 m dating to 2290 ± 115 years BP and 2600 ± 265 years BP. Varying dating results for the upper notch might be the consequence of repeated erosive phases compressed along a thin section of the cliff. In the modelled cliff a tidal notch develops at +1.8 m as a consequence of coastal displacement at 1680 yrs BP and hence misses natural conformity only at one earthquake recurrence interval. The absence of a distinct erosive evidence for an earthquake that happened 1861 AD might be the consequence of an increasing amount of downward carving due to fluctuating gradients (relative sea-level and coastal uplift).

Therefore, our model shows promise for the potential reconstruction of actual cliff faces but needs better and more accurate input values for relative sea-level change and seismic history. However, in accordance with the assumptions made the algorithm combines the results from multiple disciplines and produces cliff faces that can be compared to natural exposures to support interpretation strategies. If the sea-level rise is well constrained, far range deglaciation effects are validated, and information about paleotectonic activity is available, a separation of spatial and temporary segments might be possible. However, to produce more reliable results parameters such as maximum overburden, lithological discontinuities, exposure to wave action, and much more have to be considered [Trenhaile, 2014]. Furthermore, constraining the erosional base does not only include isostatic corrected sea-levels but also tectonic activities on and offshore. In extensional Graben systems such as the Gulf of Corinth several coast-down-throwing normal faults on land potentially influence the erosional level. This research has implications for assessing the overall and local seismic activity of a certain region. Based on reasonable assumptions reliable information on overall tectonic activity as a budget and balanced structure-linked values can be gathered. Due to its simplicity the novel algorithm is transparent and
reproducible increasing the objectivity in assessing coast-affecting tectonic activity.

5.7 Conclusion

Depending on the region and associated local sea-level history, Holocene tidal notches can form from 6,000-7,000 years BP in the Mediterranean Basin [see also Trenhaile, 2016]. Thereby, the very early stages of counterbalanced eustatic and isostatic conditions might not result in the most elevated sea-level marker at the present-day. In detail, modern cliff morphology contains indentations, nips, and deepened sections that are not true notches. This geomorphology is a product of continuous notch formation, repeated overprinting, bedrock heterogeneity, storm surge elevations. Gradual sea-level change optionally accompanied by tectonic activity shifts the erosional base along the vertical axis (see also Fig. 5.3). As a consequence, a notch sequence from top descending to sea-level does not necessarily adhere rigidly to and old-to-young chronology. Stages of almost-stagnation between regional sea-level rise and coastal uplift tend to produce more space between individual
notch generations. However, resulting notch shapes appear widened in comparison to successively older features.

The developed algorithm is not as close to reality as required for a retro-deformation due to significant generalizations and simplifications. However, the model presented is the first that incorporates actual and region-specific Holocene sea-level changes, erosion rates, and landmass movements (slow and rapid). It points out how variable tidal notch development and preservation occurs even in local scale. Paleoseismological studies should benefit from its application since it provides a method for the evaluation of field observations and interpreted meanings. Case studies considering both coastal coseismic footwall uplift from offshore normal faults and coast downthrowing onshore faults could profit by evaluating the balance of relative motions.

In conclusion:

- The algorithm makes clear how slow and rapid processes interplay and bias each other (modification)
- The visualization illustrates how slow and gradual sea-level changes can result in sequences that look similar to those generated with influence of seismic activity
- notch offset exceeding several decimeters in a present-day notch sequence is not a contrariety to typical coseismic coastal footwall uplifts since recurrent overprinting and minor sea-level variation can produce such offsets even without tectonic activity. Therefore, tidal notches may not be used as primary earthquake geological effects without considering detailed sea-level history
- Submerged notches can occur on emerging coastlines
- "relative stagnation" is not a condition with absence of relative motion but comprises a scope of minor motion (<0.5 mm/yr) in dependency of the actual penetration period and erosion rate

The model presented enables researchers to have an enhanced understanding of the evolution of tidal notch sequences and points out how important reliable data of sea-level rise and coastal uplift are to the correct interpretation of such sequences.
Nobody ever figures out what life is all about, and it doesn’t matter. Explore the world.
Nearly everything is really interesting if you go into it deeply enough
R. Feynman (1918-1988)
Making aspects of the world visible is a very basic goal of any kind of mapping by transporting messages. In geoscience, such messages may contain information on the location of an object and/or accompanying attributes. For describing the seismic landscape and obtaining new knowledge from its constituents a cardinal workflow successively incorporates the detection of earthquake geological effects, their visualisation, and data analysis and processing towards individual questions regarding palaeoseismic history. Studies and techniques presented aim to cover involved processes for individual features. In this work, one study is very technical and provides increased insights in palaeoseismological trenches, and two studies, focussing on the appearance and development of tidal notches, contribute to the so far uncertain value of these sea-level markers in palaeoseismology.

