ACTIVE FAULT INVESTIGATIONS IN THE WESTERN PELOPONNESE AND EASTERN CRETE, GREECE

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ABSTRACT

Greece is the most seismically active country in Europe and is subject to regular earthquakes. The majority of earthquakes in Greece are caused directly by the subduction zone which runs from close to the Ionian Islands in the west, past southern Crete in the south, to the south-western coast of Turkey. However, the region also hosts many normal faults which are around 10 – 25 km long and therefore smaller in length than the subducting thrusts, but still capable of destructive earthquakes. Research on many of these normal faults is yet to be carried out and their finite lengths and slip rates are largely unknown. Neotectonic, palaeoseismological and palaeoenvironmental research on several normal faults in two study areas are described in this thesis.

Study area 1 is located in the northern Kyparissiakos Gulf, western Peloponnese. Fault mapping, fault slip data, observable throw calculations and published offshore bathymetry data are used to accurately determine fault lengths for two bedrock faults: the northern and southern Lapithas faults. Empirical calculations are then used to determine maximum potential magnitudes for these faults. The northern Lapithas fault comprises two segments; a ca. 18 km long onshore segment and a ca. 15 km long offshore segment. When viewed separately, potential maximum earthquake magnitudes (Mw) of 6.5 and 6.4 could occur on the onshore and offshore segments respectively. As a worst case scenario, these two segments could both rupture during an earthquake producing a maximum magnitude (Mw) of 6.9. The total length of the southern Lapithas fault is ca. 5 km and could therefore produce a maximum potential magnitude (Mw) of 5.8. Earthquake archaeological effects are also described at Samicum and Ancient Olympia which are two nearby archaeological sites, the latter being a UNESCO world heritage site which has been largely reported to have been damaged by earthquakes in AD 522 and/or AD 551. However, a review of the historical literature and earthquake records shows that no earthquakes in these years can explain the damage.

Study area 2 is the Lastros-Sfaka Graben located within the Ierapetra fault zone in eastern Crete. Fault mapping and t-LiDAR at the Lastros fault are used to determine accurate fault lengths and postglacial scarp heights. A representative slip rate is presented from scarp height calculations in a reliable area not influenced by erosion and sedimentation. This reliable area is only 100 m out of 1.3 km of scanned fault scarp. The mean scarp height in this reliable area is 9.4 m which equates to a slip rate of 0.69 ± 0.15 mm/a when using 15 ± 3 ka for initial scarp exhumation. The natural variation in scarp height in this reliable area is ±12 % from the mean. Empirical calculations indicate that the Lastros fault has a maximum
This magnitude earthquake can cause a maximum vertical offset of 0.31 m and therefore a recurrence interval of 499 ± 100 years is obtained. Cemented colluvium is attached to the fault planes of many Cretan faults. Stable isotope analysis and descriptions of cemented colluvium at the Lastros fault show that the cement precipitated from springs comprising meteoric water located at the fault plane. This occurred after initial colluvial deposition through rockfalls and debris flows. Palaeotemperature calculations of the water from which the calcite precipitated indicate seasonal cement formation occurred soon after a glacial maximum in the Late Pleistocene. GPR on the hanging-wall indicates that cemented colluvium is also present in the subsurface, below uncemented colluvium. This indicates that the source springs were not just located where there is cemented colluvium at the present day surface, but they were once more widely distributed along the strike of the fault.

Trenching investigations at the antithetic Sfaka fault identify different generations of fissure fills. Retrodeformation analyses and ¹⁴C dating of the fill material indicate at least four events dating back to ~16 ka, with the last event occurring ~6 ka. These dates either represent activity on the Lastros fault, assuming they formed coseismically, or accommodation events. Cross sections of the Lastros-Sfaka Graben show that the finite throw is limited to around 300 m. Therefore, the 0.69 ± 0.15 mm/a slip rate for the Lastros fault indicates that both faults are relatively young having initiated 435 ± 120 ka.

Lastly, the results of a trench logging exercise at the Kaparelli fault, Gulf or Corinth, are presented. This fault ruptured during the 1981 Corinth Gulf earthquake series with a magnitude (Mₚ) of 6.4. This trenching study was undertaken for the development of a new technique to visualise trench stratigraphy in 3D using t-LiDAR and GPR.
KURZFASSUNG


Das zweite Arbeitsgebiet ist der Lastros-Sfaka Graben innerhalb der Ierapetra Störungszone im östlichen Teil der Insel Kreta. Störungskartierung und terrestrische LiDAR-Vermessungen der Lastrosstörung wurden genutzt um akkurate Störungslängen und postglaziale Geländestufenhöhen zu ermitteln. Eine repräsentative Bewegungsrate wird für solche Bereiche ausgewiesen, in welchen die Geländestufenhöhe weder durch Erosion noch durch Sedimentation modifiziert wurde. Diese Bedingungen sind in einem Bereich von lediglich 100 m entlang der 1,3 km langen gescannten Geländestufe gegeben. Hier beträgt die mittlere Geländestufenhöhe 9,4 m, welche mit einer postglazialen ($15 \pm 3$ ka) Bewegungsrate von
0,69 ± 15 mm/a korreliert. Dabei beträgt die Varianz der Geländestufenhöhe ±12 %.

Empirisch betrachtet ist die Lastrosstörung dazu in der Lage, ein Erdbeben der Magnitude (M_s) 6,42 zu generieren und damit einen vertikalen Versatz von 0,31 m zu produzieren. Daraus ergibt sich eine Erdbebenwiederholungsrate von 499 ± 100 Jahre.


Schurfuntersuchungen an der antithetischen Sfakastörung lassen verschiedene Generationen von Oberflächenfissuren identifizieren. Eine Retrodeformationsanalyse zeigt, dass mindestens vier Erdbebenereignisse zu dieser Oberflächendeformation führten. Dies wird darüber hinaus durch ¹⁴C- Datierungen bestätigt, welche den Deformationszeitraum auf ~16-6 ka eingrenzen. Entweder spiegeln diese Deformationen tektonische Aktivitäten an der Lastrosstörung oder Ausgleichsaktivitäten wieder. Den Lastros-Sfaka-Graben kreuzende geologische Profile deuten darauf hin, dass der finite Versatz auf etwa 300 m limitiert ist. Mit der Bewegungsrate von 0,69 ± 0,15 mm/a an der Lastrosstörung geht somit einher, dass beide Störungen relativ jung sind und Rupturereignisse dort etwa alle 435 ± 120 Jahre stattfinden.

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Foreword

Rationale

Throughout history earthquakes have caused extensive damage and injury to humans - no more so than in Greece where human civilisation dates back to the early Minoan times. It is only quite recently that our understanding of earthquakes has developed enough for us to study these events and the establishment of plate tectonics theory in the 20th century was the catalyst for this. The development of instrumental seismometers in the early 20th century allowed plate boundaries, where the majority of earthquakes occur, to be accurately mapped. However, we now know that earthquakes do not only occur at plate boundaries, and earthquakes on some faults can have very long recurrence periods. Neotectonic studies allow us to characterise regions that have undergone geologically recent deformation and the physics and geometry of individual fault structures can be described. Palaeoseismology offers techniques to estimate the properties and timing of surface rupturing earthquakes prior to the instrumental monitoring and historical periods. Dates of palaeoearthquakes and calculations of slip rates allow spatial and temporal earthquake patterns to be identified. The hazards associated with regions or individual fault structures can then be communicated to society.

Greece is subject to regular earthquakes and is the most seismically active country in Europe; approximately 40% of Europe’s seismic energy is released in Greece. The majority of earthquakes in Greece are caused directly by the subduction zone which runs from close to the Ionian Islands in the west, past southern Crete in the south, to the south-western coast of Turkey; the North Anatolian Fault in the northern Aegean also significantly contributes to the seismicity pattern of the area and defines the geodynamics. However, the region also hosts many normal faults which are relatively small in length but still capable of destructive earthquakes. Research on many of these normal faults is yet to be carried out and their finite lengths and slip rates are largely unknown. Research on several of these normal faults from the Peloponnese and the island of Crete is described in this thesis.

It is clear from the landscapes of Greece that seismic activity has sculpted the environment. There are many impressive mountains with large faults scarps making the region the ideal place for neotectonic research. The long history of human habitation including many damaged archaeological sites also aids in seismic activity descriptions. Some faults scarps also have cemented colluvium attached to them. Isotope studies on this phenomenon allow us to describe the palaeoenvironment and determine how faults have evolved over time.
Scope

The objectives of this research are to contribute to neotectonic and paleoseismic research by undertaking multi-disciplinary investigations on normal faults where very little or no research has been done to date. This will allow some fundamental questions to be answered which have an impact on the local seismic hazard.

The principle questions we address regarding faults in the western Peloponnese are:

- Are the existing geological data accurate to define fault lengths and potential earthquake magnitudes on individual faults?
- Can archaeological damage be used to help define individual earthquakes?
- To what extent can written reports of archaeological damage be used to identify earthquake events?

The principle questions we address regarding faults in the eastern Crete are:

- Can accurate postglacial slip rates be determined?
- When did the last large earthquakes occur?
- Can we estimate the ages of the faults?
- How do the faults fit within the regional geodynamic context?
- Does hanging-wall cementation provide good evidence to reconstruct the palaeoenvironmental history and fault evolution?

To address these questions classical field mapping and fault slip data were used in the western Peloponnese on two faults, the northern and southern Lapithas faults. Damage to local archaeological sites is described, and documented damage reports were critically studied to fully describe the study area. On Crete a multi-disciplinary investigation comprising terrestrial remote sensing, trenching, shallow geophysics and laboratory analyses were used to describe the Lastros-Sfaka Graben, part of the Ierapetra Fault Zone, in the east of the island. Lastly, the results of re-logging exercise at an open palaeoseismic trench at the Kaparelli fault in the Gulf of Corinth are presented. This re-logging was undertaken to enable the development of a method to visualise trench stratigraphy in 3D.
Thesis structure

Chapter 1 introduces the tectonics of the eastern Mediterranean and describes the evidence for large recent earthquakes. Chapter 2 provides a summary of the methods used in the thesis. At the end of Chapter 2 the rationale for choosing the study sites is described. Chapters 3-5 comprise modified academic papers produced by this research. Chapter 6 is a short chapter summarising the results of a trenching exercise undertaken on the Kaparelli fault in the Gulf of Corinth. Chapter 7 is a discussion and conclusions chapter drawing together the interesting outcomes from the study areas. Chapters 8 and 9 are respectively references and acknowledgements, and these are followed by the Appendix.
Chapter 1: Introduction

This chapter provides the reader with the geological background of the Aegean and the current theories for the development of normal faulting throughout the region. Many of these normal faults comprise bedrock scarps juxtaposed against colluvial and/or marine sediments and fault scarp formation and their preservation potential are introduced. Instrumental seismicity and the seismic risk of the broader region are presented including a summary of rupture processes and the parameters involved with earthquakes. Lastly, the geology within the two study areas is summarised. There are numerous publications which deal with the nature of tectonics in the eastern Mediterranean and for further information Allen et al. (2004), Jolivet et al. (2013) and van Hinsebergen & Schmidt (2012) should be consulted.

1.1 Seismotectonic overview of the eastern Mediterranean

The geology of the Aegean comprises a complex assemblage of terrains, island arcs, and ophiolites accreted during the closure of the Palaeo-Tethys and Neo-Tethys Oceans. This section provides an overview of the processes involved with this oceanic closure, the tectonic units which were accreted to form the current geology of the Aegean, and the current extensional regime that has produced the normal faults, which are the focus of this thesis.

1.1.1 Oceanic closure and plate collision

The tectonic history of the eastern Mediterranean is complex due to the huge number of internal forces that have affected the broader region, and there are various interpretations regarding individual phases of extension and compression. Here, a brief summary of the main events which led to the formation of the Aegean from Permian/Triassic boundary (250 Ma) to the Miocene (13 Ma) is provided (Fig. 1.1), based on the interpretations of Berra & Angioli (2014) and Blakley (2015) and references therein.

At the beginning of the Mesozoic (around 250 Ma,) the Earth’s landmass was dominated by the supercontinent Pangea. What is now the eastern Mediterranean was covered by the Palaeo-Tethys and Neo-Tethys Oceans, which separated the eastern extent of Laurasia and Gondwana (Fig. 1.1). At this time the Palaeo-Tethys Ocean was being subducted below the southern margin of Laurasia, and rifting was occurring at the northern margin of Gondwana producing a series of microplates (Angioli, 2014; Blakley, 2015). In the Early Jurassic (200 Ma) the Palaeo-Tethys Ocean closed, evidenced by the Cimmerian Orogeny (Fig. 1.1). The break-up of Pangea also began with intense extensional and strike-slip tectonics developing
on the southern margin of Europe and connected westwards to the central Atlantic, where rift basins formed leading to the detachment of Africa from North America (Blakley, 2015).

In the Early Cretaceous (125 Ma) the North Atlantic Ocean was opening to the west of Iberia, and rifting was occurring in the northernmost part of the future Atlantic Ocean between
Canada and Scandinavia (Fig. 1.1). The southern margin of Laurasia was still characterised by subduction of the Neo-Tethys Ocean beneath it. The central part of the Neo-Tethys was characterised by the northward pull of a complex puzzle of microplate blocks, including future parts of Turkey, Greece and Adria (Berra & Angiolini, 2014). At the end of the Cretaceous (75 Ma) the present-day continents were almost completely defined, with rifting only taking place in the northern Atlantic to the west of Scandinavia (Fig. 1.1). Continuing subduction of the Neo-Tethys Ocean caused the complex puzzle of microplate blocks, which will form Turkey, Greece and Adria, to approach the southern margin of Eurasia, and the subsequent collision will produce the Alpine and Turkish orogenic belts (Blakley, 2015). Minor oceanic basins formed between the Alpine and Neo-Tethys Oceans (e.g. Vardar and Pindos Oceans; Stampfli & Borel, 2002; Papanikolaou, 2013) indicating the presence of multiple verging subduction zones (see Section 1.1.2). The complex puzzle of microplates and sediments of the oceanic basins were then sandwiched within the collision zone of Africa and Eurasia through the Eocene (60 – 34 Ma) due to Africa moving towards the southern boundary of Eurasia. This gradually closed the Neo-Tethys Ocean and formed the Alpine orogenic belt (Blakley, 2015).

In the Miocene the Dead Sea Fault formed, separating the Arabian plate from the African Plate. Around 13 Ma the Arabian plate began to collide with Eurasia creating a large deformation zone and closing the Neo-Tethys Ocean (Fig. 1.1), with its remnants being named the Mediterranean Sea and Arabian Gulf. Extension to the east within the Aegean began around this time, and along with Arabia plates anticlockwise rotation whilst moving north (Bosworth et al. 2005), formed the North Anatolian fault. The processes involved in Aegean extension are covered in Section 1.1.3.

### 1.1.2 Tectonostratigraphic terranes of Greece

The Hellenides forms part of the Alpine orogenic belt and dominates the structure of Greece. The region has long been recognised to comprise numerous distinct sedimentary facies belts known as “tectonostratigraphic terranes" which have a NNW-SSE orientation (Aubouin, 1959; Papanikolaou, 1997) (Fig. 1.2). These linear terranes are thrust sheets that developed as a result of oceanic closure. Two types of terrane are recognised based on related geological units: continental terranes comprising pre-Alpine crustal basement rocks and sedimentary cover of shallow-water carbonate platforms; and oceanic terranes comprising basin sediments overlying ophiolites. Five continental terranes (H1, H3, H5, H7 and H9) and four oceanic terranes (H2, H4, H6, H8) are recognised (Fig. 1.2) (Papanikolaou, 1997; 2013).
There are three main phases which describe the tectonic evolution of each continental tectonostratigraphic terrane: 1) The rifting phase, which creates a new diverging plate boundary between a new terrane and the Gondwanan (African) plate to the south (see Fig. 1.1). 2) The drifting phase, representing the northward motion of the continental terrane within the Tethys Ocean, due to the closure of pre-existing basins further north and the opening of basins to the south. 3) The final phase is the accretion and amalgamation of the terrane to the European continent. Final accretion of the continental terranes occurred between the Eocene-Oligocene (H7) and Mid-Late Miocene (H1) (Papanikolaou, 2013).
Chapter 1: Introduction

There are also three main phases that describe the tectonic evolution of each oceanic tectonostratigraphic terrane: 1) The rifting phase, where the collapsed blocks within the rift gradually form the two continental margins, one towards Africa in the south and one towards Europe in the north. 2) The oceanic opening phase, where ophiolites form and abyssal/pelagic sediments are deposited. 3) The subduction and oceanic closure phase, where the oceanic basin is sutured and the ophiolite obducted. This final phase finished between the Eocene (H8) and Late Eocene (H2) (Papanikolaou, 2013).

There is a lot of information on these tectonostratigraphic terranes and for detailed information Papanikolaou (1997; 2013) or van Hinsbergen & Schmidt (2012) should be consulted. This thesis focusses on two study areas, one in the western Peloponnese (study area 1) and one in eastern Crete (study area 2). Both of these study areas comprise units of H1 (External Carbonate Platform) and H2 (Pindon/Cyclades oceanic basin) (Fig. 1.2). A brief overview of the geology of both study areas is provided in sections 1.3 and 1.4.

1.1.3 Aegean extension

The present day tectonic configuration of the eastern Mediterranean is shown in figure 1.3. The African plate in the south is separated from the Arabian plate in the east by the Dead Sea fault, and the Anatolian plate is separated from the Eurasian plate in the north by the North Anatolian fault (NAF). The Aegean region (approx. area shaded in pink in Fig. 1.3) is located at the western end of the NAF and covers western Turkey, the Aegean Sea and the majority of Greece. During the closure of the Neo-Tethys Ocean, oceanic lithosphere joined to the African continent was subducted at the Hellenic arc. The Hellenic arc then moved southwards due to the onset of extension within the Aegean. This subduction continues today at the Hellenic arc where it forms a trench and back-arc system (Fig. 1.3). There is debate regarding exactly when and why extension in the Aegean began. There are three different processes that are suggested to describe the evolution of the Aegean region: 1) slab retreat / subduction roll-back, 2) gravitational collapse of the overthickened (post orogenic) lithosphere, and 3) lateral extrusion (escape tectonics) associated with the collision of Arabia into Eurasia. These three processes are described below.
Subduction roll-back is the most commonly stated theory for extension in the Aegean region and is evidenced by the southward younging magmatism and metamorphic exhumations, and the high dip angle of the subducting slab. A wide range of ages from 5-30 Ma have been suggested for the onset of this process (Meulenkamp et al. 1988; LePichon and Angelier 1979; Jolivet et al. 2013); however, around 25 – 30 Ma is likely a more realistic figure (van Hinsebergen & Schmidt, 2012; Jolivet et al. 2013). van Hinsebergen & Schmidt (2012) state that extension occurred in two stages: the first from 25-15 Ma where around 110 km of extension took place, and the second from 15 Ma to present where a maximum of 290 km extension occurred.

Gautier et al. (1999) argue that gravitational collapse started ~21-38 Ma and was the result of thickened continental lithosphere, formed during the Alpine collision, collapsing southward into the Hellenic arc. Through the dating of metamorphic core complexes and associated units, the authors suggest that extension was occurring by 21 Ma. Hatzfeld et al. (1997) carried out gravity collapse analogue model experiments to represent Aegean extension. The
models consisted of a tank, half of which contains a two-layer brittle-ductile lithosphere (sand-silicone) floating on top of a heavier fluid (honey) representing the asthenosphere. This half of the tank was separated from the other half containing just honey by a vertical wall with central gate. When the gate is removed the model flows under gravity though the space. This model produced features which broadly represent those observed in the Aegean.

Mantovani et al. (2006) state that the driving force for Aegean extension is not subduction roll-back or gravitational collapse but the collision of the Arabian plate into Eurasia. This collision transfers strain westwards and the lateral extrusion of Anatolia via the North Anatolian Fault (Fig. 1.3). Approximate west-east shortening is accommodated by the southward buckling of the Aegean region (Mantovani et al., 2006).

Reilinger et al. (2006, 2010) present geodetic data of plate motions and fault slip rates within the Aegean. The authors determined a GPS velocity of 33 mm/yr for the southwest movement of the Aegean microplate (Fig. 1.3). However they also measure little internal deformation across the Peloponnese and southern Aegean Sea, which contradicts with the geological evidence for large scale post-Pliocene extension. Reilinger et al. (2010) states that this is because the normal faults observed in the Peloponnese and southern Aegean Sea pre-date movements on the NAF. An earlier phase of African slab roll-back caused the extension, and then later (< 4 Ma) the NAF developed propagating westwards and cutting off the central Aegean and Peloponnese from the Eurasian plate. There is, however, debate as to whether GPS velocities represent long-term slip rates. Reilinger et al. (2010) state that they believe the uncertainty is ±10%; however, the authors do not account for post-seismic rebound which can severely affect deformation rates (e.g. Mantovani et al., 2006).

It is most likely that extension throughout the Aegean is caused by a combination of subduction roll-back, gravitational collapse of the lithosphere and the extrusion of Anatolia. This combination has been suggested by a number of authors (Jolivet et al. 2013; Philippon et al., 2013; van Hinsebergen & Schmidt, 2012). However, the dominant process and timing of when extension began remains disputed.

1.1.4 Neotectonic domains

The Aegean region can be broadly split into three domains with different neotectonic trends (Mariolakos & Papanikolaou, 1981, 1987) (Fig. 1.4). The first domain (Domain I) has an E-W neotectonic trend. Here, north-south extension of the upper crust is occurring similar to that observed in the Corinth rift and central Greece (Mariolakos & Papanikolaou, 1981; Armijo et
al., 1996; Roberts, 1996; Papanikolaou et al., 2007). In southern Peloponnese and western Crete (Domain II), roughly NW-SE normal faulting is causing northeast-southwest extension; and in eastern Crete and the eastern Aegean islands, SW-NE faulting prevails (Domain III).