More precisely, the methodology presented in Chapter 3 combines terrestrial laser scanning with GPR investigations of the hanging walls sedimentary architecture in order to provide a more objective visualisation of stratigraphic characteristics in palaeoseismological trenching studies by multispectral analyses and 3D imaging.

In Chapter 4 terrestrial laser scanning is adopted as a tool that provides high resolution cliff morphologies for high performance curvature analysis and feature detection. Horizontally continuous line features are interpreted as remnants of tidal notch morphologies even in between major emergence. Thus, the presented workflow helps to clarify the paradox of so far observed metre scale shoreline displacements in contrast to coseismic scaling laws in regions of tectonic extension.

Chapter 5 deals with the visualization of coastal cliff evolution since the late Holocene. By incorporating Holocene sea-level change, erosion rates, slow net movements and rapid coseismic displacement to a newly developed numerical model it is possible to visualise the evolution of tidal notch sequences in time slices for the first time. It clearly images the interplay of slow and rapid processes and addresses issues between palaeoseismological research and morphological observations by using comprehensible mathematic equations. Thus, it helps to clarify debating issues in the literature regarding the development of tidal notches and their use for palaeoseismological and tectonic studies.
However, two questions arise querying the cost-benefit ratio in real-world investigations and not only for theoretical purposes:

- Why use complex multidisciplinary techniques to visualise the seismic landscape at high spatial resolutions?

- What are the implications arising from new knowledge about tidal notch formation and morphology?

### 6.1 Why use elaborative techniques to visualise the seismic landscape?

In order to validate the costs of new techniques and combining methodologies a brief review of technical innovations and their impact on assessing the seismic landscape is presented.

#### 6.1.1 What is the value for future studies?

The compilation of technological milestones in Chapter 1 points out that the majority of innovations and enhancements that are used in palaeoseismology target on the visualisation of surfaces. Even conventional trench logging is partially based on illustrating the surface of trench walls. Only a few technologies, dominantly from geophysical applications, aim to visualise properties of individual materials that are not limited to the surface. While trenching investigations and photomosaic methods have become state-of-the-art techniques to visualise constituents of the seismic landscape, capabilities of modern technologies that aim to visualise and assess physical properties or morphological characteristics appear subordinate. New methodologies comprising innovative workflows adhering to interdisciplinary techniques have been published during the past decade [e.g. Ragona et al., 2006; Sagy et al., 2007; McClymont et al., 2008; Sagy and Brodsky, 2009; Jones et al., 2009; García-Sellés et al., 2011; Zielke and Arrowsmith, 2012; Wiatr et al., 2015; Schneiderwind et al., 2016]. However, the application of efficient workflows remains rare in palaeoseismic publications, although their practice in broadly based investigations clearly increases a study's value in consideration of scientific integrity:

- Technological innovations and methodological enhancements increase the objectivity in assessing palaeoseismological circumstances

- Even detailed and explicit workflows are reproducible

- The application of approved workflows and techniques increases transparency of extensive studies

- Aforementioned essentials contribute to clearness of scientific work
Figure 6.1: The application of terrestrial laser scanning (TLS) and 'structure-from-motion' (Sfm) quantified by their appearance in scientific publications (Elsevier Library). Date: 12/2016.

- Workflows that have proven as robust increase reliability of applied investigations
- The usage of novel workflows is innovative itself

The different workflows presented in Chapter 3, 4, and 5, comply these essentials and provide valuable input for future studies. Considering multispectral imaging (that got its origins in spaceborne remote sensing techniques) during palaeoseismic trenching investigations (see Chapter 3) combines established techniques and clearly increases objectivity in distinguishing lithological units by their physical properties. The application of GPR measurements incorporated in the workflow increases objectivity even more, is easily reproducible since GPR has become a commonly used technology in palaeoseismology, and increases reliability of interpreted results since the lithologic continuity gets validated [Schneiderwind et al., 2016].

Using TLS preferred to Sfm (Fig. 6.1) is based on post-processing requirements and accuracy [e.g. Buckley et al., 2008]. Furthermore, the availability of near-infrared backscatter images offers the visualisation of physical properties that again increases the assessments' objectivity [e.g. Schneiderwind et al., 2016]. However, the majority of TLS practice is commonly limited to the field of extracting digital elevation models and geometrical characteristics representing surface morphologies are oftentimes just descriptively analysed. When fusing distinct advantages of TLS (e.g. accuracy, precision, rectification, accessible range, post-processing efficiency) with technological knowledge (computer vision feature extraction) applied in Sfm a new field of application arises that provides a new dimension of objective data analysis, reproducibility, transparency, clearness, and reliability (see Chapter 4) [Schneiderwind et al., 2017].
6. Discussion

Not imaging actual constituents of a seismic landscape but producing them virtually, visualising what would happen if assumptions were true is the content of chapter 5. For logical reasons reproducibility is the biggest advantage of numerical modelling; a simulation can be run as often as required comprising the ability to optimise the algorithm iteratively. Furthermore, if a model is well documented its application is completely transparent and accessible for interpretations. In the case of tidal notches analytical approaches focusing on actual features are always confused by geomorphological appearance and palaeoseismological dimensions [e.g. Cooper et al., 2007; Schneiderwind et al., 2017]. A solution to decrease existing discrepancies is to create a simplification of the real world simulating the evolution of tidal notch development and hence, to visualise situations that represent the truth if model assumptions sufficiently correspond to underlying driving forces.

6.1.2 Geospatial information on the "seismic landscape" at different scales

So far, only the visualisation of certain constituents of the seismic landscape is discussed. However, in order to assess and evaluate the seismic hazard of an area and to take full advantage of palaeoseismological archives broad-based and multidisciplinary studies are necessary [e.g. Michetti et al., 2005]. Rectified satellite imagery, high-quality SAR data, and high-resolution laser scans, both airborne and terrestrial, provide information about earthquake geological surface effects at different scales ($\sim 10^0 \text{ to } 10^8 \text{ m}^2$). Accompanied by the high-resolution ($\sim 10^{-4} \text{ to } 10^2 \text{ m}$) characteristics modern technologies offer the opportunity not only to investigate single expressions of coseismic activity but to measure and analyse all important (cumulative) ground effects distributed in the seismic landscape of an earthquake-prone area. Contrarily to conventional methods that aim to visualise the seismic landscape, such as sketches or logs, and also other than innovative photogrammetry methods, above mentioned technologies provide rectified and georeferenced data which allow a direct combination and correlation with field data or previously collected data towards multidisciplinary investigations. Even terrestrial laser scans of steep to vertical objects, which cannot be accessed from airborne or spaceborne platforms, can be easily implemented in broader datasets due to geotetic information included by the TLS system. Furthermore, accurate DEMs offer topographic data to be combined with field data and geophysical measurements [e.g. Bubeck et al., 2015; Mason et al., 2016, submitted].

Since software and computing resources become broadly available and affordable the option of data processing and analysis capabilities that are scalable, extensible, and innovative are provided to an increasing scientific community. Therefore, the question is not why to apply more elaborate techniques to visualise the seismic landscape but why not. Firstly, technological potentials are available and are continuously improved. This counts for data acquisition as well as for analysis and presentation. Secondly, additional benefit from innovative techniques is proven by numerous studies. Thirdly, assembling accurate information on earthquake geological effects in a geospatial context gathered by high quality innovative visualisations and reproducible measurements enables objective, transparent, and reliable evaluations for earthquake hazard assessment. Hence, advantages of
applying novel techniques and detailed workflows prevail to investments of application.

6.2 What are the implications arising from new knowledge about tidal notch formation and morphology?

In this section the advantages of the work presented in Chapter 4 and 5 are discussed. First, the arising knowledge of notch formation and sequence evolution is combined with the outcome of innovative notch identification. Second, tidal notch morphologies are included into the seismic landscape of Perachora Peninsula.

6.2.1 Tidal notches are now defined in space and time

As shown in Chapters 4 and 5 the formation of tidal notches is defined in space and time. Within the tidal range bioerosion is the driving force that develops symmetrical V-/U-shaped indentations predominantly on limestone coastlines [e.g. Pirazzoli, 1986; Evelpidou et al., 2012a; Antonioli et al., 2015]. Furthermore, in microtidal environments (tidal range <0.6 m) biological zonation is predominantly limited by the height of the waves and irregular variations of the sea due to winds and atmospheric pressure [Shennan et al., 2015]. Hence, the best preservation of tidal notches over geological timescales is given at steep and sheltered exposures [Pirazzoli, 1986] where continuous spray is absent or insignificant. Furthermore, cliff quarrying by wave action is negligible for homogenous and bulky carbonates.

When displaced from present-day sea-level tidal notch preservation is determined by the ratio of absolute difference in elevation to mean sea-level ($d_{abs}$) versus the local tide amplitude ($t_a$). A high ratio ($\gg 1$) implies that the displaced former strandline is no longer affected by bioerosion within the tidal range; if uplifted, even influences by temporary wave-action and spray is presumably no longer given [e.g. Agios Pavlos (S. Crete) $d_{abs}/t_a = 1m/0.175m \approx 5.7$; see Schneiderwind et al., 2017]. The preservation of a displaced notch at 50% requires a $d_{abs}/t_a$ ratio of 1 (see Fig. 4.4b). If a tidal notch is emerged by 50% of its vertical extent from mean sea-level, the high tide level corresponds to the point of maximum concavity. $d_{abs}/t_a$ ratios lower than 1 imply significant overwriting of pre-existing notch morphologies. As a consequence, differentiating between two notch generations is more difficult, if not impossible from apparent observations. However, applying high-performance morphological analysis on high-resolution 3D surface data enables detecting even minor morphological evidence of palaeo-sea-level marker and hence, provides newly identified notch generations to sea-level history.