For Domain I, Papanikolaou & Royden (2007) state that E-W normal to oblique-slip normal faults can be formed by the transtensional stressfield produced by the development of the Central Hellenic Shear Zone which has varying convergence rates. Domains II and III have both undergone mainly trench perpendicular (NNE-SSW) extension, which can be associated with subduction roll-back and/or gravitational collapse. These two domains have both also undergone significant trench parallel (E-W) extension. van Hinsebergen & Schmidt (2012) carried out a kinematic reconstruction of the Aegean and state that the original distance 25 Ma between SW Peloponnese and Rhodes was 100 km and now it is 750 km. The authors attribute this to opposing rotation directions of Domains II and III. This rotation or trench parallel extension is not observed in recent geodetic measurements (Reilinger et al. 2006, 2010). However, Shaw and Jackson (2010) present focal mechanism data from a number of large (Mw > 5.2) instrumental earthquakes within Domains II and III which show E-W extension (see Section 1.1.5). There is therefore no definitive explanation for these domains and due to the orientation of the Hellenic arc it is unclear how these domains interact. However, based on fault throws and tectonic geomorphology (Papanikolaou and
Lozios, 1990), fault slip rates (e.g. Grützner et al. 2015), and instrumental and historical seismicity, it is likely that Domain II is less active than Domain I.

1.1.5 Instrumental seismicity and seismic risk in the Aegean

Since the early 1900s earthquakes have been recorded throughout the Aegean by seismometers. Before their advent only historical reports could be used to locate events, and qualitative descriptions of damage were used to estimate magnitudes. For a thorough description of historical events in the eastern Mediterranean up to 1900 Ambrasays (2009) should be consulted. Figure 1.5 shows instrumentally recorded earthquake’s in the Aegean between 1900 and 2008 (see Makropolous et al. 2012 and references therein for original source data). This earthquake data has been filtered for magnitude and then scaled for magnitude and depth; earthquakes of Mw 5.5 and above are presented in figure 1.5. This is because Mw 5.5 is the lowest magnitude required to cause deformation at the ground surface (McCalpin, 2009). It can be seen that seismicity is well distributed throughout the Aegean with only parts of the central Aegean Sea not having experienced large earthquakes. There is a clear clustering of earthquakes along the Hellenic trench arcing from the western Peloponnese to southwest Turkey. There is a range of magnitudes and depths but most are within 40 km of the surface. The large deep events (>60 km; orange in Fig. 1.5) have epicentres within the Aegean Sea set back from the Hellenic trench. This is due to the African slab being subducted at shallow angle forming the so-called Benioff zone. There is a large concentration of earthquakes to the west of the Peloponnese near the Kefalonia transform fault. Here, the majority of the earthquakes are of shallow depth (<20 km) and a number of earthquakes have magnitudes of Mw 6.5 and above; e.g. the 1953 Ionian earthquake struck the southern Ionian islands with a magnitude Ms of 7.2 (Papazachos, 1997) and caused over 600 deaths; recently in November 2015 a Mw 6.4 strike-slip earthquake occurred in Lefkada causing two fatalities. The largest shallow events are caused by the NAF which has caused many fatal events in northern Turkey. Earthquake epicentres show a clear linear pattern from the Sea of Marmara into the north Aegean Sea.
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Markopoulos et al. (2012) do not provide any structural data for these earthquakes. There are however other publications who present earthquake focal mechanism data from the Aegean. Shaw and Jackson (2010) carried out an investigation of the Hellenic subduction zone using GPS and revised focal mechanisms and epicentres for a number of earthquakes based on wave modeling. The authors show that, apart from close to the Kefalonia transform fault, thrust faulting earthquakes show a divergence of slip vectors around the arc, consistent with GPS velocities. This divergence of slip vectors causes extension in the overriding material, and this is observed in shallow (<20 km) normal faulting earthquakes representing E-W extension. This very recent extension appears to be concentrated in the east and west of the arc (either side of Crete) rather than the middle (Shaw and Jackson, 2010).
Figure 1.6. a) Instrumental seismicity data for the eastern Mediterranean (modified after Becker et al. 2013). Orange vectors are plate velocities; focal mechanisms are from CMT (www.globalcmt.org) between 1977 and 2013 with depths <50 km, colour indicates the deformation style; grey circles are earthquake hypocentres with depths <50 km. b) Instrumental seismicity data for normal faulting in the Aegean, earthquake depths <50 km (from Kiratzi, 2014). Dark green shaded beach balls show N-S extension in the back-arc region, whereas the light green shaded beach balls show ~E-W intra-mountain extension (Hellenides Mts) which connects with the along-arc extension occurring in the overriding Aegean plate.
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Becker et al. (2013) present instrumental earthquake data from the CMT catalogue (www.globalcmt.org) for earthquakes throughout the globe. The data from the eastern Mediterranean (Fig. 1.6a) show a wide range of normal, reverse and strike slip earthquakes distributed throughout the Aegean region. Regarding normal faulting, Kiratzi (2014) presents instrumental earthquake data from a variety of sources (see Kiratzi (2014) and references therein) which shows that N-S extension is mainly confined to continental Greece and western Anatolia (dark green beach balls in Fig. 1.6b); whereas E-W extension is occurring along a distinct belt from the mountains of inner Albania, through the Hellenides Mountains and along the southern Aegean to western Anatolia (light green beach balls in Fig. 1.6b). This E-W extension is occurring south of the Volcanic Arc in the southern Aegean and represents deformation within the overriding Aegean plate.

As part of the SHARE (Seismic Hazard Harmonisation in Europe) project, Gardini et al. (2013) undertook a PSHA (probabilistic seismic hazard assessment) for the whole of Europe. The authors used the earthquake catalogue, fault source database, crustal strain rate models and the strong ground motion database to carry out the PSHA. Figure 1.7 shows the seismic hazard map of the Aegean region. The areas with the highest expected ground accelerations are along the west coast of Albania, mainland Greece and northwest Peloponnese, the Gulf of Corinth, southwest Turkey and eastern Crete, and the area around the NAF. However, even outside these areas, the whole Aegean region can classified has having a high seismic hazard with very few exceptions.

Seismic hazard assessment is predominantly based on historical and instrumental seismicity. However, the normal faults throughout the Aegean have low slip rates and long recurrence intervals; they may not have ruptured in the instrumental period and are often absent from historical catalogues. Earthquake recurrence intervals on individual faults are commonly several hundred to several thousand years, and therefore the time period covered by the historical record in many cases is shorter than the earthquake recurrence interval. Therefore, slip-rate and earthquake date determination is needed for these individual faults so that this data can be used to improve the accuracy of seismic hazard assessments.

There are a number of well documented earthquakes that have occurred in the last 40 years that have caused damage and destruction in the region; for example, the 1981 Corinth earthquake sequence has been extensively studied by many authors (see Jackson et al. 1982; Roberts 1996). The study areas in this thesis are the western Peloponnese (study area 1) and eastern Crete (study area 2). Two significant earthquakes have occurred in the western Peloponnese, the first occurred in 1993 near Pirgos, and the second occurred in
2008 in SW Achaia around 30 km northwest of Pirgos. For details of these events see Chapter 3. No significant earthquakes have stuck eastern Crete in last 200 years, but there are a number of events in the historical record that most probably relate to shallow earthquakes. These occurred in the following years: 1508, 1595, 1717, 1780 and 1815, and a summary of these events is provided in Chapter 4. Two events occurred within the Cretan Sea in the last century to the north of study area 2 and these are described in Chapter 4.

1.2 The rupture process for normal faults

1.2.1 The deformation cycle, earthquake recurrence and fault behaviour

The earthquake deformation cycle for normal faults can be divided into four phases: interseismic, preseismic, coseismic and postseismic. However, we can only reliably measure the coseismic and postseismic phases. This is because for faults that are presently in interseismic phase of the cycle, the preceding state’s topology is not known (for historical earthquakes the topology of the preceding state is known). Therefore, there is a lack of geological evidence in the deformation cycle (McCalpin, 2009). The interseismic phase is the
period in-between earthquakes where strain accumulates steadily. During this phase the fault is locked and little or no motion will occur. The preseismic phase is hard to define in terms of its character and mechanism. It involves precursory earthquake phenomena such as seismicity patterns and foreshocks, hydrological changes to subsurface water. There is, however, only fragmentary evidence for these precursory effects (Scholz, 2002). The coseismic phase occurs during the instant of the earthquake, and the vast majority of slip occurs during this phase. The postseismic phase refers to the slip that occurs in the relatively short time after the earthquake (days to many months), encompassing all aftershocks (Scholz, 2002). One of the best ways to observe the coseismic and postseismic phase on a large scale for recent earthquakes is differential synthetic aperture radar interferometry (DinSAR). In these images centimetre scale deformation can be observed.

In 1910, Reid proposed a model to describe strike-slip deformation based on the interseismic and coseismic phases of earthquakes. At a sufficient depth the geological material can be continuously and aseismically deformed as it is ductile and higher than 350°C (Sibson 1984; 1989). Towards the surface deformation is brittle and concentrated along faults. Far field interseismic deformation continuously occurs and induces a sigmoidal bend perpendicular to the fault, which is locked by asperities and roughness. When the stress exceeds the locking strength it ruptures and the near-field coseismic deformation catastrophically catches up with the far field deformation. In this model (Fig 1.8a), earthquakes occur whenever constant stress builds up to a given level; therefore the model suggests that the earthquake cycle is perfectly periodic, magnitudes of earthquakes are the same and occurrence is predictable at a point on a fault. Shimaki & Nakata (1980) further developed Reid’s model into two more possibilities where time or slip is predictable at a point on a fault. In the time-predictable model (Fig 1.8b), the stress threshold where failure occurs is known and constant. As we know the slip of the previous earthquake and the rate of strain accumulation (a constant slip rate is assumed), the time of the next earthquake can be calculated. In the slip-predictable model (Fig 1.8c) the stress threshold where failure stops is known and constant. As we know when the previous earthquake occurred and the constant rate of strain accumulation, the slip of the next earthquake can be calculated.
Figure 1.8 a-c: models of how a fault can cause displacement over time at a single point on the fault: a) perfectly predictable model, b) time predictable model and c) slip predictable model; the upper graph in each model shows stress over time, and the lower graph shows the cumulative coseismic displacement over time (modified from Scholtz 2002). d-h schematically show the along strike cumulative slip for each fault behaviour model (modified from Berryman and Beanland 1991). f applies most to normal faults in the Aegean; however, the maximum displacement is at the centre of the fault and diminishes towards the tips.
As slip varies along strike several other models of fault behaviour were developed (Berryman & Beanland 1991). Figure 1.8d-h shows 5 models of fault behaviour along the strike of a fault; these models try to categorise how coseismic deformation occurs along a fault over time (McCalpin, 2009). Each model relates the variability of: displacement at a point, slip rate along the fault, and the size of an earthquake. The models (Fig 1.8d-h) assume that there is a relationship between the time between surface rupturing earthquakes and the amount of strain released. The model that applies most to normal faults in the Aegean is the characteristic earthquake model (Fig. 1.8f). Here we have variable slip along strike and this slip persists and grows with each earthquake; however, for Aegean normal faults the maximum displacement is at the centre of the fault and diminishes towards the tips. This characteristic earthquake model gives rise to variable tectonic relief and bedrock blocks being stranded at intermediate structural levels between fault strands. These are observed a lot in the footwall blocks of normal faults. However, this model does assume that most strain is released in large earthquakes within a narrow, “characteristic” magnitude range within a one-magnitude (Schwartz & Coppersmith, 1984) and this may not be the case.

1.2.2 Fault segmentation, uplift and subsidence

faults often comprise several slightly irregular fault traces which are separated from adjacent fault by gaps or stepovers. These smaller individual normal faults are known as segments and collectively form larger faults or fault zones. Individual earthquakes rarely rupture entire fault zones; therefore for seismic hazard analysis an estimate for how much a given fault zone will rupture during an earthquake is required to characterise earthquake size (Depolo et al. 1991). There is no accepted criterion for the naming of individual segments or faults comprising many segments. Therefore, within the literature you will find that many named longer faults are actually composed of several segments (McCalpin, 2009). Segment boundaries for normal faults are typically defined based on static criteria such as structural, geological and geometric evidence (e.g. Fig. 1.9a). However, the best way to define segments and their linkage it to study their behaviour in terms of slip rates and palaeoearthquakes (McCalpin, 2009). In general, large M > 7 earthquakes can rupture multiple segments, whereas for smaller M < 6 earthquakes it is more likely that only one segment will rupture (Suter, 2006). The normal fault systems in Greece are often segmented and the terminations of these segments are known as transfer zones; it is here where the deformation between offset fault segments is accommodated. Figure 1.9a showing the typical features which are present in segmented fault zones in the Corinth Gulf, Greece. Here, the footwall mountains comprising pre-rift stratigraphy are juxtaposed against hanging-wall syn-rift stratigraphy overlain by fan/deltas; modern lakes and marine waters are also
found in the hanging-wall (Roberts and Koukouvelas, 1996). Segments can be arranged with both en-echelon and end-on geometry (Fig. 1.9b).

When a normal faulting earthquake occurs it involves both uplift and subsidence. The footwall block is uplifted and the hanging-wall block subsides (Fig. 1.9b), and in general there is more subsidence than uplift. Several authors have attempted to calculate the long-term ratio of uplift to subsidence (U:S) in Greece, which takes into account both coseismic movement and post-seismic rebound. Armijo et al. (1996) was the first to attempt this at the
Gulf of Corinth by correlating uplift with sea level. The authors determined a ratio of 1:3.5 if there is no regional uplift, and 1:2.7 if regional uplift is 0.2 mm/yr. McNeil & Collier (2004) calculated around 1:2 – 1:3.2 also in the Gulf of Corinth. This was determined by measuring the depth to the basement with gravity modelling and seismic reflection (Myrianthis 1984; McNeil & Collier, 2004). However, it is not clear how the authors defined their reference line to distinguish between uplift and subsidence. Stiros et al. (2007) used releveling and dislocation modeling to calculate a ratio of between 1:2 and 1:3 at the Kaparelli fault in eastern Corinth Gulf.

Figure 1.10. Coseismic differential interferogram of the April 2009 L’Aquila earthquake, covering the period between April 2008–April 2009. The red line shows the trace of the ruptured zone as identified by DinSAR. The footwall to the northeast is uplifted and the hanging-wall to the southwest is subsided (from Papanikolaou et al. 2010).

Regarding a coseismic ratio of uplift vs subsidence (no post-seismic rebound), the best way to calculate this over a large scale is with DinSAR (Differential Synthestic Aperture Radar Interferferometry) which allows cm scale vertical movements to be observed from satellite data. Papanikolaou et al. (2010) carried out a detail study of coseismic uplift/subsidence after the 2009 L’Aquila earthquake in Italy (Fig. 1.10). The authors determine an average ratio over the entire ruptures area of 1:2.9 (approx. 8.5 cm of footwall uplift and 25 cm of hanging-wall subsidence). Higher ratios have been observed in other parts of the world for individual earthquakes, e.g. the 1983 Borah Peak Earthquake in Idaho, USA, had a ratio of 1:9 (McCalpin, 2009). Post seismic rebound is usually in the region of 5-10% depending on the rheology and seismogenic thickness; therefore, for Greece and the wider Mediterranean
region a long-term U:S ratio of between 1:2 and 1.3 is likely representative for normal faulting, but further investigations are necessary to refine this figure.

Normal faults throughout the Aegean undergo greater along strike extension in the hanging-wall compared to the footwall and this asymmetry is thought to induce oblique fault slip close to segment boundaries. At the centre of the segment there is pure dip-slip movement, but at the segment boundaries there is a strike slip component (Roberts, 1996). For bedrock faults this kinematic information can be preserved as slickensides and used to study fault segmentation (see Chapter 2).

1.2.3 Aegean post-glacial fault scarps

The majority of the large normal faults or fault segments within the Aegean are comprised of bedrock fault scarps which are juxtaposed against Quaternary alluvial-colluvial deposits (Fig. 1.11) and/or marine sediments. These faults are easy to recognise as they offset smooth mountain slopes and have steeply dipping fault scarps that are several metres in height. These preserved fault scarps are coseismic and result from cumulative earthquake events on the individual fault. In the Mediterranean the common theory (Benedetti et al., 2002) is that during periglacial conditions the erosion rate of these bedrock fault scarps from vigorous freeze–thaw action, and subsequent deposition on the hanging-wall, was faster than the fault’s throw-rate. In addition, the harsh climate restricted the vegetation, promoting high erosion rates. This resulted in smoothed hillsides that are similarly graded either side of fault scarps (Giraudi 1995; Roberts & Michetti 2004; Papanikolaou et al. 2005) and the bedrock fault scarp itself was not preserved in the landscape (Fig. 1.11a).

The existence of well-preserved Holocene bedrock fault scarps along active normal faults in the Mediterranean region suggests a dramatic reduction in rates of rock weathering and erosion that correlates with the transition from a glacial to interglacial climate (Tucker et al. 2011). Tucker et al. (2011) applied a numerical model calibrated with 36Cl isotope data from the Magnola postglacial fault scarp in central Apennines and calculate that the maximum hillslope erosion rate during glaciations was thirty times higher than the modern rate of scarp weathering. The major glacial retreat phase in the central Apennines began at about 18 ka, dominating the present geomorphology of the region where slopes have been preserved because vegetation stabilised the mountain slopes and erosion rates decreased producing a thin (ca. 1–2 m) organic soil after the demise of the glaciation (Giraudi, 1995; Giraudi & Frezzotti, 1997). An extensive database of radiocarbon dates and tephrochronology reveal the absolute and relative ages of the late glaciation-related deposits and slopes in central...
A postglacial age is also confirmed by 36Cl cosmogenic isotope dating of carbonate fault scarps in central Apennines (Palumbo et al., 2004; Schlagenhauf et al., 2010) and in southern Greece (Benedetti et al., 2002). Indeed, the Peloponnese and Crete display evidence of former glaciation and intense periglacial activity (Mastronuzzi et al., 1994; Hughes et al., 2006; Hughes & Woodward 2008). In particular, Mastronuzzi et al. (1994) describe glacial landforms towards the eastern flank of the Taygetos Mt in the southern Peloponnese and Poser (1957) in Crete. These postglacial bedrock scarps vary in height along strike, exhibiting higher values towards the centre of the faults that diminish towards the tips (e.g. central Apennines (Roberts and Michetti, 2004) and southern Apennines (Papanikolaou and Roberts, 2007)). The latter confirms that these scarps have resulted by tectonic activity and excludes any exhumation by differential erosion which would imply a constant scarp throw value along fault strike (Roberts and Michetti, 2004). Therefore, using a date for first exhumation and the scarp height, a throw rate can be calculated. 15 ± 3 ka is commonly used for the date of first exhumation (Papanikolaou et al., 2013). This 15 ± 3 ka range is due to the uncertainty on the exact post-glacial age estimate; 18 ka is the initiation of the last glacial retreat (Allen et al., 1999) that dominates the present geomorphology of the region, and 12 ka is the youngest reported age in the literature since some small magnitude glacial re-advances followed by retreat phases have been recorded between 12 and 18 ka, predominantly between 14 and 18 ka (e.g. Giraudi & Frezzotti, 1997).

The faults with preserved scarps throughout the Aegean (Fig. 1.11b) are considered to be capable; a capable fault is defined as a fault that has significant potential to cause displacement at or near the ground surface (International Atomic Energy Agency (IAEA), 2010). Hence, in palaeoseismological investigations a fault is considered capable if it has
been active since approximately the last interglacial period in the Quaternary. Normal fault scarps in the Aegean range in height from decimetres to many metres. Care must be taken when measuring scarp heights to estimate long-term slip rates as natural erosion from gullies or anthropogenic activity can artificially increase the scarp height; however, the preservation of many several-metre high scarps implies that regular seismic activity must have been occurring for a significant period since the last glacial maximum.

1.3 Geological summary of Study Area 1: The Lapithas Mountain, western Peloponnese

Study area 1 lies within the region of Elis and northern Messinia and is located at the centre of the Kyparissiakos Gulf (Fig. 1.12). The Alpine formations forming the Lapithas Mountain Horst comprise parts the Gavrovo-Tripolis unit (both H1, Fig. 1.2), and also the Pindos unit (H2, Fig. 1.2). The Pindos unit comprises a Mesozoic pelagic limestone sequence. The underlying Gavrovo-Tripolis unit comprises Mesozoic to upper Eocene shallow water carbonates and upper Eocene-Oligocene flysch (Papanikolaou et al., 2007). The Lapithas Mountain forms the northern most Horst in a series of horst and graben structures which are displaced by a series of E-W orientated faults. The horsts to the south of the Lapithas Mountain comprise limestones of the Pindos unit. Within the grabens, continental and marine sedimentary sequences of Pliocene and Quaternary age unconformably overly the Alpine formations.

To the south of the Lapithas Mountain is the Zacharo Basin. The eastern part of the basin comprises a lower lacustrine sequence of Pliocene age, and in its western part is an upper sequence of marine lower Pleistocene sediments. To the north of the Lapithas Mountain is the Olympia Basin. Here, the sediments comprising the basin mirror those to the south of the mountain (Fig. 1.12). Geophysical prospecting and drilling has been undertaken in the Olympia Basin and there are over 3 km of Pliocene-Quaternary sediments (Hageman, 1977). The E-W striking neotectonic faults within the Kyparissiakos Gulf are the reason for this juxtaposition of H1 and H2 terranes against the Pliocene-Pleistocene sediments, and the thick sequences implies intense neotectonic activity throughout the Quaternary. These E-W striking faults have been previously studied (Lekkas et al., 2000; Papanikolaou et al., 2007; Fountoulis and Mariolakos, 2008); however, up to now fault-slip data have not been used to structurally analyse the bedrock faults and precise shear senses are unknown. A detailed description of the Lapithas Mountain faults is provided in Chapter 3.
1.4 Geological summary of Study Area 2: The Lastros-Sfaka Graben, eastern Crete

Study area 2 is the Lastros-Sfaka Graben located in eastern Crete (Fig. 1.13). The Lastros-Sfaka Graben forms part of the Ierapetra fault zone which consists of a roughly 25 km long zone of onshore faults segments most of which dip to the NW. The westernmost and longest is the Ierapetra fault which traverses nearly the whole width of the island. The Lastros fault dips to the SE and forms the SE boundary to the Kapsos ridge horst. The easternmost Sfaka fault has an antithetic relationship with the Lastros fault dipping to the NW (Fig. 1.13).

The footwall mountains of both the Lastros and Sfaka faults comprise crystalline limestone of the Mani unit (H1, Fig. 1.2) (Papanikolaou & Vassilakis, 2010) which is heavily fractured and brecciated close to the fault plane (Stewart & Hancock, 1990). The hanging-wall of both faults comprises limestone of the Mani Unit overlain by either Quaternary colluvial deposits, or Permian-Triassic phyllites (part of the Western Crete unit and Tripolis unit, both within H1, Fig. 1.2) partly overlain by Quaternary colluvium. In the centre of the north part of the graben near the town of Lastros, Miocene marine sediments overly the phyllites and small fluvial terraces of Pleistocene age are preserved. Intercalated within the phyllites are evaporites deposits; a large gypsum and anhydrite quarry is located to the south of Mochlos (Fig. 1.13).
In many places on both the Lastros and Sfaka faults the Quaternary colluvium has become cemented and is attached to the fault plane in various forms. A detailed description of the Lastros and Sfaka faults is provided in Chapters 4 and 5.

![Geological map of study area 2. (modified from Papastamatiou et al, 1959; Papanikolaou and Vassilakis, 2010). Inset: the island of Crete showing the study area location.](image-url)
Chapter 2: Methodology - bedrock fault scarp and hanging-wall investigations of normal faults in Greece

This chapter provides an overview of investigation techniques in neotectonics and palaeoseismology, setting the scene for the studies undertaken in this research. In the interest of completeness, a summary of absolute and relative dating of bedrock scarps is provided in section 2.1.3; however, these techniques were not used in this research. A summary of the requirements for thorough palaeoseismological investigations is also provided. Lastly, the rationale for study area selection is described. There have been a number of investigations previously carried out in both study areas and a comprehensive summary of these investigations is provided in Chapters 3, 4 and 5.

2.1 Footwall Investigations

2.1.1 Fault slip data and kinematics

Bedrock fault planes often preserve indicators which express the local conditions of faulting and show the coseismic strains the fault planes were subject to. By measuring their plunge and azimuth, and comparing it to the dip angle and dip direction of the fault plane, the relative motion of the footwall and hanging-wall can be determined; hence, the components of a normal, reverse or strike-slip shear can be defined. As first noted by Roberts (1996) many bedrock normal faults throughout the Mediterranean do not show pure dip-slip movements over their entire length. At the centre of the fault kinematic indicators show dip-slip movement, but towards the ends of the fault the kinematic indicators have a large strike slip component, converging towards the centre of the fault (Fig. 2.1a). This is because throw gradients produce stretching of the ground surface along strike, so slip-directions at the tips of the faults are oblique and converge towards the hanging-wall to accommodate this stretching (Roberts 1996; Roberts & Ganas, 2000). This is also observed for smaller fault segments making up larger fault zones. Individual ruptures do not always breach the whole fault or fault segment. Smaller breaches can lead to overprinting of the kinematics because individual ruptures will have dip-slip movement at the centre of the breach and a strike slip component towards the ends. At the centre of the fault it is therefore possible that slip directions can vary up to 90° (Fig. 2.1b). Thus, by carrying out a structural analysis on measured kinematic indicators, the orientation and variation of extension can be determined and used to accurately map faults and fault segment lengths.
Figure 2.1. a) Schematic view of the variation in fault-slip directions along a normal fault in Greece (modified after Roberts & Ganas, 2000). b) Diagram showing how kinematics may vary depending on what section of the fault was breached by individual ruptures. Coloured arrows show slip vector azimuths for three separate ruptures; rupture 1 = orange, rupture 2 = yellow, rupture 3 = red (modified after Roberts, 2007).

Kinematic indicators come in many forms such as steps, crescentic markings, tool marks, etc. (Hancock & Barka, 1987); however, by far the most common types observed are striations and corrugations (Fig. 2.2). Striations are cataclastic lineations caused by abrasion or ploughing parallel to the direction of relative fault movement. Often these striations are iron coated caused by iron precipitation when water flows down the fault plane. These are often several centimetres wide and can be traced for many metres. Corrugations are at a larger scale of several metres wide. Both the striations and corrugations were measured using a conventional compass clinometer. For the corrugations it is harder to determine the exact azimuth and plunge due to their scale and shape; however, this can be done to an accuracy of ±2 degrees. The locations of measurements are measured using a hand-held
GPS with an accuracy of several metres. The orientation of kinematic indicators can then be plotted along strike of the mapped fault allowing more accurate fault lengths to be determined. Using software programmes such as *TectonicsFP* (Reiter and Acs, 2014) principle stress orientations can also be determined at each data acquisition point along strike.

Figure 2.2. Sketch of a typical postglacial normal fault showing bedrock juxtaposed against Quaternary sediments. The bedrock scrap shows both striations and corrugations which can be used as kinematic indicators. Quaternary sediments contain structures caused by recurrent earthquakes (modified after Reicherter et al. 2003; Roberts & Ganas, 2000).

### 2.1.2 Observable throw rate calculations (GIS)

The observable throw of the footwall mountain above the fault also changes with respect to the centre of the fault (e.g. Roberts 1996; Roberts and Ganas, 2000). The term observable throw is used here rather than total throw, because the total throw would be measured from the top of the footwall mountain down to the unconformity below the hanging-wall sediments. Therefore, the observable throw, measured from the top of the footwall mountain to the fault scarp (Fig. 2.3), is considered as a proxy for total throw. As with the kinematic indicators showing dip-slip movement, the observable throw is largest at the centre of the fault and reduces towards the fault tips. There is variation in observable throw caused by non-tectonic
features such erosion as catchment gullies (Fig 2.3). However, as a general trend the observable throw decreased towards the fault tips.

Digital elevation data can be used along with fault measurements from compass clinometer and GPS to accurately determine the length of faults. Once these lengths are established, empirical calculations such as Wells and Coppersmith (1994) are used to determine maximum possible earthquake magnitudes and vertical displacements. The Wells and Coppersmith dataset was compiled from fault lengths and surface displacements obtained from field studies of earthquakes up to 1994 where the magnitude was known.

![Figure 2.3. Annotated photograph of a typical bedrock fault (Lastros fault, Crete) showing an overall decrease in observable throw (yellow arrows) to the north. Catchment gullies erode the footwall mountain and hanging-wall leading to local changes in observable throw.](image)

**2.1.3 Fault scarp dating (absolute and relative)**

Dating of earthquake events is a central issue for palaeoseismological studies. As the bedrock normal faults in the Aegean exhibit preserved fault scarps, it is possible to date when these scarps were first exposed to cosmic rays. This fault scarp dating technique, known as cosmogenic nuclide dating, has been carried out in central Greece (Sparta and Kaparelli fault, Benedetti et al., 2002, 2003), central Italy (Magnola-Velino fault, e.g. Palumbo et al., 2004), the Hebgen Lake fault in Montana US (Zreda and Noller, 1998; Gran-Mitchell et al., 2001), and northern Israel (Nahef East fault, Gran-Mitchell et al., 2001). In all of these studies carbonate fault planes were sampled in the direction of slip (determined through kinematic indicators) and subsequently dated based on cosmogenic nuclide $^{36}$Cl concentration. Locations on the fault plane where there is a rapid change in $^{36}$Cl concentration define the sections that were exhumed by different earthquakes. These, therefore, define the palaeo-ground levels or event horizons for each earthquake. When these event horizons are known, the $^{36}$Cl concentrations can be transformed into ages that
date each earthquake event (Palumbo et al., 2004; Schlagenhauf et al., 2010). The distance between event horizons represents the amount of slip during each earthquake. This can then be used, in combination with empirical relationships (e.g. Wells and Coppersmith, 1994; Pavlides and Caputo, 2004), to estimate the magnitude of palaeoearthquakes. Event horizon determination is, therefore, the controlling factor when interpreting earthquake history. Fault scarps can also be relatively dated by correlating the enrichment/depletion of rare earth elements (REE). This technique has been successfully used on the Magnola-Velino fault in Italy by Carcaillet et al. (2008) and Manighetti et al. (2010). The Magnola-Velino fault was chosen as cosmogenic nuclide dating had already been undertaken on this fault and its Holocene earthquake history is known, allowing correlations and comparisons to be made. This technique has also been carried out on the Spili fault (Mouslopoulou et al., 2011) and differing concentrations of REE have been identified on the scarp, the cause of which is interpreted as being exposure time variations. The application of the $^{36}$Cl dating technique relies hugely on sample site selection, and there are many pitfalls associated with this which need to be considered. A perfect sampling site would be a fresh bedrock scarp located away from nearby gullies; the location must also show no evidence for any anthropogenic activity which may have contributed to the scarp’s exhumation. On Crete, sites like this are few and far between. On many faults, what looks like the exposed bedrock scarp from a distance, can actually have cemented colluvium attached to the fault plane. These areas need to be avoided as studies at these locations may provide erroneous dates and lead to incorrect interpretations. Schlagenhauf et al. (2010) and Manighetti et al. (2010) provide detailed methodologies for $^{36}$Cl and REE dating respectively. For dating other bedrock lithologies, other cosmogenic nuclides like $^{10}$Be, $^{21}$Ne or $^{26}$Al should be used (Ivy-Ochs and Kober, 2008).

Terrestrial Light Detection And Ranging (t-LiDAR) scanning of faults scarps has been successfully used to provide insights into active faulting and can be used to take very accurate measurements of fault plane geometry as well as assist in fault plane sample site selection. Surface parameters changes like colour, roughness, slickensides or karstification can be revealed using t-LiDAR (e.g. Renard et al., 2006; Sagy et al., 2007, Sagy and Brodsky, 2009; Wiatr et al., 2011, 2013). Roughness analysis by Wiatr et al. (2015) has shown that fault plane roughness varies over fault plane height which is related to time since first exposure; relative dating can then be undertaken. The backscatter signal reveals the combined surface conditions on the fault plane, which can identify different ribbons of weathering (Wiatr et al., 2011). As roughness and differential weathering relate to exposure time, it is likely that palaeoearthquakes are the cause.
2.1.4 Postglacial throw and slip rate determination using t-LiDAR

As stated in Chapter 1 the common theory for bedrock fault scarp preservation in the Mediterranean is that during glacial times footwall erosion and sedimentation in the hanging-wall was higher than the throw rate of the fault. When climatic conditions improved after the LGM (Last Glacial Maximum) erosion and sedimentation rates reduced and the scarps then became preserved in the landscape. The date for first exhumation is 15 ± 3 ka (see Chapter 1 for the background information to this number). Therefore an accurate scarp profile (Fig. 2.4) in an area which has not been affected by gully erosion or anthropogenic activity can be used to calculate a throw rate. Traditionally profiles were done manually with a measuring tape and a compass (e.g. Papanikolaou et al., 2005). However, with t-LiDAR a high resolution DEM (Digital Elevation Model) can be produced and then scarp profiles can be undertaken using geographical software (e.g. Wilkinson et al., 2015).

t-LiDAR is a laser scanning method which generates point cloud data. The ILRIS 3D from OPTECH Inc. was used in this research. This static laser scanner sends a pulse of lasers in a zig zag pattern which covers the scan window. The individual lasers are then reflected off objects. The laser scanner then measures the backscattered signal that is returned to the scanner position. Each reflected laser point contains x, y and z coordinates and the backscatter signal strength. The position and orientation of the scanner is measured with GPS and compass. This allows the point cloud to be georeferenced when back in the office. Once the scan is completed, the scanner is moved to the next position and a new scan window is carried out. There must be some overlap of scan windows in order for the point cloud data to be combined accurately. Once point cloud data from all scan windows is combined, a 0.5 x 0.5 m raster is created from the combined data set. This raster provides an accurate DEM of the scanned area. With this DEM a huge amount of fault scarp profiles can be undertaken.
Figure 2.4. Typical scarp profile showing the parameters needed to calculate fault throw. This example profile has been extracted from a t-LiDAR point cloud raster.

The vertical difference between the footwall slope and hanging-wall slope can be used to determine the vertical offset which in turn can be used to determine a post glacial throw. Two methods were used to calculate this vertical offset: 1) traditional profiles, and 2) polygons and zonal statistics. To construct profiles the profile coordinates must be extracted from GIS and imported into excel. The data must then be exported from excel into MATLAB to ensure the profile is at correct scale and subsequent measurements are accurate. Lines for the average footwall and hanging-wall slope are then created and the vertical difference between where these two lines intersect the projected fault plane is measured (Fig. 2.4). The polygon and zonal statistics method create polygons within GIS which cover the scarp. To define the polygon size the slope tool is used. 10 m wide polygons are created with the top and bottom of each polygon being to the top and bottom of the scarp respectively. The slope classification is set to 5 classes. The class with the highest slope angle, and hence the fault scarp, it between 80° and 52°. The top and bottom of each polygon is defined manually as the change between this upper class and the classes surrounding it (hanging-wall and footwall). Zonal statistics are then used in each polygon to define the vertical displacement. The range in elevation between mean 20 highest and mean 20 lowest points in the polygon is used for this. The two methods are then compared to verify that the fault scarp angle is between 80° and 52°. If the profile shows a smaller or larger vertical offset then the polygon’s size is adjusted accordingly. Results are presented with the error (sigma 1) showing the variation from the mean in each polygon. The average postglacial throw in each polygon is
then used in combination with the dip of the fault and 15 ± 3 ka for first exhumation (see Chapter 1) to determine the slip rate and age uncertainty.

### 2.2 Hanging-wall investigations

#### 2.2.1 Trenching

The technique of palaeoseismic trenching has been used for decades to identify earthquake event horizons and collect samples for dating palaeoearthquakes. For normal faults where bedrock fault scarps are juxtaposed against Quaternary alluvial-colluvial sediments, trenches can obviously only be excavated in the hanging-wall where no cemented colluvium is present. Here, evidence for palaeoearthquakes may be found within the sedimentary architecture (Fig. 2.2).

The ideal palaeoseismic trench site would be in an area where there is continuous sedimentation on both the footwall and hanging-wall, such as an alluvial fan which covers both the footwall and hanging-wall (Fig. 2.5). However, sites like these are very rare on normal faults in Greece and Crete. Nevertheless, hanging-wall sedimentation is necessary to preserve stratigraphic evidence of earthquakes and locations must be where there are at least traps for sediments mobilised by intense ground shaking. High peak ground accelerations will result in slope processes such as shallow, deep seated slides and rock falls. These cause increased erosion and sedimentation which are enhanced by high slope angles (Keefer, 1994). Purpose built man-made trenches are the most common; however, other hanging-wall outcrops such as road cuts or natural exposures can also be used for palaeoseismic investigations provided their locations allows for sedimentation in the hanging-wall.

Relatively few trenching investigations have been carried out to date in mainland Greece and Crete. The 1981 earthquake series struck the Gulf of Corinth (see Chapter 1 for description), and ruptured three faults: the Pissia fault, Skinos fault and Kaparelli fault. A trenching investigation was undertaken on the Skinos fault in 1996 (Pantosti et al., 1996; Collier et al., 1998). The trench site location was chosen where the fault runs through the Bambakies alluvial fan. Three trenches were excavated within the fan and several palaeoearthquakes were identified. These derived slip rates of 0.7-2mm/yr from trench 1 and 1.2-2.5 mm/yr from trench 3. A recurrence interval of 330 years was also determined from trench 3.
The Kaparelli fault was also trenched in 2002 by Kokkalas et al. (2007). The authors excavated three trenches, two of which abutted the limestone bedrock scarp, and the third was excavated within loess deposits which cover the bedrock. Only in this latter trench (KAP-1; Kokkalas et al., 2007) were any coseismic features identified. One palaeoearthquake was dated within the fault zone which occurred 5710-5570 BP. Coarse grained fill material within the hanging-wall loess was interpreted as formed by tension fissures rapidly filled with overlying gravels. A number of radiocarbon dates were derived from these tension fissure fills; however, their relationship to individual earthquakes is poorly constrained. A re-logging and re-interpretation of this trench was undertaken as part of the development of a new technique to visualise sedimentary structure within trench walls in 3D (Schneiderwind et al., 2016). Within Chapter 6 detailed logs are presented which suggest other possibilities for the formation of these ‘tension fissures’.

Figure 2.5. Sketch showing the features of a typical capable fault on Crete and where palaeoseismological investigations should be carried out. Fault scarp dating should be avoided at locations (1) due to presence of cemented colluvium, (2) due to proximity to gullies, and (3) due to anthropogenic activity. Fault scarp dating should preferably be undertaken at (4) where the scarp height is not influenced by external processes, and trenching investigations should be carried out at (5) where there is continuous sedimentation and/or a sediment trap at the fault.

The only trenching investigation that has been carried out on Crete was undertaken on the Kera fault (a segment of the Rodopos fault zone; see Fig. 4.1) by Mouslopoulou et al. (2001). The trench site is a natural outcrop which appears to have been exposed by either landsliding or quarrying within the hanging-wall. The authors identified at least three
earthquake events thought to be from the Pleistocene and Holocene; however, the study did not provide any age constraints.

2.2.2 Stratigraphic evidence of palaeoearthquakes

Within trench walls, sedimentary structures caused by coseismic earthquakes can be visualised clearly and can be expressed in many ways (Figs. 2.2 and 2.6), e.g. colluvial wedges, disrupted and displaced strata, grabens and half-grabens, sand blows, fissure fills, etc. Datable material within or buried under these sedimentary structures can provide dates for earthquakes, and with several generations of structures formed by different earthquakes, recurrence intervals for individual faults can be determined. Different styles of faulting tend to produce different kinds of earthquake evidence. Where there are bedrock fault scarps juxtaposed against Quaternary sediments, such as the normal faults throughout the Aegean, fissure fills and displaced strata are often the primary sedimentary structures present in the hanging-wall stratigraphy (McCalpin, 2009). This is because the footwall scarp erodes too slowly to form a post-seismic colluvial wedge. During an earthquake the fault refracts enough to form a fissure which does not fully collapse during shaking. Repeated faulting can lead to nested fissure fills. Where aggradation of the hanging-wall occurs between earthquakes, paired deposits form comprising coseismic fissure fills and interseismic graben fill deposits (McCalpin, 2005; 2009).

2.2.3 Trench logging and photomosaics

Trench logging is the process of reporting information from the exposed trench wall to paper with as much accuracy as possible. These logs are used as evidence to present to the scientific community and should be as least ambiguous as possible. However, where stratigraphy is unclear and complex interpretations by loggers may differ. A trench log combines background knowledge, experience, observations and interpretations of the logger, and is therefore not raw data set like photographs are.

Trench walls are first cleaned of any traces left by mechanical shovel (if opened using an excavator) or weathering effects (if open for a long time, e.g. a road cut). A string grid is then installed, usually 1 x 1 m or 0.5 x 0.5 m, and attached to the wall by nails of appropriate size for the stratigraphy. To set the grid accurately a string line bubble level is used for the horizontal string lines. For the vertical string lines a plumb line is used. This establishes a 2 dimensional coordinate system. Logs are drawn at an appropriate scale for the features within the trench, which is usually 1:20. At this scale a 0.5 mm pencil line represents 1 cm on the trench wall. The log is then sketched describing different sedimentological units and
taking appropriate measurements of any features using a tape measure and the string grid. Personal interpretations should be minimised and the log undertaken as objectively as possible.

Figure 2.6. Diagrams illustrating stratigraphic and structural criteria used to identify the occurrence and timing of paleoearthquakes in trench walls: (1) Vertical fault offset sealed by a levelling sedimentation episode; (2) Liquefaction along a fault offset sealed by subsequent sedimentation episodes; (3) Soft-sediment deformation sealed by subsequent sedimentation episodes; (4) Fault-scarp eroded and sealed by subsequent sedimentation; (5) Colluvial wedge derived from scarp degradation on normal fault offset; (6) Filling of open surface fissures along a vertical fault plane, sealed by a levelling sedimentation episode; (7) Normal bedrock fault scarp with offset slope deposits, typical of carbonate fault plane in the Mediterranean Region; (8) Reverse fault offset sealed by a levelling sedimentation episode (from IAEA 2015; modified from Grant, 2002; and Audemard, 2005).
Photographs are used to have a raw data set of the trench wall as trench logs will always be partly interpreted. Photographs provide objective information for verification and discussions when back in the office. When taking pictures of trench walls the main caution is to avoid optical distortion which takes place because trench walls are planar and camera lenses are convex. Distortion can then be quite high at the edges of the photograph. If possible the best way to avoid distortion is to take the photos from as far distance as possible and zoom into the trench wall. This minimises the distortion. Once the whole trench wall is photographed appropriate software, e.g. Adobe Photoshop, can be used to combine the photos forming a photomosaic. If significant distortions have occurred, these can be corrected for within the software using the string grid as a reference.

2.2.4 Earthquake timing: radiocarbon dating

There are two ways to determine earthquakes that occurred prior to the instrumental period: 1) historical records, which are numerous in Greece due to the long history of human habitation; and 2) geological and geomorphological evidence from palaeoseismological investigations. Historical records come in many forms such as ancient scripts, diaries of traders and traveller journals. These records should never be thought of as thorough because records may be missing due to destruction or widespread illiteracy. Furthermore, erroneous translations or misinterpretations can also lead to misrepresentative historical records. If the findings of one inaccurate historical report are published in a respectable journal/book, subsequent citations can over time make this inaccurate history become common knowledge and accepted fact. Therefore, when using historical reports in earthquake studies, care must be taken to find the original source so that there are no ambiguities.

Trenching is the most common technique in palaeoseismology and most trenching studies use radiocarbon dating to constrain the timing of earthquakes. There are other techniques such as OSL (optically stimulated luminescence); however, only radiocarbon was used in this thesis. Radiocarbon dating uses the radioactive decay of carbon ($^{14}$C) to determine how much time has passed since an organism was living. $^{14}$C is formed in the atmosphere from the collision between cosmic ray neutrons and $^{14}$N. $^{14}$C then decays back to $^{14}$N by $\beta$ radioactive decay over a half-life of 5730 years (Walker, 2005). This process occurs in the upper atmosphere. It then gets dispersed throughout the rest of the atmosphere, the oceans and other exchangeable carbon reservoirs on Earth. Living plants and animals absorb carbon from the atmosphere, oceans and from living materials they consume, and therefore contain $^{14}$C. Once an organism dies its ratio of $^{14}$C begins to decrease according to its half-
life. Therefore by measuring the amount of $^{14}$C that remains in the organism, and comparing this to modern $^{14}$C standard material, an age can be inferred for the death of the organism (Walker, 2005). The isotope ratio of $^{14}$C/$^{13}$C or $^{14}$C/$^{12}$C is then measured by beta counting or using a mass spectrometer. This ratio is then used to determine the fraction of modern carbon and it’s uncertainty. Then using the half-life of $^{14}$C a radiocarbon age is determined. This is usually reported as a Gaussian distribution with a 1σ error reflecting the variation from the mean value. Years before present (BP) fixed at 1950 follow the radiocarbon age. This is because the amount of atmospheric $^{14}$C has almost doubled since 1950 due to atomic bomb tests. The amount of $^{14}$C has fluctuated over time due to interactions between solar radiation and the atmosphere, and more recently by human actions in the industrial revolution. Therefore, the radiocarbon age must be corrected for this fluctuation using a calibration curve describing these fluctuations. This results in a non-uniform probability distribution function describing a range of possible ages between 2σ uncertainty limits. The calibration curve IntCal13 (Reimer et al., 2013) and the calibration program OxCal 4.2 (Bronk Ramsey, 2014) were used in this research. Obviously, dates can only be inferred for datable materials. Organic compounds like bones, teeth, snail shells, charcoal chunks and soil may be dated through radiocarbon analysis and samples should be strategically taken from trench walls to date organic material killed soon after an earthquake.