For the Mediterranean Sea tidal ranges of 0.3-0.5 m are described, based on in-situ observations and satellite altimetric data [e.g. Evelpidou et al., 2012a; Antonioli et al., 2015; Schneiderwind et al., 2017]. Compiling a simple conceptual model from tidal range data (here: 0.4 m) and empirical scaling laws [e.g. Wells and Coppersmith, 1994] imply minimum magnitudes of $M \geq 6.5$ to preserve a tidal notch by 50% on an uplifting coastline. However, numerical modelling (see Chapter 5) showed that a tidal notch occurring different from present-day sea-level is not necessarily a result of coseismic coastal uplift/subsidence [see Figs. 5.7-5.11; Schneiderwind et al., submitted].
6. Discussion

The numerical model considers the dynamic sea-level change during the late Holocene for a particular region. For the first time both, rapid coseismic coastal uplift/subsidence and slow gradual relative sea-level changes are considered in tidal notch modelling and visualise the significant influence of modification on the projected cliff morphology. Furthermore, periods of notch formation are quantifiable. In particular, the beginning of Holocene notch formation of a certain region is dependent on the rate of relative sea-level change in accordance to isostatic and coseismic landmass movements, and dates back to ~6.5 kyr BP on average in the Mediterranean. Moreover, characteristics of a particular sea-level curve might shift the erosional base along a vertical cliff face and yield to notch widening, deepening, or partial modification of older features. Interestingly, Antonioli et al. [2015] concluded that local tide amplitudes are less than observed tidal notch heights, and the maximum convexity does not correspond to the maximum tidal ranges. However, almost all model predictions show recent notches developing asymmetrically due to gradual sea-level changes slightly falling below landmass motion rates. This is in particular true for moderately uplifting coastlines (Fig. 5.6 a, b, c). Hence, most recent notches are likely to have their roof slightly higher located than high tide stand although formed within the tidal range.

If accurate sea-level curves and values on landmass motion pose inputs to the model the projected cliff face represents a surface morphology that should have been formed if applied input parameters are correct. It should be noted that calculations made in chapter 5 also show very minor morphologies and remnants of overprinted notches (see Fig. 5.13). Furthermore, simulated cliff evolution is based on simplifying assumptions representing an idealised cliff profile and hence, provides only 2D information. However, notch profile heterogeneity on local scale has been described for numerous actual cliff faces [e.g. Kershaw and Guo, 2001]. In order to compare predicted notch profiles to reality high resolution TLS and high performance edge detection and feature extraction (Chapter 4) enable to visualise even minor evidence of tidal notch morphologies and further, allow evaluating occurrence in a 3D manner. If the predicted notch profile significantly matches with actual cliff morphologies coastal tectonic activity is described in 4D comprising an accurate sea-level history and well-balanced throw values of on- and offshore tectonic structures.

6.2.2 The case: Perachora Peninsula

Assessing the seismic landscape of Perachora Peninsula considering the application of tidal notches as a valuable archive of sea-level history requires detailed information on deformation characteristics, earthquake recurrence intervals, detailed mapping of notches along the coastline, and a suitable sea-level curve.

Deformation characteristics

Two major north-dipping faults of the south Alkyonides Fault Zone, namely the Pisia fault and the Schinos fault (see Fig. 2.3), ruptured during the 1981 Alkyonides earthquake sequence and produced mean surface displacements of 0.9 m (Pisia fault) and 0.3 m (Schinos fault) [Jackson et al., 1982; Taymaz et al., 1991; Leeder et al., 2005], respectively. Hubert et al. [1996] provided an elastic dislocation map for the events on February 24th
Figure 6.2: Geodetic constraints of Perachora Peninsula. Fault traces, slip orientations [Morewood and Roberts, 2002], scarp heights, and elastic deformation [Hubert et al., 1996] provide useful information for assessing the seismic landscape with respect to late Holocene coastal uplift/subsidence. Faults and coastal lithology after Bornovas et al. [1984b,a]. Cumulative scarp heights are derived by own mapping. Photographs collected at marked positions are shown in Fig. 6.3.
and 25th 1981 which are attributed to above mentioned faults. Stiros et al. [2007] provided the same geodetic constraints for the antithetic south-dipping Kaparelli fault that ruptured during the third event of the Alkyonides earthquake sequence on March 4th 1981. Both datasets consistently describe footwall uplift versus hanging wall subsidence ratios of 1/4 to 1/2 referring to 0.3-0.4 m subsidence at the Kaparelli fault [Stiros et al., 2007] and up to 1.0 m subsidence at the Pisia fault [Hubert et al., 1996]. Figure 6.2 shows the elastic deformation map modelled by Hubert et al. [1996]. The model also provides predictions of coastal uplift/subsidence. Interestingly, even coastal areas that clearly show late Holocene uplift (Fig. 6.3 g-j) are affected by predicted subsidence coverage.