2.2.5 Palaeoenvironment: stable isotope analysis of colluvial cement

As stated in Chapter 1, the hanging-wall of normal carbonate faults throughout the Aegean often contains carbonitic colluvial material that has been eroded and fallen from the footwall mountain above the scarp. In places this colluvial material has become cemented and is stuck to the fault plane either as sheets of varying thickness, or as large talus lobes forming protuberances in the hanging-wall morphology. This cemented colluvium can be classified as a type of travertine, more specifically a cemented rudite (Pentecost, 2005). Pentecost (2005) defines a cemented rudite as consisting of cemented scree, alluvium, breccia and gravel formed as a result of ground and surface water degassing. To study this phenomena stable isotope analysis of carbon and oxygen can be undertaken on the cement that binds the carbonate colluvium together. The cement is formed of CaCO$_3$ (calcium carbonate) which has precipitated from water, and mass spectrometers are used to measure the isotope composition of the carbonate. The ratios of two isotopes pairs $^{13}$C/$^{12}$C and $^{18}$O/$^{16}$O, which are respectively expressed as the delta values $\delta^{13}$C and $\delta^{18}$O, are measured and used determine the source of water from which the cement precipitated, and if meteoric the palaeoenvironmental conditions at the time of precipitation. These ratios relate the isotopic concentration of the sample to that of a standard, and delta value are said to be either
heavier (enriched) or lighter (depleted) than the standard (Pentecost, 2005). The equations for $\delta^{13}C$ and $\delta^{18}O$ are as follows:

$$\delta^{13}C = \frac{^{13}C/^{12}C}_{\text{sample}} - \frac{^{13}C/^{12}C}_{\text{standard}} \times 1000 \text{‰}$$

$$\delta^{18}O = \frac{^{18}O/^{16}O}_{\text{sample}} - \frac{^{18}O/^{16}O}_{\text{standard}} \times 1000 \text{‰}$$

The standard used for carbonates is the VPDB - Vienna Pee Dee Belemnite (originally from the Cretaceous Pee Dee Formation, Urey et al., 1951 but has now been replaced by the artificial Vienna PDB which is a marble) (Pentecost, 2005).

Travertines have a wide range of $\delta^{13}C$ values from -12 to +11 ‰, and the value depends on the nature of the water from which they precipitated. If the travertine has precipitated from meteoric water it is known as a meteogene and has $\delta^{13}C$ values averaging at -10 ‰ (Pentecost, 2005). If the travertine has precipitated from thermal water it is known as a thermogene and typically has higher values, from -3 to +8 ‰. Thermogenes have higher values because of higher amounts of CO$_2$ in thermal waters and limestone decarbonisation. Vegetation and soil activity in wet periods also make $\delta^{13}C$ values more negative and this is having a negligible effect in thermogenes (Pentecost & Viles, 1994; Pentecost, 2005).

The travertine $\delta^{18}O$ values are dependent on the temperature of the water from which they precipitated, and if meteoric the amount of $^{18}O$ in the atmosphere at the time of precipitation. In cooler climatic periods evaporation of water molecules containing lighter $^{16}O$ isotopes is easier than water molecules containing heavier $^{18}O$ isotopes. This leads to a low concentration of $^{18}O$ in the atmosphere compared to warmer climatic periods. Therefore for meteogene travertines with low $\delta^{18}O$ values are indicative of cool climatic conditions as the heavier $^{18}O$ isotopes remain in water reservoirs. High $\delta^{18}O$ values indicate warmer climatic conditions because the heavier $^{18}O$ isotopes have also evaporated and then returned to earth in rainfall. $\delta^{18}O$ values are therefore also dependent on the atmospheric temperature and the geographical position of the travertine (e.g. proximity to the Ocean) (Pentecost, 2005). Furthermore, palaeotemperature calculations of the parent water from which calcite cement precipitated can be undertaken (e.g. Hays and Grossman, 1991; Kim and O’Neil, 1997). These calculations assume a value for the oxygen isotopic composition of the parent water.
2.2.6 Ground Penetrating Radar

Ground penetrating radar (GPR) is a non-invasive, fast and precise investigation technique based on the propagation of electromagnetic waves (Bristow and Jol, 2003; Neal, 2004). During the last years it has shown its potential for locating and correlating sedimentary structures (e.g. bedding, faults, joints & folds in sediments; Bristow and Jol, 2003; Neal, 2004). Davis and Annan (1989) first stated that radio waves with frequencies ranging from 10 MHz to 1 GHz are adequate to image resistive material. Resolution of up to 5 cm can be achieved when using antennas with high frequencies. However, such high frequencies are strongly absorbed by conductive geological materials (wet clay, soil). This is a major problem in temperate regions, but in semi-arid regions such as Greece fieldwork can be undertaken at the correct time of year (April – September) to avoid this. More limiting factors include: uneven and/or rough surface conditions, changing lithologies/soil humidity within a study area, anthropogenic modification of the subsurface, and the presence of metallic objects or other radar wave sources. GPR has been used extensively in fault investigations. Stratigraphic offset of sediments has been identified by many authors (e.g. Reicherter and Reiss, 2001; Anderson et al., 2003; Reiss et al., 2003; Alasset and Meghraoui 2005). Colluvial wedges have been identified in GPR data by a number of authors (e.g. Chow et al., 2001; Christie et al., 2009; Denith et al., 2010); Reiss et al. (2003) identified event horizons beneath colluvial wedges indicating two co-seismic ruptures and post seismic sedimentation. Other structures that have been successfully imaged include small graben structures caused by antithetic faults (Christie et al., 2009), sand blows and fault related folding (Chow et al., 2001).

2.3 Archaeological damage

Studying the effects of past earthquakes from archaeological remains and evidence in the geo-archaeological record is known as archaeoseismology. This discipline constitutes a new independent tool for the analysis of past earthquakes, complementing, validating, checking, calibrating or rejecting the description in the literature from the classical period (Stiros and Jones, 1996). The main questions to be answered by archaeoseismological investigations are: (1) how probable are seismic ground motions or secondary earthquake effects to be the cause of damage observed in manmade structures from the past; (2) when did the damaging ground motion occur; and (3) what can be deduced about the nature of the causative earthquake (Hinzen et al., 2011).
Based on the guidelines of the EEE produced for the implementation of the INQUA ESI-2007 Intensity Scale (Michetti et al., 2007), a classification of Earthquake Archaeological Effects (EAE) has been developed (Rodríguez-Pascua et al., 2011). This classification uses the classical techniques of geological structural analysis used in geology for ductile and brittle deformation to describe earthquake damage in ancient buildings. The classification is based on strain structures due to 'primary effects' (direct effects) and structures generated by 'secondary effects' (indirect effects). Primary effects occur when a seismic wave is transmitted from the seismic focus to the buildings, i.e. due to primary faulting (Rodríguez-Pascua et al., 2011). Primary effects include: geological effects such as fault planes crossing ancient towns, damage by landslides and rockfalls, liquefaction producing tilted walls, monuments buried by seismities, folded and fractured pavements; and building fabric effects, which comprise strain structures generated by permanent ground deformation, tilted and displaced walls, folded and fractured walls, strain structures generated by transient shaking, fallen and orientated columns, displaced masonry blocks, dropped keystones in arches, folded steps and kerbs, impact block marks, collapsed walls and vaults, and dipping broken corners or chipping marks (Rodríguez-Pascua et al., 2011). Secondary effects are caused as a consequence of major earthquakes and include the reconstruction and repair of earthquake related damage such as repaired keystones in arches or replaced fallen columns (Dinsmoor, 1941; Stiros and Jones, 1996; Rodríguez-Pascua et al., 2011). By structurally analysing the aforementioned primary and secondary effects of archaeoseismic deformation, the estimation of the strain ellipsoid associated with the ground motion during the causative paleoearthquake can be achieved. This allows for the identification of the preferable orientation of the theoretical ground motion (ey axis) which can then be correlated in relation to relevant nearby neotectonic structures (Rodríguez-Pascua et al., 2011). Geometric relationships between the orientation of damage and likely seismic sources can then be determined.

2.4 Palaeoseismological research requirements

In order to determine the earthquake history for the individual capable faults in Greece, the individual fault needs to be extensively studied. The flow chart in figure 2.7 summarises the general procedure and these steps are explained in detail below. Figure 2.5 schematically summarises the locations where investigations should be undertaken. Not all of these steps were used in this thesis due to various constraints; however, the following text explains the ideal requirements for future research.
The first step is regional reconnaissance which will begin with a desk study. Here, it is necessary to obtain aerial photographs and satellite images of the study area in order to start a detailed study of potential fault scarps prior to undertaking time-consuming and expensive field surveys. For example, Klinger at al. (2003) used 1 m resolution images to map sections of the Kunlun fault in Tibet where the magnitude (Mw) 7.9 Kokoxili earthquake occurred in 2001. The high zooming capability allowed the authors to identify metre scale displacement features such as ridges and deflected river streams. Purchasing TerraSAR-X Tandem satellite images is recommended. These will deliver radar imagery to 1 m resolution. As the terrain of the study area is mountainous and most areas can only be accessed by foot, these images will provide fast, unique and critical information so that the field mapping campaign can be focused on faults where neotectonic movement indicators can be observed. Even higher resolution images (cm scale) can be acquired with photographs taken from unmanned aerial vehicles (UAVs; e.g. hexacopter) or helikites (combination of helium balloon and a kite). Airborne LiDAR can also be used to obtain high resolution DEMs with the advantage that most vegetation can be removed from the model. With aerial and satellite images, a tectonic geomorphology study can also be undertaken. This will allow the tectonic influence on the morphology and drainage pattern to be determined by quantitatively measuring geomorphic indices such as slope gradient, mountain-front sinuosity, percentage of faceting along mountain fronts, valley floor ratios, and stream gradient changes (see Burbank and Anderson, 2011). Once the individual fault extent has been determined from aerial and satellite images, field reconnaissance needs to be undertaken to define a study area. This will involve fault mapping of individual faults or fault segments in order to define their length and select areas for detailed investigations. Fault plane measurements should be undertaken noting the presence of cemented colluvium on the fault scarp and/or the hanging-wall. Gullies should be mapped as well as any areas which have the potential to be affected by human activity; access possibilities for equipment should also be considered. Upon choosing and gaining access permission to an appropriate study area detailed reconnaissance can begin.

If wanting to undertake a fault plane dating investigation, a study area containing no cemented colluvium on both the fault plane and hanging-wall is a prerequisite. If any cemented colluvium was mapped along the fault, high resolution GPR should be carried out on the hanging-wall beneath the potential sampling site. If cemented colluvium is present another sample site should be chosen. Terrestrial Light Detection and Ranging (t-LiDAR) can then be used to scan the bedrock fault plane and identify areas where the surface roughness changes rapidly. Furthermore, the t-LiDAR backscattered signal provides additional information on the surface conditions (e.g. surface colour, lineation, Riedel shears,
karstification, brecciation, lichen growth). The backscatter signal, therefore, reveals the combined surface conditions on the fault plane, which can identify different ribbons of weathering (Wiatr et al., 2011). As roughness and differential weathering relate to exposure time, it is likely that palaeoeartquakes are the cause. This information can then be used to identify the best locations for $^{36}$Cl carbonate bedrock and REE sampling when anthropogenic influences (e.g. quarrying and terracing) can be excluded. Schlagenauf et al. (2010) and Manighetti et al. (2010) provide methodologies for $^{36}$Cl and REE dating respectively. For dating other bedrock lithologies, other cosmogenic nuclides like $^{10}$Be, $^{21}$Ne or $^{26}$Al should be used (Ivy-Ochs and Kober, 2008).

Figure 2.7. Flow chart summarising the general palaeoseismological investigation procedure for normal faults.
If wanting to undertake a trenching investigation, geophysical reconnaissance needs to be undertaken on the hanging-wall to aid trench site selection. GPR and ERT should be used to characterise the spatial variability of sedimentary structures, and determine whether the sediments are appropriately faulted. Geophysical measurements are essential for the selection of trenching sites, which will be chosen to expose the longest range of temporal deformation history. The trench should be excavated perpendicular to the fault’s strike with one end up against the bedrock scarp. This will allow the smallest wedges or fissures to be visualised. Alternatively, if geophysical results show that there are secondary fault planes in the hanging-wall as shown in figures 2.2 and 2.5, a trench should also be excavated at this location. McCalpin (2009) describes further trenching techniques. After trench logging and photography, t-LiDAR can be used to scan the trench walls in order to characterise small scale sedimentary changes which are hard to visualise with the naked eye. Furthermore, non-visible physical measurements such as magnetic susceptibility can be undertaken on the trench walls, which will provide additional quantitative information on the nature and distribution of exposed sediments. t-LiDAR can also be used to scan the terrain surface surrounding the trench so that a Digital Terrain Model (DTM) can be produced for each trench site. The strategic sampling of suitable materials should then be undertaken. In order to do this event horizons need to be identified, which are the ground surface at the time of each earthquake. Samples from above and below event horizons should be taken to constrain the minimum and maximum age respectively. The ages of the samples will then be determined through radiocarbon dating and the age of the earthquakes are determined by calibrating the ages for atmospheric $^{14}$C (Reimer et al., 2013). There is no guarantee that there will be $^{14}$C rich materials in the samples taken. It is, therefore, suggested that samples should also be taken for Optically Stimulated Luminescence (OSL) dating in order to mitigate the risk of not being able to date event horizons. This technique involves dating the last time a sample containing quartz or feldspar was last exposed to sunlight. The utmost care needs to be taken when sampling so that the sample is not directly or indirectly exposed to sunlight; therefore, sampling at night time or using shielded steel tubes is recommended. More information on $^{14}$C, OSL and other Quaternary dating methods is provided by Walker (2005).

If wanting to look back further in time into palaeoenvironmental conditions during fault evolution, cementation of hanging-wall colluvium can provide valuable information. The stable isotopes of carbon ($^{13}$C and $^{12}$C) and oxygen ($^{16}$O and $^{18}$O) can be analysed to determine whether the supersaturated water from which the cemented precipitated had a meteoric or thermal origin. Palaeotemperature calculations of the parent water from which the cement precipitated can also be undertaken allowing the palaeoclimate to be accurately determined. Absolute dating techniques such as U-Th should also be applied.
2.5 Study area selection

The scope of this project was to undertake multi-disciplinary investigations on normal faults where little or no research has been done to date, covering, where appropriate, neotectonics, palaeoseismology and palaeoenvironment. The two main study sites (western Peloponnese and eastern Crete) were chosen for different reasons, which are summarised in the following sections.

2.5.1 Study Area 1: The Lapithas Mountain, Western Peloponnese

Study area 1 in the Western Peloponnese comprises the Lapithas Mountain which is a horst bounded to the north and south by two parallel E-W trending faults. Little research had been undertaken at these faults; only geological mapping of the general region (Lekkas et al., 1993), and also estimations of long term throw rates based on hanging-wall sediment thicknesses and offshore seismic data (Papanikolaou et al., 2007). What is of specific interest here is the proximity of these faults to nearby archaeological sites which have a lot of damage. One of these archaeological sites known as Samicum is only a few hundred meters from the northern Lapithas fault; and the other is Ancient Olympia, a UNESCO world heritage site which is widely reported to have destroyed by earthquakes, which is only around 10 km north of the Lapithas Mountain. Field reconnaissance showed that the faults themselves contain kinematic indicators. Therefore an investigation was undertaken in 2012 documenting archaeological damage and studying historical reports. Using fault plane structural measurements, fault lengths were accurately mapped and expected magnitudes determined through empirical calculations.

2.5.2 Study Area 2: The Lastros-Sfaka Graben, Eastern Crete

Study area 2 in Eastern Crete comprises the Lastros-Sfaka Graben which is situated within the Ierapetra Fault Zone (see Fig. 4.1). Two publications (Caputo et al. 2006; Caputo et al. 2010) provide broad summaries of the active normal faults throughout the Island of Crete. Both of these publications use scarp heights to determine postglacial slip rates, and fault lengths to determine maximum expected magnitudes. Two of the faults with the highest slip rates were reported to be the Lastros fault and the Sfaka fault with slip rates of 1.3 and 1.0 mm/a respectively (Caputo et al. 2010). During a reconnaissance trip in 2013 we visited the majority of the bedrock faults on Crete, and the Lastros and Sfaka faults were two of the best examples of normal faulting on the island. Furthermore, the Lastros fault contained good examples of hanging-wall colluvium which had become cemented and attached to the fault plane. This travertine-like material, which was observed on many Cretan faults, would therefore provide good data to study how the Lastros fault has evolved through time looking
at the palaeoenvironment of cement formation. Therefore, a multi-method investigation was undertaken over two field campaigns in 2013 and 2014. Methods including t-LiDAR, trenching (road cuts), GPR and laboratory analyses were carried out to study and understand these faults in detail.
Chapter 3: The Lapithas Mountain faults and nearby archaeological damage, western Peloponnese, Greece.


3.1 Abstract

Fault slip data are analysed from the E-W striking northern and southern Lapithas Mountain faults, which respectively form the northern and southern boundary between the Alpine mountain horst and the Quaternary basins. Kinematic indicators such as corrugations and striations show that slip directions vary on both these individual faults, which is typical for the Aegean region. Slip directions converge towards the centre of the fault where there is pure dip-slip movement; towards the lateral tips the slip is more oblique with a strike-slip component. These fault slip orientations and calculations of along strike variations in observable throw, are used in combination with published earthquake and offshore bathymetry data to accurately determine the lengths of both the northern and southern Lapithas faults. The northern Lapithas fault comprises two segments; a ca. 18 km long onshore segment and a ca. 15 km long offshore segment. Based on empirical calculations and when viewed separately, potential maximum earthquake magnitudes (M_w) of 6.5 and 6.4 could occur on the onshore and offshore segments respectively. As a worst case scenario, these two segments could both rupture during an earthquake. A multi-segment rupture totalling 33 km in length is capable of producing a maximum magnitude (M_w) of 6.9. The total length of the southern Lapithas fault is ca. 5 km and could therefore produce a maximum potential magnitude (M_w) of 5.8. There are archaeological sites in the vicinity of the Lapithas Mountain which show evidence of possible earthquake damage. Samicum, which is located on the western end of the Lapithas Mountain, contains collapsed towers and displaced walls and blocks; so far this damage has not been attributed to earthquakes, but because of the site’s location, the Lapithas Mountain faults are a likely source. Ancient Olympia, which is located 10 km to the north, has oriented fallen columns which have been attributed to earthquakes in AD 522 and/or AD 551. However, a review of the literature and earthquake records shows that no earthquakes in these years can explain the damage; further work is need to determine
whether the damage was caused by earthquakes and, if so, to find the causative fault and accurately date the damage.

3.2 Introduction

The majority of the normal faults throughout the Peloponnese comprise Mesozoic carbonate fault scarps which are juxtaposed against Quaternary marine/colluvial sediments and occasionally flysch deposits. These faults are easy to recognise as they offset smooth mountain slopes and have steeply dipping fault scarps that are several metres in height containing slickenside kinematic indicators. These preserved fault scarps are coseismic and result from cumulative slip during earthquake events on the individual fault (Benedetti et al., 2002). In the Mediterranean region the common theory is that during glacial conditions the erosion rate of these bedrock fault scarps, and sediment deposition on the hanging-wall, was faster than the fault's slip-rate; the preservation of the bedrock fault scarp within the landscape was therefore not possible (Benedetti et al., 2002; Roberts and Michetti 2004; Papanikolaou et al., 2005). In postglacial times, however, the improved climatic conditions reduced erosion rates, and vegetation stabilised the mountain slopes producing a thin organic rich soil which allowed fault scarps, caused by recurrent earthquakes, to be preserved. Fault scarps can only be preserved when the slip rate, which is usually lower than 1 mm/yr for bedrock faults in the Mediterranean region (Roberts and Michetti 2004; Papanikolaou et al., 2005), is higher than the erosion rate (Benedetti et al., 2002). This is the situation we have throughout the Peloponnese and the Aegean, and the preserved faults are considered to be capable; a capable fault is defined as a fault that has significant potential to cause displacement at or near the ground surface (IAEA SSG-9, 2010) and the preservation of many several-metre high scarps implies that regular seismic activity must have been occurring for a significant period since the Last Glacial Maximum (LGM) before the onset of the Holocene.

Coseismic fault planes contain preserved slickenside kinematic indicators which express the local conditions of faulting in their texture, type and composition; the striations on slickenside planes indicate the local displacement direction and the shear sense of the fault can be deduced. These are basic requirements in brittle tectonics analysis (e.g. Petit, 1987). The kinematic indicators found on bedrock fault planes can take many forms including grooves, crescentic markings, corrugations, steps, fractures, tool marks, etc. (for a full description of the variety of kinematic indicators see Hancock and Barka, 1987). The preservation of these indicators is, however, highly dependent on the time that has passed since exposure, and also the local conditions of the bedrock fault plane (Fig. 1a). For example if the footwall
mountain above the scarp has a morphological expression which allows rainfall to be channelled so it flows over the scarp at a catchment, kinematic indicators are quickly eroded away due to dissolution of the carbonate scarp. However, where anthropogenic activity, such as terrace construction or quarry excavation, has artificially exposed part the fault plane relatively recently, kinematic indicators can be extremely well preserved (Fig. 1a). As first noted by Roberts (1996), kinematic indicators on Aegean normal fault planes converge towards the centre of the fault where there is the largest scarp height and throw; here pure dip-slip movement is observed. Towards the ends of the fault where the scarp height and throw is much smaller, the kinematic indicators have a strike-slip component (Fig. 1b). Therefore, structural analysis of Aegean normal faults using kinematic indicators is a reliable way to determine the neotectonic evolution of an area (e.g. Roberts and Ganas, 2000; Reicherter and Peters, 2005) as long as it is ensured that entire faults and fault segments are mapped; the orientation of extension or shortening, and variations in these processes, can then be quickly determined. Moreover, since this converging slip direction pattern varies with throw and distance, it has been used in several settings among other methods to help define fault lengths (Roberts and Michetti 2004; Papanikolaou and Roberts 2007).