**Earthquake recurrence**
The free-face height at the centre of the Pisia fault is $\sim 9$ m; along strike of the Schinos fault the escarpments appear smaller and have average heights of 6 m (Fig. 6.2). Generally, fault scarp mapping is suitable to give first order estimates on post-glacial slip rates [Benedetti et al., 2002; Papanikolaou et al., 2005; Reicherter et al., 2011]. However, the fault scarps across the Perachora Peninsula are located on steep slopes within forested environments. These circumstances prevent the application of TLS for continuous scarp height measurements [see Mason et al., 2016] and also make manual measurements difficult [see Papanikolaou et al., 2005]. Only from palaeoseismological trenching Holocene vertical slip rates for the Schinos fault of $1.2 - 2.4$ mm/yr are available. Furthermore, palaeoseismological trenching provides mean surface displacements of $\sim 0.6$ m per event [Pantosti et al., 1996; Collier et al., 1998], which is in good correlation with observed displacements during the 1981 earthquake sequence [e.g. Jackson et al., 1982; Taymaz et al., 1991], and earthquake recurrence intervals of $\sim 330$ years [Collier et al., 1998]. Assuming a seismic cycle comprising evenly distributed earthquake events with recurrence intervals of $330\pm 66$ years (20 %) and applying a footwall uplift to hanging wall subsidence ratio of 1:3, which is comparable to the 1981 events [Hubert et al., 1996; Stiros et al., 2007], to occurring surface displacement values of $0.6\pm 0.12$ m (20 %) a statistically generated earthquake history (Tab. 6.1) may pose input for notch profile modelling (Tab. 6.2) (see Appendix A for the "History Maker" Matlab code).
Figure 6.3: continued
Figure 6.3: continued
Notches along the coast of the Perachora Peninsula
Palaeo-strandlines along the southwestern coastline of Perachora Peninsula have been extensively investigated by Cooper et al. [2007]. The authors point out that notches are only preserved where limestone or lithified Quaternary conglomerates crop out (Figs. 6.2 and 6.3a-c). At Lake Vougliameni a composite volcano-sedimentary series comprising basic rocks crops out. Here, notch development obviously occurred. However, heavy degradation (Fig. 6.3d) led to significant modifications so that the quality of preserved notches is not suitable for accurate investigations [see Schneiderwind et al., 2017].

The northwestern coastline is mainly made of thick bedded lower Jurassic limestones (Fig. 6.2). The most-elevated strandline occurs by mean at ~3 m. Minor variations of coseismic coastal uplift along shore are not easy to quantify since the occurrence of abrasional notches (Fig. 6.3f, j) indicate that currents and strong wave action affect the shoreline.

At Strava and further to the East (Fig. 6.2) late Holocene coastal subsidence is denoted by Leeder et al. [2005]. Evidence for coastal subsidence pose submerged tidal notches (Fig. 6.3m-o). These palaeo-strandlines are again incised into steep limestone coastlines and appear submerged down to 0.7 m (Fig. 6.4a,b).

Close to Schinos a minor south-dipping fault juxtaposes schists and limestone in its footwall from quaternary alluvial deposits within the hanging wall. West to Schinos a small bay, crossed by this fault, is located in mean slip-direction of the Pisia fault and the Schinos fault and experienced major subsidence up to 1.2 m predicted by the dislocation model [Hubert et al., 1996]. Figure 6.3o shows the south-dipping fault plane comprising the recent strandline indicated by a dark greyish band of marine fauna. Sea-urchins settled directly underneath the shown water level are a sign of low tide stand, and furthermore, indicate that the cliff is sheltered from abrasive processes (Fig. 6.4). Therefore, this location appears suitable for tidal notch preservation without significant modification. Moreover, this site most likely experienced repeated hanging wall subsidence caused by

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**Figure 6.3:** Notches of Perachora Peninsula. Photo plate comprises abrasional notches (f, j, k), tidal notches (a-e, g-i, m-n) and other evidence of sea-level (d, l). p represents a cliff mainly made of schists east of Schinos showing no more palaeo-sea-level evidence. Massive limestone cliffs appear more suitable for notch preservation than coasts comprising flysch (e) or basic rocks (d).
Figure 6.4: Submerged notch close to Schinos. Natural appearance (a, b) and predicted notch shape for two different scenarios (c). The resulting profiles (d) demonstrate the difference between a scenario without offshore coastal uplift contribution (dashed orange line) and a scenario with contributions from offshore normal faults (red line). The scenario details are listed in Tab. 6.1 and Tab. 6.3.
Table 6.1: Statistically generated earthquake history of the Perachora Peninsula as input for tidal notch modelling without contributions from offshore coast-uplifting faults. Magnitudes are derived from empirical regressions after Wells and Coppersmith [1994].

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crustal movements along the Psia fault and Schinos fault.

Notch profile modelling at Schinos
Incorporating the sea-level curve for Peloponnese [Lambeck and Purcell, 2005], repeated coseismic subsidence of -0.35 – -0.55 m in recurrence intervals of 330±60 years [see Collier et al., 1998] (Tab. 6.1), and a net uplift rate of 0.7 mm/yr [Stewart and Vita-Finzi, 1996; Roberts et al., 2009], the model predicts a notch profile as shown in Figure 6.4d. Consistent in model prediction and natural expression no notch morphology occurs above present datum. Differences are expressed by the shape of the profile especially towards the bottom of the notch. The actual notch has a distinct roof expressed by a sharp edge ~0.45 m below mean sea-level. From the roof downwards incision evenly increases reaching its maximum incision at ~0.75 m on the actual cliff face. The lowermost section of the natural exposure is gradually shaped and shows no obvious edge representing the bottom of the notch. In contrast, the modelled profile for a scenario with no coseismic
6. Discussion

**Table 6.2:** Input parameters applied for tidal notch modelling at Schinos, Perachora Peninsula.

<table>
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Contributions from offshore faults projects a distinct roof at $\sim-0.15$ m and provides no distinct vertex for cliff sections just below present day sea-level. However, similar in both, the natural and predicted notch profiles, is a gradually carved shape down to $\sim-1.0$ m. Subsequently downwards, a distinct notch with its vertex at $\sim-1.3$ m is predicted resulting from $\sim325$ years of relative stagnation and significant coastal subsidence of $\sim0.52$ m 1717 years BP. This feature is not observed at the actual cliff face.

**Table 6.3:** Statistically generated earthquake history of the Perachora Peninsula including offshore coastal uplift contributions. Magnitudes are derived from empirical regressions after Wells and Coppersmith [1994].

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*Table continued on next page.*
It should be noted, that landmass net uplift and sea-level change are in unison only in the interval between ~3ka and 1.5ka BP. Prior to that phase, sea-level change outpaced the net uplift yielding in interseismic upward grazing of the cliff. During the balanced period only coastal coseismic subsidence shifts the erosional base and lifts it to a corresponding level of ~1.8 m below present-day sea-level. The latest phases since 1.5 ka BP are characterised by repeated coseismic coastal subsidence events and slow and gradual downward grazing of the cliff since the net landmass uplift outpaces sea-level change. However, gradual shifts after coseismic activity do not reach previous levels and thus, only partially overprint pre-existing morphologies.

The natural cliff face exhibits a distinct notch roof and a clear vertex incised by ~0.4 m. The maximum incision indicates either longer periods of relative standstill - which would be ~800 years applying an erosion rate of 0.5 mm/yr - or repeated erosive phases for the same parts of the cliff. However, a scenario incorporating only onshore coast downthrowing normal fault activity seems not be able to result in cliff profiles with significant similarities to the actual morphology.

If the offshore Strava fault and western Alkyonides fault (Fig. 6.2) contribute coastal uplift a different geodetic path (Tab. 6.3) results in a profile which shows more similarities to the modern cliff shape (Fig. 6.4). Adding hypothetical coseismic events offshore accompanied by geodetic constraints as they are assumed to be characteristic for the eastern Corinthian Gulf (~0.2 m uplift) and slip rates about 2.4 mm/yr [Sakellariou et al.,

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yield to profile predictions that almost match the shape of the actual exposure. A clear notch is predicted at -0.45 m with a distinct roof at ~0.25 m. Here, the maximum incision is about 0.2 m and therefore a bit less than for the natural feature. Additionally, a second notch is obvious just below with its vertex at ~0.7 m. Both indentations are differentiated by a distinct rim at ~0.6 m which poses notch bottom and notch roof, respectively. Contrasting to the model scenario without coseismic coastal uplift, here slow downward migration and coseismic displacements vertically oriented both ways yield in significant overprinting of pre-existing morphologies. The lower notch is at least once modified between 1.5 ka and 0.7 ka BP. The upper notch is the result of at least two erosive phases. Interestingly, very short periods between coastal subsidence and coastal uplift do not provide enough time to significantly incise the cliff, whereas the periods between uplift and subsidence are longer and therefore pose erosive phases. These conditions are the reason for the distinct rim separating both notches. Indeed, these conditions are the result of model assumptions and do not necessarily represent natural recurrence intervals for all contributing faults. However, they demonstrate how slightly different conditions on individual faults result in distinct notch levels that again exhibit multiple notch formation phases. If the applied assumptions are correct, the model proves the coseismic activity of offshore faults that uplift coastal regions and furthermore sustains obtained fault activity estimates from other coseismic contributors within the seismic landscape.