For this paper two normal faults located at the Kyparissiakos Gulf in the western Peloponnese were investigated. These two faults are major structures which bound the Lapithas Mountain horst (see section 2.1). To help determine the capability of these faults, and provide information regarding their formation and associated seismic hazard, structural analyses have been undertaken using identified kinematic indicators, and estimations of observable throw have been undertaken using elevation information.

The study area is of historical importance due to the presence of archaeological sites where damage may have been caused by earthquake activity. The closest site is Samicum located on the Lapithas horst itself only a few hundred metres south from the northern Lapithas fault. Ancient Olympia is located only 10 km to the north of the mountain and is a UNESCO world heritage site which is reported to have been heavily damaged by earthquakes. Studying the effects of past earthquakes from archaeological remains and evidence in the geo-archaeological record is known as archaeoseismology and involves classifying and describing earthquake damage in ancient buildings known as EAE (Earthquake Archaeological Effects) (Rodríguez-Pascua et al., 2011). These effects can be ‘primary’ (direct geological or building fabric effects) or ‘secondary’ (indirect effects such as reconstruction or repair work). For a full description of the various types of EAE see Rodríguez-Pascua et al. (2011). A review of observed and reported damage at Samicum and Ancient Olympia is presented.
3.3 Geological and tectonic setting

The orogenic belt of the Hellenides dominates the structure of Greece and has long been recognised to comprise numerous distinct sedimentary facies belts known as “tectonostratigraphic terranes” which have a NNW-SSE orientation (Aubouin, 1959; Papanikolaou, 1997) (Fig. 2a). It is now understood that these linear terranes are thrust...
sheets developed as a result of oceanic closure (Papanikolaou, 1997). Westernmost mainland Greece, the northwest Peloponnese and the Ionian islands are all located in the Ionian unit, which is a classic example of a thin-skinned linear fold and thrust belt (Broadley et al., 2006) forming part of the external carbonate platform (H1, Fig. 2a). The study area lies within the region of Elis and northern Messinia and is located at the centre of the Kyparissiakos Gulf (Fig. 2b). The Kyparissiakos Gulf area comprises parts of the Ionian unit and the Gavrovo-Tripolitza unit (both H1, Fig. 2a), and also the Pindos unit (H2, Fig. 2a). These units are all displaced by a series of E-W orientated faults which bound horst and graben structures; the northernmost horst is the Lapithas Mountain (see section 2.1) where our study is undertaken. These E-W neotectonic faults have been extensively studied (Lekkas et al., 2000; Papanikolaou et al., 2007; Fountoulis and Mariolakos, 2008); however, up to now fault-slip data have not been used to structurally analyse the bedrock faults and precise shear senses are unknown.

The Aegean region can be broadly split into three domains with different neotectonic trends (Mariolakos and Papanikolaou, 1981, 1987) (Fig. 2c). The Lapithas Mountain is located in the first domain (Domain I) which has an E-W neotectonic trend. Here, north-south extension of the upper crust is occurring similar to that observed in the Corinth rift and central Greece (Mariolakos and Papanikolaou, 1981; Armijo et al., 1996; Roberts, 1996; Papanikolaou et al., 2007). In southern Peloponnese and western Crete (Domain II), roughly NW-SE normal faulting is causing northeast-southwest extension; and in eastern Crete and the eastern Aegean islands, SW-NE faulting prevails (Domain III). There are a number of theories regarding the formation of these domains and how they interact with the Hellenic trench (e.g. Armijo et al., 1992; Papanikolaou and Royden 2007). For Domain I, where the study area is located, Papanikolaou and Royden (2007) state that E-W normal to oblique-slip normal faults can be formed by the transtensional stressfield produced by the development of the Central Hellenic Shear Zone which has varying convergence rates. It is, however, still unclear how these domains interact; furthermore, it is still unknown whether a particular domain is more active than the other. Regarding seismicity, however, the whole of the Aegean region is classified as having a high hazard. The study area has a 10% probability of exceeding 0.4 g ground acceleration in 50 years (Giardini et al., 2013).

The last destructive earthquake in the study area occurred on 23rd March 1993 (Koukouvelas et al., 1996) with an epicentre 2 km southeast of Pirgos (Fig. 2b). This earthquake had a magnitude (Mw) of 5.5 and maximum intensities of EEE (Earthquake Environmental Effects - Papanikolaou et al., 2009) were VIII and included landslides, liquefaction, ground ruptures and changes in aquifer level (Lekkas et al., 2000). Another earthquake located approximately 30 km to the northwest of Pirgos occurred in June 2008 in SW Achaia. This earthquake had
a magnitude ($M_w$) of 6.4 and occurred at approximately 22 – 25 km depth with clearly defined strike-slip movement determined from the fault plane solution (Koukouvelas et al., 2010). Associated with the strike-slip fault movement was secondary normal faulting which accommodated the diffuse deformation. The vertical offsets of the secondary normal fault rupture segments ranged from 5 cm to 25 cm (Koukouvelas et al., 2010; Mavroulis et al., 2010). Environmental effects including landslides and rockfalls, surface fractures, and liquefaction were observed (Mavroulis et al., 2010; 2013).

Figure 3.2. a) Simplified tectonostratigraphic terrane map of the Hellenides; H1: External Carbonate Platform, H2: Pindos/Cyclades oceanic basin, H3: Internal Carbonate Platform, H4: Vadar oceanic basin, H5: Paikon-Lesvos Carbonate Platform, H6: Levos-Circum Rhodope oceanic basin, H7: Pangeon Carbonate Platform, H8: Volvi – Eastern Rhodope oceanic basin, H9: Vertiskos/Sidironero basement; continental terranes are drifted Gondwana fragments with Mesozoic carbonate platforms; oceanic terranes are oceanic basins with ophiolites sutured and obducted during the Jurassic-Oligocene. b) Simplified neotectonic map of the Peloponnese. c) Structural sketch of the Aegean region showing plate boundaries and the distinction of three domains; NAFZ: North Anatolian Fault Zone (modified after Papanikolaou, 1997; Papanikolaou et al., 2007).
A paleoseismological study has been carried out on one of the secondary normal fault rupture segments known as the Nisi Fault (Zygouri et al., 2011; Zygouri et al., 2015). The authors determined a slip rate in the order of 1 mm/yr and a recurrence interval ranging between 300 and 600 years for magnitude 6 earthquakes. To date, this is the only paleoseismological study that has been carried out in the region and more studies are needed to comprehensively determine slip rates and recurrence intervals for other active faults.

The Hellenic trench which lies approximately 70 – 80 km off the western coast is the source of a lot of the seismic events that occur in Greece; these subduction earthquakes can be the cause of tsunamis. Historical reports and tsunami research have identified a minimum of four tsunami events that have struck the Gulf of Kyparissia over the last 300 years (Koster et al., 2015) and many other tsunamites have been dated from around 4000 BC to AD 1400 (Vött et al., 2011; Hadler et al., 2015; Willershäuser et al., 2015a,b). Therefore, the study area is not only prone to seismic hazard, but to the hazard posed by tsunamis along the western Peloponnese coast.

3.3.1 The Lapithas Mountain

This study focuses on the Lapithas Mountain which, as previously stated, forms the northernmost horst in a series of fault bounded horst and grabens structures orientated approximately E-W along the Kyparissiakos Gulf (Fig. 2b). The Lapithas Mountain is approximately 17 km in length. The western 13 km of the mountain has the largest relief and comprises Eocene sediments of the Gavrovo-Tripolis unit (H1, Fig. 2a) and these sediments can be divided into two main formations; netritic carbonate rocks in the west and flysch deposits of alternating sandstones and shales in the east (Lekkas et al., 1993). The mountain’s easternmost 4 km comprises pelagic carbonate rocks of the Pindos unit (H2, Fig. 2a) which are Late Maastrichtian – Eocene in age (Lekkas et al., 1993). The mountain is bounded by two main faults; the southern fault bounds the mountain to the Zacharo basin and the northern fault bounds the mountain to the larger Olympia basin, both of which contain marine sediments of Quaternary and Neogene age. Papanikolaou et al. (2007) carried out a neotectonic study at the Kyparissiakos Gulf and estimated Quaternary throw rates for some of the E-W trending faults. This was based on the topographic difference between the top of the Alpine formations comprising the horsts and their unconformity below the Quaternary marine sediments in the grabens. The authors estimate that the northern Lapithas fault has a throw rate of over 1.0 mm/yr. This is a significant throw rate and
indicates that the northern Lapithas fault is a major capable fault and poses a high risk in terms of seismic hazard.

### 3.4 Methods and Results

#### 3.4.1 Fault investigations

In order to obtain fault data for structural analysis, geological mapping was undertaken on both the northern and southern Lapithas faults in 2012. Data collection involved traversing the fault along strike and tracking its location with hand-held GPS. Where the fault plane was exposed dip and strike measurements were taken and, where present, the plunge and azimuth of kinematic indicators. These measurements were all taken with a compass clinometer with an accuracy of approximately ± 2 degrees. The best preserved kinematic indicators were found on the southern Lapithas fault, especially where the fault plane had been exposed by anthropogenic activity such as terracing or road construction. As shown in Figure 3, striations several metres in length are exposed and clear corrugations can be seen. Where there has been no anthropogenic activity, kinematic indicators are harder to identify; erosion of the carbonate fault plane removes striations (Fig. 1a), and the channeling of water runoff caused by small catchments leads to grooves and gullies which can mask the corrugations. Another indication that the fault plane has been anthropogenically exposed is lichen. Where the fault plane has been exposed by natural processes, enough time has passed for lichen to grow and colonise on the exposed bedrock; however, for exposed planes which were exhumed recently, the lichen has not had a chance to develop (Fig. 3b). Moreover, there has been no recent historical or instrumental earthquake that can explain the exposure of very fresh fault scarps.

Processing of the fault kinematic data including visualisation and analysis was carried out using the program *TectonicsFP* by Reiter and Acs (2014). The raw data were entered into the software which is then corrected so that all kinematic indicators lie perfectly on the respective fault plane and there is no misfit; after this the orientations of the stress axes were calculated using a τ angle of 30°, where τ is the angle between the compressional and extension axis which both lie on the fault plane and are defined by the orientation of fault plane and kinematic indicators (Marrett and Allmendinger 1990). Separation of fault populations belonging to distinct stress tensors was then undertaken.
Figure 3.3. Photoplate showing the southern Lapithas fault and kinematic indicators: a) view of the southern fault's western 2km where limestone bedrock is juxtaposed against Quaternary sediments, yellow arrows indicate fault scarp location and the locations of photos b and c are marked; b) ca. 8 m high fault plane at location S3 (inset is the Angelier plot of fault-slip data), the lower three metres have been exposed by anthropogenic activity – note the dashed red line below which no lichen is growing and the paleoground surface to the east; c) iron coated striae with polished surfaces; d) asperity ploughing and iron coated striae located in a quarry near the fault’s eastern end.
In addition, a calculation of the observable throw was carried out in GIS based on SRTM elevation data. This was done by comparing the elevation of each fault plane (northern and southern faults) over their mapped lengths with the highest elevation of footwall topography or apex of the mountain which is the main drainage divide; it should be noted that the actual throw measured from the top of the mountain horst to the unconformity below the Quaternary hanging-wall sediments is considerably higher and the observable throw must be considered as only a proxy for actual throw. Furthermore, this analysis does not take into account differential erosion which may be occurring on the footwall mountain. Individual profiles were then created every 100 m between the mountain’s apex and each fault using the average dip of the fault for profile orientation. For the southern and northern faults an average dip direction of 175° and 343° were used respectively.

Figure 3.4. Updated neotectonic map of the Lapithas Mountain and surrounding area based on Lekkas et al. (1993). Locations where fault-slip data were taken are shown along with corresponding Angeller plots. Stress axes of locations grouped according to their stress tensor are also presented. Note the ancient archaeological site Samicum located on the mountain’s western end.
3.4.1.1 The Southern Lapithas Fault

The southern fault was mapped for approximately 4.5 km where the fault comprises the carbonate footwall of the Alpine basement juxtaposed against the hanging-wall Quaternary marine sediments of the Zacharo basin. Five locations along the main fault were encountered where reliable kinematic indicator measurements could be undertaken (Fig. 4 and Table 1). The kinematic indicators recorded mainly comprised decimetre to metre scale corrugations, and several mm to cm wide ploughing striations (Fig. 3); other indicators including crescentic markings, iron coated striations and tool marks were also recorded.

Figure 4 is a revised neotectonic map of the Lapithas Mountain showing Angelier plots (displaying the fault plane as great circles with the arrows pointing in the direction of the relative slip of the hanging-wall) of the fault plane and kinematic indicators at each location. The Angelier plots from locations S1 – S5 show that the kinematic indicator orientations are converging and vary according to distance along the fault; fault-slip directions vary by ca. 18° over 3 km.

Between locations S2 and S3, the fault-slip directions converge which indicates that the centre of the fault is located between these two locations. Further evidence for this can be seen from Figure 5a where the observable throw has been calculated. Here, the highest observable throw, which should be at the centre of the fault (Roberts, 1996), is located between S2 and S3 at a distance of 1,600 m along the profile. This, therefore, indicates that the southern fault is most likely only around 5 km long, because if the southern fault continued offshore to the west, the fault-slip directions of S1 and S2 should also all have an azimuth to the west. Furthermore, stress axes orientation for locations S1 – S5 (Fig. 4) appear to belong to the same stress tensor indicating that there has been no overprinting from a varying extensional stress field. Location S6 represents the boundary between the netritic carbonate rocks in the west and flysch deposits of alternating sandstones and shales in the east. These two formations have been juxtaposed against each other by a normal fault which shows a major strike-slip component.
Table 3.1. Fault plane and kinematic indicator measurements and calculated mean stress axis vectors ($\sigma_1$, $\sigma_2$ and $\sigma_3$) for each locality.

<table>
<thead>
<tr>
<th>Locality</th>
<th>Coordinates</th>
<th>Fault Plane</th>
<th>Kinematic Indicators</th>
<th>Stress Axes Orientation - Mean Vector (azimuth°/plunge°)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Lat (d:m:s)</td>
<td>Long (d:m:s)</td>
<td>dip direction (°) / dip (°)</td>
<td>azimuth (°) / plunge (°)</td>
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<tr>
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<td>21:37:52</td>
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<td>169/55</td>
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<tr>
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<td>169/56</td>
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<tr>
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<td>21:38:31</td>
<td>192/69</td>
<td>197/69</td>
</tr>
<tr>
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<td>21:39:29</td>
<td>162/55</td>
<td>174/52</td>
</tr>
<tr>
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<td>184/43</td>
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<td>343/51</td>
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<td>030/80</td>
<td>311/80</td>
</tr>
</tbody>
</table>

Chapter 3: The Lapithas Mountain Faults
3.4.1.2 The Northern Lapithas Fault

The northern fault was mapped for 12.5 km where the fault comprises the carbonate and clastic footwall of the Alpine basement juxtaposed against the hanging-wall Quaternary marine sediments of the Olympia basin. In the east where the hanging-wall is comprised of the Neogene Platiana formation (Fig. 4), the fault scarp is no longer visible. However, due to a road cut at location N6 the fault’s location was proven in the east and a total length of 18 km can be inferred. There has been less anthropogenic activity here compared to the southern fault and this has resulted in relatively poor exposure of kinematic indicators (striations). Five locations were encountered along the main boundary fault (locations N1, N2, N3, N5 and N6) where reliable kinematic indicators were present and comprised decimetre to metre scale corrugations, tool marks and several mm wide ploughing striations. Figure 4 shows that at locations N2, N3 and N5, the slip-directions are all close to the average dip of the fault and only the pitches of the lineations vary. This indicates that the centre of the fault is located close to these locations (Fig. 5b); also, locations N2, N3 and N5 have the same stress axes orientation indicating that they belong to the same stress tensor. At location N1 and N6, the normal movement has a strike-slip component and the stress axes plots shows different orientations (Fig. 4) compared to N2, N3 and N5. At N1 this is partly due to the strike of the fault plane at this location as the slip directions are similar for all locations (N1, N2, N3 and N5) and only the pitches of the lineations vary. The increasing strike-slip component with increasing distance from the fault centre (Fig. 1b) in shown in Figure 5b, where the azimuth of the kinematic indicators is plotted in map view for each location. Location N6 has a large strike-slip component indicating that it is close to the fault’s eastern tip. Further evidence showing the location of the fault’s centre can be seen from the plot of observable throw (Fig. 5b); the highest values occur at 7,300 and 8,700 m along the profile which complements the near pure dip-slip kinematic data at location N5.
Figure 3.5. Elevation profiles along the main southern (a) and northern (b) Lapithas faults showing the variation in footwall topography; red arrows indicate fault displacement directions in map view. Top image shows the locations where 100 m wide polygons were used to calculate the observable throw. Elevation profiles were constructed using SRTM data.
Chapter 3: The Lapithas Mountain Faults

At location N4, a near perpendicular fault was encountered striking approximately N-S (Fig. 4 – inactive fault). Although no cross cutting relationships were observed in the field between this N-S fault and the main boundary fault, the perpendicular N-S fault is interpreted as being older than the main fault. This is due to there being no sudden change in main fault elevation close to location N4 (Fig. 5b). The perpendicular fault represents the eastern flank of a valley that can be traced to the top of the mountain. This is clearly shown in Figure 5b where the elevation of the apex of the mountain begins to fall sharply at approximately 7,300 m along the profile. Another interpretation of this structure is that it is a release fault, and therefore not older than the main Lapithas northern fault. Although release faults, which strike parallel to the main extension direction, usually begin at the main fault scarp and extend into the hanging-wall (Destro, 1995), their presence within the footwall could accommodate the observed throw variability along strike.

3.4.2 Local earthquake records

As previously mentioned, the western Peloponnese is structurally complex due to its proximity to the Hellenic Arc and the transition between structural domains I and II (Fig. 2c) and this complexity can be seen in recent earthquake activity. Fault plane solutions from recent earthquakes in the western Peloponnese and Hellenic arc where structural data is available from the University of Athens Seismological Catalogues (UASC, 2008) are presented in Figure 6a and visualised using the program FaultKin 5.5 by Allmendinger (2012). The focal mechanisms show a variety of normal, thrust and strike-slip earthquakes having been recorded since 1986. In order to determine whether any instrumentally recorded earthquakes can be attributed to the Lapithas faults, over 110 years’ worth of earthquake data has been compiled from the (UASC, 2008) and Makropoulos et al. (2012). Greece has one of the longest historical earthquake catalogues worldwide with the oldest recorded events in 550 B.C. However, this catalogue is only considered complete for events M≥7.3 since 1500 and for events M≥6.5 since 1845 (Papazachos et al., 2000). The location accuracy of pre-instrumental earthquakes is too poor for our requirements. From 1900 to 2012, over 170 Mw 4 and above earthquakes have been recorded within 50 km of the Lapithas Mountain, of which approximately 100 are shown in Figure 6b; these include the previously mentioned 1993 Pirgos earthquake and the June 2008 SW Achaia earthquake as well as their aftershock activity. The largest earthquake that occurred close to the northern Lapithas fault was a normal faulting earthquake which occurred in 1987 with a magnitude of 5.4, located approximately 2 km north of Samicum (see fault plane solution in Fig. 6b).
Figure 3.6. a) Fault plane solutions for selected earthquakes over $M_w$ 5.5 from 1985 to 2008 plotted on Google Earth satellite imagery. Focal mechanism data are from The University of Athens Seismological Catalogues (UASC); diameters of focal mechanisms are scaled for magnitude. b) Earthquakes over $M_w$ 4 which have occurred close to the Lapithas Mountain from 1900 to 2012, visualised using GeoMapApp (http://www.geomapapp.org/). Earthquake data taken from Makropoulos et al. (2012) and the University of Athens Seismological Catalogues (UASC); i = M 4.1 on 15/02/2003 at 20 km depth, ii = M 4.1 on 28/04/1982 at 15 km depth, and iii = M 4.3 on 30/04/1982 at 26 km depth. Note: earthquakes have been scaled for magnitude and colours represent different depths.
However, this earthquake occurred at a considerable depth of 57 km, and therefore was associated with the subducting slab of the Hellenic trench and not shallow extension. There are, however, three earthquakes that occurred offshore to the west of the Lapithas Mountain. These three earthquakes (i, ii and iii in Fig. 6b) had magnitudes between 4.1 and 4.3, and occurred at depths between 15 and 26 km. Therefore, these are shallow enough to be attributed to shallow crustal extension and due to their geographical location are indicators that the northern Lapithas fault may still be active. Unfortunately, no structural data is provided for these earthquakes, and therefore their focal mechanisms are unknown. Overall, when considering the recurrence intervals of these faults and the completeness period of the historical record, the historical catalogues are generally too short. Therefore, paleoseismological studies that extend the history of fault slip back many thousands of years are necessary (Yeats and Prentice 1996; Grützner et al., 2013).

3.4.3 Archaeoseismology in the vicinity of the Lapithas Mountain

The Elis province is most famous for the archaeological site of Ancient Olympia. This site is located ca. 10 km to the north of the Lapithas Mountain and is birthplace of the Olympic Games, which were held there for over 1100 years; many authors have attributed damage at the site to earthquakes. Samicum is located at the western end of the Lapithas Mountain Ridge (Figs. 2b and 4) and also has experienced damage which may be attributed to earthquakes. Here we describe the damage from these two archaeological sites and summarise historical reports related to the damage.