**What’s next?**

Tidal notches different from present-day sea-level are not exclusively raised or subsided by coseismic activity and thus, are no earthquake geological effect in general. As shown in Chapter 5, Holocene tidal notches potentially form as a function of the local sea-level curve and coseismic activity, both on- and offshore, in different periods at varying erosional levels, in correspondence to a modern cliff face, since ~ 6.5 kyr. The particular outcomes and conclusions for palaeoseismological studies are:

- Displaced tidal notches, different from present-day sea-level are not necessarily the result of coseismic deformation
- The tidal notch development and preservation can be highly variable even at local scale
- Reliable data on relative sea-level change and coastal net movements are of major importance regarding the formation and preservation of tidal notches
- A modern notch sequence from the most elevated notch gradually descending to present datum is not inevitably of decreasing age
- Cliff morphology is continuously modified, resulting in widening, in deepening, in separation of tidal notches, and quite often in overprinting of pre-existing notch morphologies
- The vertical position of a tidal notch on a modern cliff face is the result of biasing slow relative sea-level changes and rapid coseismic contributions from both on- and offshore seismogenic origins during the late Holocene
This is also true for the coast of Perachora Peninsula where palaeostrandlines appear raised from present-day sea-level as well as submerged. The earthquakes of February 24 and 25, 1981, subsided the coastline at Schinos and displaced a distinctive tidal notch ~ 0.7 m below sea-level. Modelling the notch showed how tectonic circumstances and seismicity of on- and offshore normal faults could have happened and estimate a late Holocene slip rate of ~ 1 mm/yr at the Strava fault or western Alkyonides fault. Discrete identification of individual palaeoshoreline levels in combination with tidal notch profile modelling along shore of Perachora Peninsula will complete datasets on the seismic activity of the SAFZ and extend knowledge on offshore tectonic activity.
One of the basic rules of the universe is that nothing is perfect. Perfection simply doesn’t exist. Without imperfection, neither you nor I would exist.

S. Hawking

---

No great discovery was ever made without a bold guess.

I. Newton (1643-1727)
The seismic landscape of an earthquake-prone area comprises geological constituents that vary in preserved quality and individual impact on seismic hazard assessment. This work mostly deals with earthquake geological effects in regions of tectonic extension. However, the development of new technologies and their continuous improvements during the past 60 years enabled to design techniques and combining workflows that aim to visualise and quantify palaeoseismological archives, each with regards to the prevailing tectonic stress regime. Typically, earthquake magnitudes in extensional tectonic settings are in the order of one magnitude smaller than in compressional frames. As a result, also earthquake geological effects appear in a smaller scale (e.g. shoreline displacement). Therefore, the presented multidisciplinary visualising methodologies target the identification and verification of normal faulting earthquake geological effects with high-resolution and high-accuracy techniques applied in innovative and effective workflows.

In fact, two pioneer applications for the use of terrestrial laser scanning (TLS) into the fields of palaeoseismology, coastal and tectonic geomorphology as well as an innovative tidal notch modelling approach are presented. Moreover, a multidisciplinary workflow comprising traditional palaeoseismological trench logging, geophysical measurements of the shallow subsurface, applications of remote sensing and geographic information systems (GIS), statistical evaluations as well as algorithm testing and development, is integrated. By taking advantage of the high spatial resolution and the near-infrared backscatter signal of TLS its application has been successfully expanded to palaeoseismological trenching studies and to palaeoshoreline identification. TLS has been merged with ground-penetrating radar (GPR) investigations for 3-dimensional imaging of normal fault hanging wall sedimentary architectures. The presented multiparametric approach offers a 3D view of the trench wall stratigraphy and provides a higher objectivity in grouping different features within the trench by applying multispectral analysis. A second TLS application for the identification of palaeoshoreline tidal notches is introduced using high spatial resolution point cloud data for high performance curvature analysis on modern cliff faces. Based on such data, a newly developed custom Fuzzy Logic edge detection algo-
rithm enables to trace previously undiscovered tidal notches or features corresponding to multiple notches. This is in particular important in extensional tectonic settings, where the uplift caused by normal faults is in the order of a few decimetres and overwriting of pre-existing older features is common. Moreover, numerical modelling has been used to visualise tidal notch formation, evolution, and deformation during the late Holocene and provides new insights to long standing debates regarding the relationships between palaeoshoreline tidal notches and active tectonics.

The herein used instruments TLS and GPR have become standard field techniques and help to assess geomorphological and lithological expressions in 3D. However, dating different types of earthquake geological effects is not always possible so that palaeoseismologists have to rely on empirical scaling laws in order to infer historic earthquake data. The fusion of high quality data and innovative workflows does not only improve objectivity in assessment and makes palaeoseismological interpretations robust but also enables to put all gathered information into a geospatial framework that is crucial for reviewing a seismic landscape and hence, allowing justified statements on the seismic activity of an area. Only if individual components of the seismic landscape are adequately accessed, addressed, analysed, and correlated to each other, conclusions regarding seismic hazard assessment are well-founded.

The most important junction in this overall system is the applied scaling law. Future studies may consider climatic forces that modify the seismic landscape and thus enhance empirical regressions. Furthermore, high-performance analytic methods applied to continuously collected high-resolution and high-quality data has the potential to gather detailed information of deformation processes and poses significant input for fine-tuning existing scaling laws and generates new links between earthquake geological effects and palaeoseismological consequences. Interdisciplinary studies combining conventional palaeoseismological methodologies with geophysical techniques, applications of remote sensing, and image analysis implementations of computer vision, have the capabilities to collect the essential datasets, to perform top-level analyses, and to produce the required information in order to understand seismic activity far back into history and hence, provide valuable support for the seismic hazard assessment of particular regions.
References


References


References


References


References


References


References


References


References


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References


References


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APPENDIX A

Matlab Code: EQ history maker

This EQ history maker calculates a fictive earthquake history based on average displacements and recurrence intervals for a given normal fault! In general, lower and upper value boundaries are defined by 20% errors for displacement values $d$ and individual recurrence intervals $Rec$.

Required Inputs:

- Number of Events "$NOE$"
- Average Displacement "$ad$"
- Average Recurrence Interval "$AveRecInt$"
- Uplift:Subsidence Ratio "$U":"S$"

Output:

- Fault rupture history matrix $Ft_hist$ size(length(NOE) x 6).
  Matrix header: [Years_[BP] Interval Magnitude Displacement Subsidence Uplift]
- Mean Magnitude
- Average Displacement

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Contents

- Inputs
- Calculations
- Results
A. Matlab Code: EQ history maker

Inputs

Number of Events: \( NOE \)
Please change to your needs

\[ NOE = 40; \]

Average displacement: \( ad \)
Please change to your needs

\[ ad = 0.6; \]

Average recurrence interval: \( AveRecInt \)
Please change to your needs

\[ AveRecInt = 330; \]

Uplift:Subsidence \((U:S)\) Ratio
Please change to your needs

\[ U = 1; \quad \% \quad \text{Uplift} \quad U \]
\[ S = 3; \quad \% \quad \text{Subsidence} \quad S \]

Calculations

\% multiplier
\[ mR = U + S; \]

\% displacement per fictive event
\[ uplim_d = ad \times 1.2; \]
\[ lowlim_d = ad \times 0.8; \]
\[ d = \text{zeros}(NOE,1); \]
\[ d = (uplim_d - lowlim_d) \times \text{rand}(NOE,1) + lowlim_d; \]

\% subsidence at up:down = 1:3
\[ \text{subs} = (d \times mR \times S) \times -1; \]

\% uplift at up:down = 1:3
\[ \text{up} = (d \times mR \times U); \]

\% recurrence history
\[ uplim_{Rec} = AveRecInt \times 1.2; \]
\[ lowlim_{Rec} = AveRecInt \times 0.8; \]

\[ \text{Rec} = \text{zeros}(NOE,1); \]
\[ \text{Rec} = (uplim_{Rec} - lowlim_{Rec}) \times \text{rand}(NOE,1) + lowlim_{Rec}; \]
% Magnitude estimates after Wells & Coppersmith (1994) using d as average
% displacement parameter ("AD")
M = zeros(NOE,1);
M = 6.78 + 0.65 * log(d); % Wells & Coppersmith [1994]

Results

Fault rupture history matrix [Year_[BP] Interval Magnitude Displacement Subsidence Uplift]

Ft_hist = [round(cumsum(Rec)), Rec, M, d, subs, up]

% Mean Magnitude
Mean_Magnitude = mean(M)

% Average Displacement
Average_Displacement = mean(d)

Ft_hist =

1.0e+04 *

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<td></td>
<td>0.7330</td>
<td>0.0354</td>
<td>0.0007</td>
<td>0.0001</td>
<td>-0.0001</td>
<td>0.0000</td>
</tr>
</tbody>
</table>
A. Matlab Code: EQ history maker

```
0.7599 0.0270 0.0007 0.0001 -0.0001 0.0000
0.7873 0.0273 0.0007 0.0001 -0.0001 0.0000
0.8206 0.0333 0.0006 0.0001 -0.0000 0.0000
0.8482 0.0277 0.0006 0.0001 -0.0000 0.0000
0.8854 0.0372 0.0006 0.0001 -0.0000 0.0000
0.9226 0.0372 0.0006 0.0001 -0.0000 0.0000
0.9586 0.0359 0.0006 0.0001 -0.0000 0.0000
0.9870 0.0284 0.0007 0.0001 -0.0000 0.0000
1.0221 0.0351 0.0006 0.0001 -0.0000 0.0000
1.0553 0.0332 0.0006 0.0001 -0.0000 0.0000
1.0945 0.0392 0.0006 0.0001 -0.0000 0.0000
1.1295 0.0350 0.0007 0.0001 -0.0001 0.0000
1.1665 0.0370 0.0007 0.0001 -0.0000 0.0000
1.1989 0.0324 0.0007 0.0001 -0.0001 0.0000
1.2310 0.0321 0.0007 0.0001 -0.0001 0.0000
1.2683 0.0373 0.0006 0.0001 -0.0000 0.0000
1.2958 0.0275 0.0006 0.0001 -0.0000 0.0000
1.3239 0.0282 0.0006 0.0001 -0.0000 0.0000

Mean_Magnitude =

6.4505

Average_Displacement =

0.6063
```