3.4.3.1 Samicum

The fortified village of Samicum located on the western end of the Lapithas Mountain Ridge was most probably founded at the end of the 5th cent. BC (classical Hellenistic period) and was occupied until the 2nd cent. AD (Roman period) (Pausanias, AD 115-180). Samicum’s most prominent feature is the so-called pseudo-polygonal wall c. 1500 m in length that surrounds the village (Fig. 7a). The wall is trapezoid in shape forming an acropolis and is characterised by several collapsed towers (Fig. 7b/c).

Indicators of seismic damage include broken, moved and rotated masonry blocks (Fig. 7d), corner break outs, missing triangular masonry blocks, (Fig. 7e) and the collapsed towers. Some towers have collapsed into the village whereas some towers have collapsed to the outside onto steep slopes. If these towers were destroyed by anthropogenic activity they should all have collapsed to the inside; the steep slopes surrounding the village would
significantly restrict the ability of humans to pull these towers down. Furthermore, the village was not totally destroyed and many sections of the wall remaining largely intact. This differential destruction is commonly observed in structures which have undergone seismic shaking.

Up to now this damage has not been described as related to earthquakes; there is no historical account for earthquake damage at Samicum. Pausanias (AD 115-180; Book 5, Chapter 6, Sections 1-3) refers to his journey to Samicum in the 2nd cent. AD when he found the village already in ruins indicating that the destruction and abandonment was pre 2nd cent. AD. Due to Samicum’s proximity to the northern and southern Lapithas faults, it is highly likely that the observed damage at the site can be attributed to one of these faults.

3.4.3.2 Ancient Olympia

Ancient Olympia is a UNESCO World Heritage site known as the location of the classical Olympic Games which were staged here from 776 BC to AD 393. The sanctuary or Altis resembles a variety of buildings of different use and age during more than 3000 years (from c. 2500 BC until 6th cent. AD). The area was then flooded and buried by 6 m of sediments from the nearby Kladeos Creek and Alfeios River (Fouache and Pavlopoulos, 2010) and was forgotten until AD 1829; excavation of the site then began in AD 1875 by the German Archaeological Institute. Two of the most famous temples in the Altis are the Philippeion Temple and the Temple of Zeus; the latter is widely reported to have been destroyed by earthquakes (see below). Many earthquake archaeological effects (EAE) can be observed throughout the site including orientated fallen columns (Fig. 7f) (Stiros and Jones, 1996), dropped keystones and dipping broken corners (Fig. 7g/h).

Dinsmoor (1941) carried out an archaeological investigation at the site and discovered two different styles of column drums, on which the connection to the adjacent drum can be seen. Of the 135 drums which could be accurately classified, 30 contained Lewis holes and 105 contained only a central hole. Through these replacements, repair work such as clamps at the western corners, and the study of ancient texts, Dinsmoor (1941) was first to suggest that an earthquake must have occurred soon before 169 BC.

It is widely accepted, but not proven, that the Temple of Zeus at Ancient Olympia was destroyed by earthquakes in AD 522 and/or AD 551. Many textbooks on ancient Greece, archaeology, and even archaeoseismology (see Stiros and Jones, 1996) all state this fact,
with nearly all presenting photographs of the Temple of Zeus’ fallen columns (Fig. 7f) as typical examples of earthquake archaeological effects (EAE). However, through the study of historical earthquake documentation, the timing of the destruction of the Temple of Zeus is not that clear-cut. The AD 522 event occurred near Corinth and caused serious damage to the city which was then rebuilt with imperial aid (Ambraseys, 2009). The UASC (2008) estimates that the earthquake had a magnitude of 6.7 and actually dates this event to have occurred in AD 521. Papazachos and Papazachou (1997) also state that this event occurred in AD 521. There is, however, no question that these sources are all referring to the same event as the preceding and succeeding earthquakes in Greece occurred in AD 505 and AD 541 respectively. In AD 551 three large earthquakes occurred in Greece; for locations of damaged/destroyed cities see Fig. 2b: (1) a destructive earthquake in Boeotia that destroyed Chaeronea, probably damaging Corinth; (2) an earthquake and associated tsunami in the Maliac Gulf; and (3) an earthquake in the Gulf of Corinth heavily damaging both Nafpaktos and Patras (Ambraseys, 2009; Papazachos and Papazachou, 1997). Some authors such as Mallet and Mallet (1858), Guidoboni (1989) and Guidoboni et al. (1994) amalgamate these three earthquakes into one huge earthquake but this possibility can be omitted on physical grounds; many important centers would have been destroyed such as Chalkis and Delphi and in fact they escaped damage (Ambraseys, 2009). Moreover, there is no such fault structure with the dimensions and the orientation that can cause such damage in such different tectonic sub-settings. The UASC (2008) also amalgamate these events as there is only one earthquake recorded in AD 551 which had an estimated magnitude of 7.1. The preceding and succeeding earthquakes are both recorded as occurring in Turkey in AD 543 and AD 553 respectively. Ancient Olympia is over 65 km south of Patras (closest AD 551 event) and over 115 km from Corinth (AD 522 event) (Fig. 2c and 6a). These events would have had to be huge earthquakes to have caused the ground acceleration needed to cause the observed damage at Ancient Olympia. The region around Ancient Olympia is characterised by a large number of neotectonic faults (Lekkas et al., 1993; Lekkas et al., 2000; Papanikolaou et al., 2007; Fountoulis and Mariolakos, 2008) such as those at the Lapithas Mountain located only 10 km to the south of the archaeological site. It is far more likely that a paleoearthquake on one of the more nearby neotectonic faults would have caused the observed damage.
Figure 3.7. Photoplate showing features and Earthquake Archaeological Effects at Samicum (a-e) and Ancient Olympia (f-h): a) Google Earth imagery aerial view of Samicum showing the c. 1500 m pseudo-polygonal wall – for location see Fig. 4; b) collapsed towers of the southern wall; c) rotated wall and displaced blocks; d) collapsed wall and tower; e) missing triangular masonry blocks; f) oriented fallen columns of the Temple of Zeus; g) and h) dipping broken corners at the Philippeion Temple, some of which have been recently repaired.
3.5 Discussion

As first noted by Roberts (1996), the slip-directions vary on individual normal faults in Greece, even if the overall extension is perpendicular to the fault’s strike and faulting is mainly dip-slip. Fault slip directions converge towards the centre of the fault where the throw is greatest; pure dip-slip movement is observed at the centre of the fault, and towards the lateral tips there is a strike-slip component (Fig. 1). This has been observed on many faults in the Aegean (e.g. Roberts, 1996; Roberts and Ganas, 2000; Roberts and Koukouvelas, 1996) and the Apennines (Roberts and Michetti 2004, Papanikolaou and Roberts 2007), and now also on the Lapithas Mountain faults. During normal faulting hanging-wall subsidence exceeds footwall uplift; therefore, greater along strike extension occurs in the hanging-wall than in the footwall along segmented normal faults (Ma and Kusznir, 1995). Indeed, throw gradients on faults produce stretching of the ground surface along strike, so slip-directions at the tips of the faults are oblique and converge towards the hanging-wall to accommodate this stretching (Roberts 1996; Roberts and Ganas, 2000). This along strike variability has also been demonstrated in 3D modelling (Maniatis and Hampel 2008). This variation in slip-direction is clearly shown on both the southern and northern Lapithas faults and the convergence takes place where the observable throw is highest (Figs. 4 and 5a/b.).

On the southern fault, the variation in slip-direction and observable throw indicates that the fault’s full length is only around 5 km as the centre of the fault is located at approximately 1,600 m along the profile shown in Figure 5a. Furthermore, offshore bathymetry data (Papanikolaou et al., 2007) and seismic profiling (Wardell et al., 2014) do not show an offshore prolongation of the fault. However, based on microseismicity data in combination with more recent swath bathymetry data (Camera et al., 2014), Papoulia et al. (2014) infer a large offshore continuation of the southern Lapithas fault. Based on our variation in fault-slip data, we believe this offshore propagation of a Lapithas southern fault not to be the case, as this would require the fault to extend for many more kilometres onshore to the east, which it does not; furthermore, three months of microseismicity data may not be enough to accurately infer faults locations. The bathymetry data from Camera et al. (2014) does indicate a near-linear feature extending offshore from the Kyparissiakos Gulf. However, this is most likely an erosional feature caused by the nearby onshore catchment and not an offshore continuation of the fault. Therefore, the total length of the southern Lapithas fault is ca. 5 km. According to empirical relationships of surface rupture length vs magnitude, and surface rupture length vs maximum displacement (Wells and Coppersmith, 1994), the southern fault could produce an earthquake with a maximum magnitude (M_w) of 5.8 with a maximum displacement in the range of tens of centimetres. This is still a significant event and the hazard remains high for
those living close to the mountain. Iron coated striations were identified at a number of localities (Fig. 3). Iron coated striations are formed by iron precipitation as groundwater flows down the fault plane. Occasionally these appeared very smooth due to fault movement occurring after iron precipitation. However, no reliable slip rates can be estimated using these structures alone; further work including paleoseismological investigations are needed to determine this.

On the northern Lapithas fault, data from locations N2, N3 and N5 show that the centre of the fault is most likely around these locations as close to pure dip-slip movement is observed belonging to the same stress tensor. At locations N1 and N6, the normal movement has a strike-slip component and the stress axes plot shows different orientations (Fig. 4). At N1 this is partly due to the strike of the fault plane at this location as the slip directions are similar for all locations (N1, N2, N3 and N5) and only the pitches of the lineations vary. However, at location N6 there is a large strike-slip component and when the azimuths of the kinematic indicators are plotted along the fault at each location (Fig. 5b), the typical Aegean fault structure can be observed. Furthermore, calculations of observable throw indicate that the centre of the fault is close to location N5. As the fault is of significant length, it may extend offshore to the west. Papanikolaou et al. (2007) present a bathymetry study and lithoseismic profiles of the continental slope off the Kyparissiakos Gulf. The edge of the continental platform is a very good indicator of Holocene-postglacial deformation since it was formed during the glaciation and undergoes continuous sedimentation. Papanikolaou et al. (2007) infer that the northern Lapithas fault may continue up to ca. 15 km offshore, producing large discrepancy sliding; the authors state that the total length of the northern Lapithas fault is ca. 25 km. Furthermore, instrumentally recorded earthquake data (Fig. 7b) also indicate that the fault may continue offshore. Camera et al. (2014) present an updated bathymetry survey of the continental margin off the western Peloponnese. The bathymetry data also show evidence for a continuation of the northern Lapithas fault offshore. Seismic profiling shows sharp sub-vertical offsets cutting lithified sediments and large scale discrepancy sliding; this is most likely associated with fault activity and complements the interpretations of Papanikolaou et al. (2007). However, we interpret this offshore continuation of the northern Lapithas fault as another segment of a fault array, which are observed a lot in the Aegean (Roberts, 1996; Roberts and Ganas, 2000). This is because the northern fault’s finite throw diminishes rapidly as it approaches the coast (Fig. 5b). The Lapithas Mountain outcrops Alpine rocks at over 700 m a.s.l. only 5 km from the coast, and these rocks disappear westwards into the sea. Therefore, the offshore continuation observed in bathymetry data must be associated with another segment. The northern fault therefore comprises two segments: an 18 km long onshore segment, and a 15 km long offshore segment. When
viewed separately these onshore and offshore segments are capable of producing earthquakes with maximum magnitudes ($M_w$) of 6.5 and 6.4 respectively (Wells and Coppersmith, 1994). As a worst case scenario these two segments could interact during a seismic event (Cowie and Roberts, 2001; Roberts and Michetti, 2004). A multi-segment rupture totalling 33 km in length would be capable of producing a displacement of up to 2 metres and a magnitude ($M_w$) of 6.9 (Wells and Coppersmith, 1994).

The archaeological site known as Samicum located on the mountain’s western tip shows evidence of earthquake damage. It is highly likely that this damage can be attributed to one of the Lapithas faults due to their proximity to the ancient site; however, we cannot currently determine whether it is the northern or southern fault. Further investigations into the orientations of the observed EAE may help determine which fault is the causative one.

The Temple of Zeus at Ancient Olympia is widely reported to have been destroyed by earthquakes either in AD 522 and/or AD 551. However, earthquake records show that in these years no earthquake struck the nearby region. The archaeological evidence for either the AD 522 and/or AD 551 earthquakes having caused the collapse of the Temple of Zeus is itself rather uncertain. Ambraseys (2009) states that it is based on a theory by Boetticher (1883) who studied ancient scripts from Lucian, and unearthed coins found underneath fragments of the temple. The unearthed coins were dated from the period of Justinian II (AD 527-565) which led Boetticher to believe that the temple must have been destroyed between AD 426 and AD 565. Due to the temple’s huge size and the way the columns had fallen he decided it must have been caused by an earthquake and after studying the earthquakes that occurred during that time he concluded that is must have been the AD 522 and/or AD 551 events (Ambraseys, 2009). However, our detailed examination of these events show that this is very unlikely to be the case due to the earthquakes in AD 522 and AD 551 being of insufficient magnitude and located too far from Ancient Olympia. A lot of damage at Ancient Olympia could be attributed to earthquakes; further investigations are, however, required to characterise and catalogue all EAE and provide a clear representation of the events that destroyed the ancient site.
3.6 Conclusions

A study of fault slip data and calculations of observable throw at the northern and southern Lapithas faults provides a better estimation of the fault’s lengths and the hazard that they pose. The literature states that the northern Lapithas fault is ca. 25 km in length; however, the fault extends further inland than previously thought and most likely also extends offshore as another segment of a fault array. The onshore segment length is ca. 18 km and the offshore segment length is ca. 15 km; based on empirical calculations and when viewed separately, the onshore and offshore segments could produce maximum earthquake magnitudes (Mw) of 6.5 and 6.4 respectively. A multi-segment rupture is also possible for the northern fault. This is considered as a worst case scenario and is capable of producing a maximum earthquake magnitude (Mw) of 6.9. The hazard posed by the northern fault is, therefore, higher than previously estimated. The southern Lapithas fault does not continue offshore; the total fault length is ca. 5 km and could produce a maximum magnitude of 5.8.

The archaeological site known as Samicum located on the mountain’s western tip shows evidence of earthquake damage. The earthquake source can most likely be attributed to one of the Lapithas faults due to their proximity to the site, which is ca. 500 m and 2 km for the southern and northern faults respectively. Ancient Olympia located 10 km to the north of the mountain is a UNESCO World Heritage site which is widely reported to have been damaged by earthquakes; however the evidence for earthquake destruction is unclear and as we have shown the date of this damage is uncertain.
Chapter 4: Fault structure and deformation rates at the Lastros-Sfaka Graben, Crete


4.1 Abstract

The Lastros and Sfaka faults have an antithetic relationship and form a ca. 2 km wide graben within the Ierapetra fault zone in eastern Crete. Both faults have impressive bedrock fault scarps many metres in height which form prominent features within the landscape. t-LiDAR investigations undertaken on the Lastros fault are used to accurately determine vertical displacements along a ca. 1.3 km long scanned segment. Analyses show that previous estimations of post glacial slip rate are too high because there are many areas along strike where the scarp is exhumed by natural erosion and/or anthropogenic activity. In areas not affected by erosion there is mean scarp height of 9.4 m. This leads to a slip rate of $0.69 \pm 0.15 \text{ mm/a}$ using $15 \pm 3 \text{ ka}$ for scarp exhumation. Using empirical calculations the expected earthquake magnitudes and displacement per event are discussed based on our observations. Trenching investigations on the Sfaka fault identify different generations of fissure fills. Retrodeformation analyses and 14C dating of the fill material indicate at least four events dating back to $16055 \pm 215 \text{ cal BP}$, with the last event having occurred soon after $6102 \pm 113 \text{ cal BP}$. The Lastros fault is likely the controlling fault in the graben, and ruptures on the Lastros fault will sympathetically affect the Sfaka fault, which merges with the Lastros fault at a depth of 2.4 km. The extracted dates from the Sfaka fault fissure fills therefore either represent activity on the Lastros fault, assuming they formed coseismically, or accommodation events. Cross sections show that the finite throw is limited to around 300 m, and the derived slip rate for the Lastros fault therefore indicates that both faults are relatively young having initiated $435 \pm 120 \text{ ka}$. 
4.2 Introduction

The island of Crete is located within the active extensional regime of the Aegean (Fig. 4.1a). Many of the associated normal faults throughout the island have bedrock scarps which form prominent features within the mountainous landscape. These normal faults comprise footwall limestone bedrock scarps which are mainly juxtaposed against hanging-wall Quaternary colluvial and/or marine sediments. The faults are easy to recognise as they offset smooth mountain slopes and have steeply dipping preserved fault scarps that are many metres in height. The preserved fault scarps were generated coseismically and result from cumulative earthquake events on the individual fault plane. In the Mediterranean the common theory (Benedetti et al., 2002) is that during glacial conditions the erosion rate of these bedrock fault scarps, and sediment deposition on the hanging-wall, were faster than the fault’s slip-rate. This resulted in the bedrock fault scarp being covered and not visible in the landscape. In postglacial times, however, the climatic conditions reduced erosion rates allowing fault scarps caused by recurrent earthquakes to be preserved (Benedetti et al., 2002; Papanikolaou et al., 2005; Reicherter et al., 2011). Scarp preservation is response to the progressive variation of several natural phenomena and their role and intensity in shaping the Earth surface. These include: i) the annual amount of precipitation, and rainfall event duration and intensity; ii) vegetation type and density; iii) mean annual temperature and daily/seasonal variations. These parameters, in turn, strongly influence and govern the overall rate of erosion, the rate and size of clastic production in the footwall block, and the transport energy. Accordingly, it is likely that the growth of a cumulative fault scarp with a relatively stationary velocity (i.e. constant slip-rate) began at 15 ± 3 ka, which is some ka after the last glacial maximum. This ± 3 ka age uncertainty is due to the fact that even though the major glacial retreat in the Mediterranean began at ca. 18 ka, some small magnitude glacial re-advances followed by retreat phases have been recorded between 12 ka and 18 ka (Allen et al., 1999; Giraud & Frezzotti, 1997; Papanikolaou et al., 2013). There are over 20 known bedrock normal faults/fault segments (Caputo et al., 2010) located throughout Crete (Fig. 4.1b) which are considered to be capable and have large exposed bedrock fault scarps. These faults range from 4 to 25 km in length, some of which stand alone and some of which form segmented parts of more complex fault zones; all faults/fault segments have the potential to generate moderate to large earthquakes $M_w > 5.5$ (Wells & Coppersmith, 1994) and are associated with shallow crustal earthquakes of ~10 – 15 km depth.

This paper presents the results of a multi-disciplinary investigation on the Lastros-Sfaka Graben located within the Ierapetra fault zone (IFZ), eastern Crete (Figs. 4.1b, 4.2). The Lastros-Sfaka Graben is approximately 2 km wide and consists of two opposing faults, the
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Lastros fault and the Sfaka fault (Figs. 4.1b, 4.2), both of which strike approximately NNE – SSW and have prominent fault scarps. Methods including geological mapping, terrestrial Light Detection And Ranging (t-LiDAR), trenching (road cuts) and laboratory analyses were carried out at these faults. The results allow us to infer slip rates based on scarp heights and also bracket the dates of the most recent earthquakes, providing an insight into the neotectonic evolution of this part of the island.

4.2.1 Geological and tectonic setting

The orogenic belt of the Hellenides dominates the structure of the Aegean and has long been recognized to comprise numerous distinct sedimentary facies belts known as “isotopic zones” which have an approximate N-S orientation. It is now understood that these linear isotopic zones are thrust sheets developed as a result of oceanic closure. The island of Crete forms part of the external Hellenides (Aubouin, 1959) and comprises a number of these thrust sheets which have been successively imbricated by other sedimentary and metamorphic units (van Hinsbergen et al., 2005; Papanikolaou & Vassilakis, 2010). In the IFZ in eastern Crete the stratigraphically lowest Mani unit, also known as the Plattenkalk unit (Fig. 4.2), is the so called autochthon basement comprising crystalline limestone. This unit was imbricated by the Western Crete unit (Fig. 4.2) which comprises mainly Permo-Triassic phyllites and Middle to Late Triassic evaporites. The Tripolis unit (Fig. 4.2) was then imbricated and comprises a thick sequence of flysch, limestone, dolomite, andesite, diabase and phyllites, all of which are preserved in eastern Crete. These rocks form the volcano-sedimentary base to the shallower carbonate rocks of the Tripolis carbonate platform, exposed in both Crete and the Peloponnese (Papanikolaou & Vassilakis, 2010). The
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deformation history of these units within Crete can be summarised as: (i) compressional
deformation producing arc-parallel east–west-trending south-directed thrust faults in
Oligocene to Early Miocene time (e.g. Bonneau, 1984); (ii) extensional deformation along
arc-parallel, east–west-trending detachment faults in the Middle Miocene, with hanging-wall
motion to the north and south (Fassoulas et al., 1994; Papanikolaou and Vassilakis, 2010;
Zachariasse et al., 2011); and (iii) Late Miocene–Quaternary transtensional deformation
along high-angle normal and oblique normal faults that disrupt the older arc-parallel
structures (Papanikolaou & Vassilakis, 2010; Peterek and Schwarze 2004). The Hellenic Arc
and Trench System is located to the south of Crete (Fig. 4.1a) and the Late Miocene–
Quaternary extension is attributed to crustal back arc extension, interpreted as a response to
the southward slab-rollback of the Hellenic margin, the southwestward expulsion of the
Aegean microplate and the anticlockwise rotation of the African lithosphere relative to
Eurasia (Meukenkamp et al., 1988; Reilinger et al., 2006). The southward slab-rollback is the
predominant mechanism; from the Middle Miocene the central and southern Aegean domain
began to extend rapidly in a north-south direction implying the rapid migration of the Crete
trench relative to northern Greece (Angelier et al. 1982; Royden and Papanikolaou, 2011).
Extension is occurring orientated both arc-perpendicular and arc-parallel, which has led to a
complex pattern of normal faulting throughout the region (Fig. 4.1a) which affects the entire
pile of tectonic units (Fig. 4.2). Crete has normal faults roughly oriented both NNE – SSW
and ESE– WNW (Fig. 4.1b). Mountrakis et al. (2012) state that for northwestern Crete the
WNW-ESE trending faults are older and now inactive as they do not affect Quaternary
deposits and are overprinted by N-S trending faults. Moreover, focal mechanisms for recent
normal faulting earthquakes throughout the whole of the island (Heidbach et al., 2008) also
show predominantly E-W extension, indicating that the NNE – SSW trending faults may pose
a higher hazard.

The study area is the Lastros-Sfaka Graben located in eastern Crete (Figs. 4.1b, 4.2, 4.3).
The Lastros-Sfaka Graben forms part of the IFZ which consists of a roughly 25 km long zone
of fault segments most of which dip WNW (Fig. 4.2). The westernmost and longest is the
Ierapetra fault which likely traverses the whole width of the island. The Lastros fault dips to
the ESE and forms the ESE boundary of the Kapsos ridge (Figs. 4.2, 4.3b). The easternmost
Sfaka fault has an antithetic relationship with the Lastros fault dipping WNW. The footwall
mountains of both the Lastros and Sfaka faults comprise crystalline limestone of the Mani
unit (Papanikolaou & Vassilakis, 2010). In the south of the graben the hanging-wall of these
faults comprises limestone of the Mani Unit and Permian-Triassic phyllites of the Western
Crete unit overlain by Quaternary colluvial deposits, whereas in the north the hanging-wall
mainly comprises Permian-Triassic phyllites of the Tripolis unit overlain by Quaternary
colluvium. In many places on both faults this young colluvium has become cemented. A detailed description of the Lastros and Sfaka faults is provided in sections 2 and 3 respectively.

Figure. 4.2. Simplified geological map of the Ierapetra Fault Zone (IFZ) in eastern Crete showing the main Alpine tectono-stratigraphic units, post-Alpine deposits and active faults. The Mani unit is the so called autochthon basement comprising crystalline limestone; the Western Crete unit comprises mainly Permo-Triassic phyllites and Middle to Late Triassic evaporites; the Tripolis unit comprises flysch, limestone, dolomite, andesite, and phyllites. Extension on the active normal faults began in the Late Miocene and is still ongoing and affects the entire pile of tectonic units (modified from Papastamatiou, 1959; Papanikolaou & Vassilakis, 2010). The locations of figure 4.3a and cross sections A and B (see Fig. 4.11) are shown.
A number of investigations have been undertaken at the Lastros-Sfaka Graben. Stewart & Hancock (1990) carried out an investigation into footwall brecciation on a number of Aegean faults including the Lastros fault. The authors show that the footwall behind the scarp comprises a compact breccia sheet, an incohesive breccia sheet with chaotic arrangement of fractures, and a broader shatter zone with an orthogonal network of fractures. These different styles of breccia are attributed to different stages of fault zone evolution. Caputo et al. (2006, 2010) undertook a study of 21 individual faults/fault segments throughout Crete. The authors determined the fault lengths and estimated potential maximum magnitudes using empirical calculations (e.g. Wells & Coppersmith, 1994), determining most likely magnitudes of 6.4 and 6.3 for the Lastros and Sfaka faults, respectively (Caputo et al., 2010). They also estimated scarp heights and long-term slip rates using an assumed 13 ka (post glacial) as a starting date when the endogenic forces (i.e. scarp formation by linear morphogenic events; Caputo, 2005) began to prevail over the exogenic ones (i.e. detrital production uphill, footwall erosion and transport downhill capable of burying and dismantling the scarp during the interseismic periods). The authors calculated long-term slip rates of 1.3 mm/a and 1.0 mm/a for the Lastros and Sfaka faults, respectively. These rates are very fast compared to other bedrock faults of comparable size in the Mediterranean (Roberts & Michetti. 2004; Papanikolaou et al. 2005; Wilkinson et al., 2015), and if these are accurate, the Lastros and Sfaka faults would pose one the largest threats of the normal faults on Crete. As scarp heights were only estimated (Caputo et al., 2010) further investigation is needed to definitively determine long-term deformation rates for the Lastros fault. Gaki-Papanastassiou et al. (2009) carried out an investigation into displaced marine terraces on the southern coast near the town of Ierapetra. For the southern part of the Ierapetra fault the authors calculated an average throw-rate of 0.1 mm/a for the last ~400 ka, and because slip fades towards fault tips it is likely that the throw-rate in the central sector of the Ierapetra fault is significantly higher.

There is a long history of human habitation in the area. Evidence for human habitation in the ancient town of Mochlos, located to the north of the Lastros and Ierapetra fault’s footwall (Figs. 4.2, 4.3a) on an island 200 m off the coast, dates back to Early Minoan II (2900–2300 BC) (Davaras, 1975). Many authors also state that there must have been an isthmus between the mainland and Mochlos Island in the Minoan times, which is now underwater. Furthermore, there are fish tanks carved into the coastline rock on the mainland opposite Mochlos Island. These fish tanks are of Roman age (2nd Century AD) (Soles & Davaras, 1992; Mourtzas, 2012) and are now 1-1.5 m underwater, indicating either significant sea level rise or subsidence through earthquakes. As there is no other account of sea levels rising by this amount in the last 2000 years, subsidence by differential movement along the Hellenic Trench (Mourtzas, 2012), or by one of the nearby on-shore faults, may be
responsible. Moreover, Jusseret et al. (2013) identified potential earthquake archaeological effects in Mochlos and to date the responsible fault has not been identified.

Figure 4.3. a) Geological map of the study area showing the locations of the northern part of the Ierapetra fault, Lastros fault and Sfaka fault, modified from Papastamatiou et al. (1959) and Papanikolaou & Vassilakis (2010), contour lines are at 80 m spacing; b) Schematic diagram of the Lastros-Sfaka Graben; c) View of the Lastros Fault; d) View of the Sfaka fault's northern segment.
4.2.2 Historical seismicity

Crete has been designated as having a high seismic hazard due to its location close to the Hellenic arc and trench system (McKenzie, 1972; LePichon & Angelier, 1979; Fountoulis & Mariolakos, 2008) and the island has a long record of destructive earthquakes. Evidence for earthquakes from archaeological remains dates back to the early Minoan civilisation which began in approx. 3000 BC, though the first well documented event is only from the first century BC (Papadopoulos, 2011). Many large earthquakes have occurred along thrust faults associated with the Hellenic Trench; e.g. the last very damaging earthquake, estimated to be M 8 – 8.5, occurred in AD 365 and reportedly caused the destruction of the Cretan towns of Kissamos and Eleutherna (Stiros & Papageorgiou, 2001; Stiros, 2010; Shaw et al., 2008). This earthquake, which occurred off Crete’s western shore, caused the west part of the island to be uplifted up to 9 m, evidenced by notches and elevated sea level marks (Pirazzoli et al., 1982; Stiros, 2010).

In eastern Crete the historical record (Papazachos & Papazachou, 1997) does not show any earthquake before AD 1500. Therefore, it is highly probable that the historical record is incomplete in this area for the pre-AD 1500 era. In addition, no major damage or earthquakes have been reported in eastern Crete from AD 1887 to the present day. There have however been reports of earthquakes in the Sea of Crete (Fig. 4.1b) in the last century. An earthquake of M~6.0 was recorded on 29th February 1940 in the Sea of Crete north of the Lastros-Sfaka Graben and was felt (intensity V) in Sitia (Comninakis & Papazachos, 1986). A similar event of M~6.0 occurred on 30th July 1956 also within the Sea of Crete to the north of the graben; this earthquake was felt (intensity V+) in both Ierapetra and the Fournoi islands in the central Aegean Sea (Fig. 4.1).

Major events in 1810 (M~7.8, intensity IX experienced in Heraklion, estimated depth 90 km), 1856 (M~8.2 intensity IX in Heraklion) and 1887 (M 7.5, estimated focal depth 100 km, intensity VII in Heraklion) were widely felt in the Eastern Mediterranean and are attributed to the subduction zone due to their large magnitude and the wide distribution of isoseismals. There was also a lack of aftershock activity which is a clear indication of deep generated earthquakes (Papazachos & Papazachou, 1997).

The events in the historical record (Papazachos & Papazachou, 1997; Shebalin et al. 1974) that affected eastern Crete and most probably relate to shallow earthquakes occurred in 1508, 1595, 1717, 1780 and 1815; further descriptions of damage caused by the 1780 and 1815 events are provided by Mallet & Mallet (1858) and Sieber (1823), respectively. All of
these events are older than 200 years and therefore a significant error must be implied in both the estimated magnitude (up to ± 0.5) and epicentral locality (up to ± 30 km) (Papazachos & Papazachou, 1997). Seismicity in the last 10 years shows there have been over 20 shallow earthquakes between magnitude 3 and 4 which could be attributed to the faults of the IFZ (EMSC, 2015). This indicates that although there have been no large ruptures in the last 200 years, strain is still being released within the seismogenic zone.

4.3 The Lastros fault

The Lastros fault is located on the western side of the graben and strikes NNE-SSW (020° -200°). The fault was re-mapped in its central sector for approximately 5 km and comprises two segments with a right-stepping geometry (ca. 150 m of stepover) (Figs. 4.2, 4.3a,b,c). The fault extends both north and south showing a total length of 11 km (Caputo et al., 2006). The footwall consists of limestone of the Mani tectonic unit (Papanikolaou & Vassilakis, 2010) making the Kapsos Mountain ridge. The fault’s northern segment comprises footwall limestone juxtaposed against hanging-wall phyllites which are overlain by a colluvial sequence. The southern segment hanging-wall close to the scarp comprises the same carbonate rocks as the footwall overlain by colluvium. The colluvium is mostly unconsolidated material that has fallen from the footwall mountain above the scarp and settled on the hanging-wall; however, in many places over the whole fault the colluvium has become cemented and forms either sheets of varying thickness which are attached to the fault plane, or large talus lobes. Compass measurements show that the southern segment has an average dip angle of ~70° and the northern one an average of ~65°. To the north where the footwall relief significantly drops the fault could not be traced; if the fault does continue to the north towards the coast through the more erodible lithology (Figs. 4.2, 4.3a), erosion and sedimentation within the basin have disguised signs of recent movement. Throughout the fault length the fault scarp ranges in height quite dramatically due to the presence of large gullies which cut both the footwall limestone and hanging-wall sediments (Fig. 4.3b,c). Furthermore, anthropogenic activity along the northern part of the fault also affects the height of the scarp. Terraces have been constructed on the hanging-wall at the base of the scarp to allow the space to be used for agriculture and utilise water runoff. This slope management has artificially exposed the scarp in some areas. The bedrock scarp itself is in most areas brecciated as first observed by Stewart & Hancock (1990).
4.3.1 t-LiDAR analyses

After initial mapping of the fault to determine the best locations for subsequent investigations we scanned the fault’s footwall and hanging-wall with t-LiDAR (terrestrial Light Detection And Ranging) to acquire a 3D point cloud. This was undertaken using the ILRIS 3D laser scanner from Optech Inc. The dataset consisted of 12 scan windows which covered approximately 1.3 km of the fault’s northern segment (Fig. 4.3a). The southern segment could not be scanned because the hanging-wall beneath the scarp is too steep and the oblique scanning angle conceals the majority of the scarp; furthermore, scanning from the southeast side of the graben could not be undertaken as the distances involved were too great for a reliable data set to be acquired. The scanned 1.3 km of the northern segment is deemed representative because the scarp heights of both segments are comparable as determined through manual measurements. The individual scan position point clouds were then processed by cleaning the raw data of isolated points (e.g. from vegetation) and then mathematically and geometrically aligning the individual scan windows. Each scan window’s point cloud contains its own relative coordinate system and these are combined into one global system which totalled more than 25 million measurement points. This point cloud was then georeferenced using GPS data taken in the field, and was then converted into a raster format with a respective 0.5 m cell size resolution using a geographical information system (GIS, ArcMap ESRI®) which creates a spatially regular distribution of the point cloud data. The morphology of the resulting high resolution digital elevation model (HRDEM) was then analysed. To calculate the scarp’s vertical displacement 10 m wide polygons were created each covering the scarp. Using slope angle visualisation the base of the scarp was defined manually as the lower boundary of each polygon. This is easy to define as there is a sharp boundary between the high angle fault plane and the shallower hanging-wall. The upper boundary of the scarp is harder to define as it has undergone some degradation since exhumation of the free face. This has led to a more gradual change in slope angle. The upper boundary of each polygon was therefore defined where the scarp angle becomes the same as the footwall slope above the scarp. The mean difference in elevation between the highest 20 and lowest 20 points in each polygon was then used to determine the scarp height every 10 m along strike. Long-term slip rates were then calculated assuming a constant fault dip angle of 65° and a postglacial exhumation date of 15 ± 3 ka (Allen et al., 1999; Giraudi & Frezzotti, 1997; Papanikolaou et al., 2013). To confirm that the calculated vertical displacements are accurate, profiles perpendicular to the fault’s strike were also generated from the HRDEM (Figs. 4.4c, 4.5). As the fault scarp dips at a higher angle than the footwall and the hanging-wall, the projected slope angles from the footwall and hanging-wall as well as the knickpoints were used to delimit the scarp height allowing the vertical displacement to be calculated and compared.
Figure 4.4. Plots showing the relationship between the Lastros fault scarp geometry and postglacial rates of movement: a) Slope angle map of ca. 1.3 km of the Lastros fault along strike acquired from t-LiDAR point cloud data; b) section view of fault scarp height and elevation along strike; c) vertical displacement of the fault and long-term slip rate assuming a postglacial exhumation date of 15 ka, a 65° dipping fault plane and dip-slip movement (red line), possible error of 1 standard deviation per 10 m segment is also plotted (dashed black lines). Profiles are plotted with a possible error of 20%.
Figure 4.5. Scarp profiles (P1 to P8) derived from the linear regression of t-LiDAR data. \( V = \) vertical displacement (throw), \( T = \) total slip. For profile locations see Fig. 4.4a,c.
Figure 4.4 shows how the scarp vertical displacement (throw) and slip rate (assuming 15 ka for post-glacial exhumation) vary along strike of the scanned segment. The scarp is clearly visible in the HRDEM (Fig. 4.4a) as the discontinuous strip of high slope angle morphology. The scarp is not continuous because of catchment gullies which have eroded the scarp and/or are too deeply incised to be observed in the t-LiDAR scan. There is an overall decrease in scarp elevation to the north of the segment (Fig. 4.4b) and an overall increase in the vertical displacement towards the north (Fig. 4.4c). At some localities there is over 20 m of vertical displacement, and in others there is less than 1 m. There are four main reasons for this variation in vertical displacement: (1) man-made terraces, (2) the presence of cemented colluvium, (3) the proximity of the scarp to channel gullies, and (4) the variation of displacement along strike. The location and scarp heights of selected profiles along strike are shown in figure 4.4c with a 20% variability error (see Papanikolaou et al. 2005; Papanikolaou & Roberts, 2007). The profiles themselves showing the footwall, hanging-wall and fault scarp are presented in figure 4.5.

At approximately 880 – 1020 m along strike (Fig. 4.4c vii) and 1150 – 1310 m along strike (Fig. 4.4c ix) there are man-made terraces located in the hanging-wall below the scarp; at 510 - 830 m along strike (Fig. 4.4c v) there is evidence of ancient terraces having once been present. It is likely that the construction of these terraces has artificially exhumed parts of the scarp. Profiles P5, P6 and P8 (Fig. 4.5) are at locations with contemporary or ancient terraces in the hanging-wall and these profiles show large vertical displacements of 13.2, 12.7 and 17.1 m respectively. Calculations of long-term slip at locations with terraces are therefore inaccurate. Figure 4.4c shows this calculation assuming 15 ka for first exhumation. The slip rate regularly exceeds 1 mm/a and on one occasion exceeds 1.2 mm/a (Fig. 4.4c v). At approximately 1020 – 1150 m is a talus lobe of hanging-wall cemented colluvium. This lobe covers the majority of the scarp. The cementation has likely caused the colluvium to fuse to the footwall block; therefore, when a surface rupturing earthquake occurs the bedrock fault plane is not exhumed because the cemented colluvium does not subside with the rest of the hanging-wall. Vertical displacement above the centre of the lobe is very low at 0.5 - 2 m (Fig. 4.4c viii; Fig. 4.5 P7) and this in turn has led to a very low slip rate of around 0.1 mm/a (Fig. 4.4c viii). The scanned segment contains a number of catchment gullies which also affect the scarp. The catchment gullies lower the scarp base and erode the free face so that in the centre of the catchment the free face is significantly decreased in height (Fig. 4.4c iv) or no longer visible in the scan (Fig. 4.4b, 4.4c ii and vi). Towards the south of the segment, from 0 – 260 m, the free face height begins to decrease (Fig. 4.4c i). This is because the scarp is approaching the end of the segment, and therefore vertical displacements and slip rates are not maxima at these locations (Fig. 4.5 P1). The most reliable areas for vertical
displacement and slip rate calculations are from 340 to 440 m (Fig. 4.4c iii; Fig. 4.5 P2, P3 and P4) as there is little evidence of erosion or sedimentation. Here the average vertical displacement is 9.4 m (1 standard deviation is 0.95 m) which gives a slip rate and age uncertainty of 0.69 ± 0.15 mm/a.

4.4 The Sfaka fault

The Sfaka fault borders the graben to the east and has an average strike of NNE-SSW (020° - 200°). The Sfaka fault was mapped for approx. 5 km, and, like the Lastros fault, it consists of two segments separated by a step over; in the south the fault jumps to the west by approx. 500 m shortening the hanging-wall (Fig. 4.3a). The footwall of the majority of the fault comprises crystalline limestone of the Mani tectonic unit (Papanikolaou & Vassilakis, 2010). The hanging-wall of the northern segment also comprises this limestone, phyllites from the Tripolis unit and colluvium; the hanging-wall of the southern segment comprises limestone of the Mani unit and phyllites from the Western Crete unit. The average dip of the Sfaka fault plane is 63°. As with the Lastros fault, where the footwall relief significantly drops in the north the fault could not be traced (Fig. 4.3a). The fault scarp varies in height quite dramatically due to the presence of large gullies which cut both the footwall and hanging-wall limestone. The largest of these gullies is the Kavousi gorge which runs approximately E-W and completely dissects the fault in the south (Fig. 4.3a). Here the fault scarp is almost completely obscured by high rates of erosion. The fault plane itself is brecciated (Stewart & Hancock, 1990) and appears quite degraded in comparison to the Lastros fault; only where anthropogenic influences have exposed the plane does it appear undegraded with a height of more than 7 m. At the northern segment where there is colluvium in the hanging-wall, the fault follows a narrow valley in which a dirt road has been constructed to access nearby olive groves. Here the dirt road cuts the fault plane at two locations, approximately 40 m from each other. These two road cuts represent vertical trenches cut approximately 75° from the fault’s strike (Fig. 4.6a).

The exposures of the southern road cut, herein referred to as Trench 1 (Fig. 4.6b), and the northern road cut, herein referred to as Trench 2 (Fig. 4.6c), were cleaned of vegetation and the outermost 10 – 25 cm of debris. Furthermore, both trenches were deepened by 0.5 – 1.0 m using hand tools. The trenches were then logged (Fig. 4.7a,b) and photographed and samples were taken from fill units for ¹⁴C dating; samples were also taken from other horizons for classification analyses. ¹⁴C was undertaken at the faculty of Mathematics and Natural Sciences at the University of Cologne.
4.4.1 Trench 1

4.4.1.1 Stratigraphy

Trench 1 is around 7 m-long and 2 m-high and has been excavated 75° from the strike of the bedrock scarp (Fig. 4.6a,b). 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 a fault gouge approximately 1 m thick, which corresponds to about 0.9 m when correcting for the 75° trench angle from fault strike. Using the classification of Woodcock & Mort (2008) this fault gouge is a (meso)cataclasite as it contains around 60% fine-grained cohesive matrix and around 40% clasts. The matrix comprises light yellow clayey silt and the clasts consist of gravels and cobbles of footwall limestone. The western side 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 fill material and they have been radiocarbon dated. CFS1 (Crack Fill South 1) contains light brown to reddish brown very gravelly silty clay with occasional cobbles and roots/rootlets, and CFS2 (Crack Fill South 2) contains light brown to brown gravelly clay including rare cobbles. Both these fissure fills have a high clay content and there is a sharp contact with the colluvial layers further to the west (Fig. 4.7a). The remaining sediments within the trench are colluvial deposits CS1 to CS5. CS1 is cemented colluvium located at the western end of the trench. As with the cemented colluvium observed at the Lastros fault, the cemented colluvium in trench 1 can be described as breccia with sub-angular to sub-rounded clasts in a relatively hard, reddish carbonate matrix containing some pedogenic iron; however, it is not attached to the fault plane as has been observed on the Lastros fault. One possible reason for this is that due to the steepness of the narrow valley where the trench has been excavated, groundwater may be drained slightly away from the fault allowing the precipitation of calcite at these locations. CS2 to CS5 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. Above the most recent colluvium of CS5 is PS (palaeosol) which has been deformed by the most recent faulting event. PS is a brownish red silty clay containing gravels and some rootlets and root traces. The degree of reddening caused by oxidation and the sharp upper boundary to CFS2 indicates the deposit was once at the surface but has now become buried. Interestingly, no palaeosols are exposed/preserved within the colluvial units CS2 to CS5. Above all of the aforementioned units is a layer of recent topsoil, which can be described as a reddish orange brown silty gravelly clay containing active roots/rootlets.
4.4.1.2 Palaeoseismic interpretation

Trench 1 is not dominated by scarp derived colluvial wedges. Instead earthquake evidence comes in the form of fissure fills which have developed within the hanging-wall adjacent to the (meso)cataclasite fault gouge. These fissure fills are faulted against colluvial material which is only partly scarp derived. Due to the nature of the sloping hanging-wall and the location of both trenches close to an alluvial/colluvial fan (Fig. 4.6a), we suggest that the main source of colluvial layers CS2 to CS5 is hanging-wall colluvium from the south at higher elevations. The topographic gradient is high enough for colluvial processes to develop on the hanging-wall, and at present the alluvial/colluvial fan is located 85 m to the west of the trench (Fig. 4.6a). At least two displacement events can be inferred based on the crack fills and
colluvial stratigraphy (Fig. 4.8). Dip slip faulting causes the hanging-wall to be downthrown and tilted; due to a slightly irregular/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 secondary faults within the colluvial layers. The fissure is then filled with scarp derived and local hanging-wall material. The slope surface then stabilises allowing gravelly topsoil to form. The second displacement event then occurs and the above described process is repeated. Numerical age control comes from radiocarbon dating of the crack fill material (see section 4.4.3).

Figure 4.8. A possible retrodeformation of Trench 1 based on two surface rupturing events.
4.4.2 Trench 2

4.4.2.1 Stratigraphy

Trench 2 (Figs. 4.6a,c; 4.7b) is approximately 8 m-long and 3 m-high and shows similar stratigraphy to Trench 1. The degraded limestone footwall is exposed at the eastern end of the trench and is adjacent to (meso)cataclasite fault gouge. The western end of the fault gouge is the primary slip location where there is an abrupt contact to the contiguous units. We interpret these units to be different generations of fill material deposited soon after surface rupturing earthquakes. CFN1 (crack fill north 1) is a dark reddish brown gravelly clay with numerous rootlets. CFN2 (crack fill north 2) can be described as light brownish red gravelly sandy clay. CFN3 (crack fill north 3) is a light brownish red gravelly sandy silty clay. The remaining sediments within the trench comprise colluvial layers CN1 to CN6. As in trench 1, CN1 at the western end of the trench comprises cemented colluvium. The uncremented colluvial layers CN2 to CN6 contain a number of small displacement secondary faults.

4.4.2.2 Palaeoseismic interpretation

Earthquake evidence in trench 2 also comes in the form of fissure fills juxtaposed against the (meso)cataclasite fault gouge. Different generations of fissure fills have become nested due to repeated displacements (Fig. 4.7b). Through retrodeformation analysis (Fig. 4.9) at least three displacement events are identified and the stratigraphy involves formation and deformation processes similar to those described for trench 1. At trench 2 dip slip faulting with a slightly irregular/concave fault plane has created tectonic fissures which are subsequently filled with scarp and local hanging-wall debris. Predominantly hanging-wall derived colluvium from higher elevations to the south then buries the crack fill deposits and the freshly exposed scarp. The slope then stabilises and the whole process is repeated. CFN2 is significantly smaller than the oldest (CFN1) and most recent (CFN3) fissure fills. This is most likely related to the magnitude of the fissure creating event.
4.4.3 Fissure fill age constraints

Six soil samples from the exposed fissure fills and palaeosol in trench 1 and trench 2 were radiocarbon dated and calibrated for atmospheric carbon using the calibration curve IntCal13 (Reimer et al., 2013) using the program OxCal 4.2 (Bronk Ramsey, 2014). The results of the
radiocarbon analyses are shown in table 4.1 and figures 4.7 and 4.10. The oldest three calibrated radiocarbon ages come from trench 2 and the three youngest calibrated radiocarbon ages come from trench 1. All of the calibrated ages come from bulk soil samples of fill material or palaeosols; no charcoal was found within these deposits. Therefore, these represent maximum ages for the deposition, since the samples may have had a significant radiocarbon age before their final deposition as fill material. CFN1 in trench 2 contains the oldest crack fill material indicating that the first observable faulting event occurred after 16055 ± 215 cal BP (Event W). CFN2 in trench 2 contains the next oldest crack fill material indicating that the second oldest observable faulting event occurred after 13168 ± 107 cal BP (Event X). CFN3 in trench 2 and CFS1 in trench 1 both have similar calibrated ages, 10403 ± 151 and 9484 ± 63 cal BP respectively. We interpret that the deposition of these two units was due to a single faulting event (Event Y). The age of the youngest observable crack fill in trench 1 (unit CFS2) is 6626 ± 122 cal BP. This age is similar to that of a palaeosol (unit PS) buried beneath CFS2; the age of PS is 6102 ± 113 cal BP. During rupture the hanging-wall would have been downthrown creating space for the palaeosol to partly fill the crack/cavity. CFS2 was then deposited, partly burying PS. We interpret both of these units to have been the result of the same event, and therefore 6102 ± 113 cal BP is the maximum age for the last observable earthquake (Event Z) in trench 1.

Table 4.1. Radiocarbon ages of fissure fills and vertical displacements at the Sfaka fault.

<table>
<thead>
<tr>
<th>Trench</th>
<th>Sample</th>
<th>Unit</th>
<th>Depth (m)</th>
<th>Sample material</th>
<th>$^{14}$C Age, years B.P.</th>
<th>Calibrated Age, years B.P. *</th>
<th>Throw (m) **</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>S-S01</td>
<td>PS</td>
<td>0.6</td>
<td>Bulk soil</td>
<td>5325 ± 48</td>
<td>6102 ± 113</td>
<td>0.4</td>
</tr>
<tr>
<td>1</td>
<td>S-S03</td>
<td>CFS2</td>
<td>0.8</td>
<td>Bulk soil</td>
<td>5839 ± 46</td>
<td>6626 ± 122</td>
<td></td>
</tr>
<tr>
<td>1</td>
<td>S-S02</td>
<td>CFS1</td>
<td>1.3</td>
<td>Bulk soil</td>
<td>8481 ± 54</td>
<td>9484 ± 63</td>
<td>0.25</td>
</tr>
<tr>
<td>2</td>
<td>S-N02</td>
<td>CFN3</td>
<td>0.8</td>
<td>Bulk soil</td>
<td>9233 ± 55</td>
<td>10403 ± 151</td>
<td></td>
</tr>
<tr>
<td>2</td>
<td>S-N04</td>
<td>CFN2</td>
<td>1.4</td>
<td>Bulk soil</td>
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<td>13168 ± 107</td>
<td>0.1</td>
</tr>
<tr>
<td>2</td>
<td>S-N01</td>
<td>CFN1</td>
<td>1.8</td>
<td>Bulk soil</td>
<td>13355 ± 64</td>
<td>16055 ± 215</td>
<td>0.2</td>
</tr>
</tbody>
</table>

* using calibration curve IntCal13 (Reimer et al., 2013).

** conservatively determined through retrodeformation analysis (Figs. 4.8 and 4.9).
4.5 Discussion

4.5.1 The Lastros fault

Our t-LiDAR investigations show that there is a large variation in displacement throughout the 1.3 km-long scan of the Lastros fault, which is predominantly due to the external influences of erosion and cementation. A slip rate for the Lastros fault has been estimated by Caputo et al. (2006, 2010) based on a maximum throw of 15 m, a dip of 60° and 13 ka for first exhumation. The authors then derive a long-term slip rate of 1.3 mm/a. 15 m of vertical displacement is close to the average vertical displacement of the fault's northern segment but only in areas where the scarp is anthropogenically affected (Fig. 4.4 v, vii and ix). Cemented colluvium in the form of a talus lobe has a serious effect on slip rate calculations by reducing the scarp height by many metres (Fig. 4.4c viii; Fig. 5 P7). The dip angle also decreases (Fig. 4.5 P7) where the talus lobe is present. This lower dip angle of 51° is most likely due to the cemented colluvium being attached to the footwall so that the bedrock scarp is not exhumed during successive uplift events. Therefore the fault plane here has had a significant amount of time to erode to form a shallower angle. This hanging-wall cementation was not just observed as lobes but as sheets of varying thickness attached to the fault plane. The sheets do not significantly affect slip rate calculations.

Based on the 1.3 km-long t-LIDAR survey, and excluding the sector that we consider affected by external influences (Fig. 4.4c iii) and unreliable for extracting data, the throw has an average of 9.4 ± 0.95 m (1 sigma standard deviation), which is lower than previously suggested (Caputo et al., 2006, 2010). This area of postglacial scarp regarded as undisturbed by external influences has a considerable degree of natural variability (Fig. 4.4c iii). Within this area the throw range is from 8.3 to 10.7 m, indicating that the total natural range is ±12.8% of the mean. This range is produced by changes in fault strike and dip and also local morphological differences in both the footwall and hanging-wall, which are themselves caused by localised erosion and sedimentation. Table 4.2 shows possible slip rates for the Lastros fault using these mean, maximum and minimum throw values, a constant fault dip of 65° and exhumation dates between 12 and 18 ka. When the maximum throw is used along with 13 ka for first exhumation the slip rate is 0.91 mm/a, 0.4 mm/a less than reported by Caputo et al. (2006, 2010). As previously stated in section 1, there is an age uncertainty for fault scarp exhumation and 15 ± 3 ka covers the fact that some small magnitude glacial re-advances followed by retreat phases have been recorded in the Mediterranean between 12 ka and 18 ka (Allen et al., 1999; Giraudi & Frezzotti, 1997; Papanikolaou et al., 2013). We therefore present 0.69 ± 0.15 mm/a (Table 4.2) as the representative slip rate for the scanned section of the Lastros fault, which uses the mean
measured throw in the reliable area (Fig. 4.4c iii), and provides a more representative figure as local morphological variations caused by minor amounts of differential erosion and sedimentation are hidden.

The t-LiDAR derived slip rate of 0.69 ± 0.15 mm/a is representative for the scanned section. However, the southern segment remains unanalysed and there is a possibility of larger scarp heights since it located at the centre of the fault (Fig. 4.3a). At the hanging-wall of this southern segment there is very little evidence of erosion from anthropogenic activity, and therefore only natural erosional effects need to be considered when choosing representative areas. As stated in section 2.1 the southern segment could not be scanned because of the oblique scanning angle caused by the steepness of the hanging-wall. Therefore, UAV (unmanned aerial vehicle) survey using photographs and structure from motion software would be most appropriate at this location.

The t-LiDAR derived slip rate of 0.69 ± 0.15 mm/a is comparable with other faults on mainland Greece. For example the Sparta fault in the Peloponnese has a throw rate of 0.8 – 1 mm/a (Benedetti et al., 2002; Papanikolaou et al., 2013); however, the Sparta fault is considerably longer structure, over 60 km in length. We mapped a clear scarp at the Lastros fault for around 5 km. There is no scarp present to the south of Kavousi gorge (Fig. 4.3a) but based on the morphology and the contact between limestone and phyllites, the fault most likely extends to the south by at least 4 km. Caputo et al. (2006, 2010) provide a fault length of 11 km, extending the fault further to the north beyond the mapped bedrock scarp by 2 km. No scarp was observed in the sediments to the north. However, we believe a total length of 11 km is a reasonable assumption as faults do not terminate abruptly and Pleistocene fluvial terraces north of the bedrock fault scarp have a linear northwestern boundary against the Tripolis phyllites extending for around 1.5 km (Fig. 4.3a), indicating that the fault is present here. No evidence for recent displacement in the shoreline deposits was observed indicating that the Lastros fault most likely does not extend offshore. With a total length of only around 11 km there would have had to be very regular seismic activity throughout the post-glacial period to produce 9.4 m of vertical displacement. Using empirical relationships (Pavlides & Caputo, 2004) based on fault length, vertical displacement and 13 ka for first exhumation, Caputo et al. (2010) estimate a recurrence interval of 289 years for the Lastros fault. When the same calculation of \( \log(\text{Maximum Vertical Displacement}) = 1.14 \cdot Ms - 7.82 \) (Pavlides & Caputo, 2004) is used with a maximum earthquake magnitude (Ms) of 6.42 (i.e. 11 km fault length using \( Ms = 0.9 \cdot \log(\text{Surface Rupture Length}) + 5.48 \); Pavlides & Caputo, 2004) there is a maximum vertical displacement of 0.31 m per event. Using a total vertical displacement of 9.4 m over 15 ± 3 ka, a recurrence interval of 499 ± 100 years is obtained. With this
recurrence interval, the Lastros fault may therefore account for one or more of the historical earthquakes that affected the region (section 1.2). This is, however, a small recurrence interval and other factors which may influence this number must be considered. The fault could be greater in length extending many kilometres offshore to the north allowing a much higher vertical displacement per event. However, there is no clear imprint in the geology, topography or bathymetry to justify this. Another possibility is that scarp exhumation is influenced by erosion over the whole scarp length and not confined to areas affected by gullies or anthropogenic activity. However, the process of erosion occurs both in the footwall and hanging-wall, and in reliable areas there is no reason why erosion should be confined to only the hanging-wall. In the reliable area (Fig. 4.4c iii) the total natural scarp height range is ±12.8% of the mean, which shows that minor amounts of erosion must be occurring to produce the variability. This is further justification for using the mean scarp height in the reliable area for the representative slip rate.

Table 4.2. Throw derived from t-LiDAR DEM and slip rates in the reliable area along strike (between 340 and 440 m; Fig. 4.4c iii)

<table>
<thead>
<tr>
<th>Throw (m)</th>
<th>12 ka</th>
<th>13 ka</th>
<th>14 ka</th>
<th>15 ka</th>
<th>16 ka</th>
<th>17 ka</th>
<th>18 ka</th>
<th>15 ± 3 ka</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mean</td>
<td>9.4 *</td>
<td>0.86</td>
<td>0.80</td>
<td>0.74</td>
<td>0.69</td>
<td>0.65</td>
<td>0.61</td>
<td>0.57</td>
</tr>
<tr>
<td>Maximum</td>
<td>10.7</td>
<td>0.99</td>
<td>0.91</td>
<td>0.85</td>
<td>0.79</td>
<td>0.74</td>
<td>0.70</td>
<td>0.66</td>
</tr>
<tr>
<td>Minimum</td>
<td>8.3</td>
<td>0.76</td>
<td>0.71</td>
<td>0.65</td>
<td>0.61</td>
<td>0.57</td>
<td>0.54</td>
<td>0.51</td>
</tr>
</tbody>
</table>

* 1-sigma standard deviation is 0.95 m  
** Slip rate based on dip-slip faulting and a constant dip angle of 65°

4.5.2 The Sfaka fault

Repeated faulting on the Sfaka fault has led to nested fissure fills (Fig. 4.7a,b) which are common where bedrock scarps are juxtaposed against unconsolidated deposits (McCalpin, 2009). As the source of the colluvium is most likely predominantly from the hanging-wall at higher elevations, aggradation at the downthrown block at the trench site is occurring between earthquakes. This has led to paired deposits of slip derived fissure fills and post-slip hanging-wall colluvial deposits which are best preserved in trench 2. McCalpin (2005) refers to this as the fissure graben model. The displacement caused by the events has been estimated by retrodeformation analysis on both the trenches (Figs. 4.8 and 4.9). Uncertainties regarding the retrodeformed vertical displacements concern the effect of the fault gouge. A major amount of slip is assumed to have occurred at the boundary between the fault gouge and the colluvial sediments. This is because clasts within the fault gouge were vertically aligned at this boundary indicating slip at this location. It is possible that the
gouge is accommodating some slip during every event, and/or slip is occurring only on the bedrock plane; however, this is not measurable in either trench and was not considered in the retrodeformation. Further uncertainties concern the geometry of the colluvial deposits. Colluvial deposits should be assumed to have been deposited with non-horizontal top and bottom contacts (McCalpin, 2009); however, the colluvium at the Sfaka trench site is mostly hanging-wall derived from higher elevations, and the depositional geometry at the trench site would most likely not have a falling gradient from the scarp into the hanging-wall. The colluvial layers have therefore been restored to near horizontal positions. Thus, the presented vertical displacements ranging between 0.1 and 0.4 m are minima and represent conservative estimates.

Figure 4.10 shows recurrence intervals calculated using the preferred ages for each faulting event. Intervals were calculated assuming the maximum age for crack fill deposition as the age of the faulting event. The recurrence interval is quite regular for the four identified faulting events, ranging from 2887 to 3684 years. However, the last observable faulting event
Chapter 4: Fault structure and deformation rates at the Lastros-Sfaka Graben

has a maximum age of 6102 ± 113 cal BP. There are a number of possibilities as to why there are no faulting events after this date. The trench location is at the northern extent of the mapped fault and would therefore undergo less displacement per event than at the centre of the fault, and not every surface rupture may have reached the trench site in the north. The fact that Event Z was not even detected in Trench 2 indicates that a complete earthquake record from two trenches at Sfaka fault is unlikely. The Sfaka fault is, however, a secondary structure and antithetic to the Lastros fault, which is the controlling fault in the graben (Figs. 4.3b, 4.11). The Sfaka fault merges with the Lastros fault at a depth of 2.4 km and it likely moves sympathetically depending on the need for volume accommodation in the hanging-wall block of the Lastros fault. Assuming the fissure fills at Sfaka fault were formed coseismically, the extracted dates also represent activity on the Lastros fault. However, the fissure fills may not be coeval with ruptures on the Lastros fault and may only represent accommodation events. If this is the case only when a certain threshold of extension is reached is there an accommodation event. Using the average recurrence intervals for the Sfaka and Lastros faults, 3318 years (Fig. 4.10) and 437 years (section 4.4.1) respectively, an accommodation event may occur after the Lastros fault has ruptured 7-8 times which produces around 1 m of extension.

A recurrence interval at the Sfaka fault ranging from 2887 to 3684 years is not a fully accurate representation due to the nature of the fill material and the dated bulk samples. Not only will there be an inherited \(^{14}\)C signature within the soil before it was deposited in the crack fill, but roots can introduce younger materials into older horizons (Walker, 2005). Further fieldwork is needed to better constrain these dates. As the Sfaka fault has an impressive fault scarp produced through cumulative earthquake events (Fig. 4.3d), a scarp height and exposure investigation using cosmogenic nuclides on an area where there is little evidence of scarp erosion or sedimentation may help better constrain these dates.

4.5.3 Synthesis

The Ierapetra fault, which cuts the whole width of the island (Figs. 4.1, 4.2) and may continue offshore to the north and the south, has been the controlling fault within the IFZ. The Ierapetra fault is an oblique normal fault with a sinistral strike slip component (Papanikolaou & Vassilakis, 2010). Even though slip is oblique on the Ierapetra fault, a significant total throw has accumulated since the development of the fault in the Late Miocene. Cross section A (Fig. 4.11) undertaken close to the southernmost point where the Ierapetra fault still displaces Alpine units (for location see Fig. 4.2) shows an accumulated total throw of around
700 m. Cross section B (Fig. 4.11) undertaken in the north of the IFZ shows the Ierapetra, Lastros and Sfaka faults (for location see Fig. 4.2). Here the Ierapetra fault shows a total accumulated throw of around 500 m. This reduction in total throw indicates that the Ierapetra fault is dying out to the north. The Lastros fault has accumulated around 300 m of total throw (Fig. 4.11). Using a constant slip rate of 0.69 ± 0.15 mm/a, which equates to a throw rate of 0.62 ± 0.14 mm/a, the Lastros fault only began to develop ca. 435 ± 120 ka, making it particularly young. For the same time period the Ierapetra fault displays relatively low slip rates based on displaced marine terraces (Gaki-Papanastassiou et al., 2009). As the Lastros fault is the controlling fault in the Lastros-Sfaka Graben, the Sfaka fault will have a similar or slightly younger age. We believe that the relatively young age of the Lastros and Sfaka faults may be the result of a changing stress regime. The Ierapetra fault is the major fault controlling the geological formations in the IFZ. The Ierapetra fault's orientation and horizontal slip component suggests it may have formed contemporaneously with the Pliny and Strabo trenches in the subduction zone (Fig. 4.1a). Due to the angle of the subduction there is a horizontal component developing in this part of the subduction zone. Regular oblique normal faulting on the Ierapetra fault since the Late Miocene led to the development of up to 700 m of throw in the Alpine units and dramatic footwall morphology (Fig. 4.11). It is likely that in the Mid-Late Pleistocene the stress regime then changed to more pure extension which caused the development of the Lastros and Sfaka faults.

The results presented here provide some age constraints for the studied faults, but without continued research on all of the normal faults the dynamics and neotectonic evolution of the island will remain largely unknown. The proximity of the island to the Hellenic trench gives the whole island a high seismic hazard. However, the way these normal faults have interacted with the trench system and other normal faults throughout the Quaternary is not known. It is likely that some of the faults are less active than others, as we have suggested for those within the IFZ, and stress transfer may leave some fault zones inactive for long periods. Offshore studies are needed to constrain the activity and characteristics of the faults that likely continue offshore, such as the Ierapetra fault. Furthermore, the coastline of Crete shows evidence for large uplift events caused by the Hellenic Trench both in notches now raised by the AD 365 event (Pirazzoli et al., 1982; Stiros, 2010) and ancient shorelines dating back to over 40 ka (Tiberti et al., 2014). We plan to investigate the coastline to determine differential uplift in adjacent areas; local differences in uplift will be due to normal faulting events and will allow us to focus palaeoseismological investigations on the most active faults.
4.6 Conclusions

Geomorphological analysis on the Lastros fault using t-LiDAR shows that previous slip rate estimations are too high and are the result of scarp height measurements in areas influenced by natural erosion and/or anthropogenic activity. Out of 1.3 km of investigated scarp, only a 100 m long section shows little influence from erosion or sedimentation. The natural range of scarp height in this 100 m section is 8.3 to 10.7 m (±12.8% from the mean), which leads to corresponding mean slip rate of $0.69 \pm 0.15$ mm/a using a postglacial ($15 \pm 3$ ka) date for scarp exhumation.

Dating of fissure fills on the Sfaka fault indicate that four events have occurred since $16055 \pm 215$ cal BP. Recurrence intervals range between 2887 and 3684 years and the last
observable event occurred soon after 6102 ± 113 cal BP, indicating that we may be overdue for an event on the Sfaka fault.

The Lastros fault is the controlling fault of the graben with the Sfaka fault being antithetic and merges with the Lastros fault at a depth of 2.4 km. Slip on the Lastros fault will therefore sympathetically affect the Sfaka fault. The extracted dates from the Sfaka fault fissure fills therefore either represent activity on the Lastros fault, assuming they formed coseismically, or the dates represent accommodation events.
Chapter 5: Hanging-wall colluvial cementation along active normal faults

This chapter constitutes a paper which has been submitted for publication in Quaternary Research in August 2016. Reference: Mason, J., Schneiderwind, S., Pallikarakis, A., Mechernich, S., Papanikolaou, I., Reicherter, K. (2016). Hanging-wall colluvial cementation along active normal faults.

5.1 Abstract

Many active normal faults throughout the Aegean juxtapose footwall limestone against hanging-wall colluvium. In places this colluvium becomes cemented and forms large hanging-wall lobes or sheets of varying thickness attached to the bedrock fault. Investigations at the Lastros fault in eastern Crete allow us to define criteria to distinguish between cemented colluvium and fault cataclasite (tectonic breccia) which is often present at bedrock faults. Macro- and microscopic descriptions of the cemented colluvium show that the colluvium was originally deposited through both rockfalls and debris flows. Stable isotope analysis of oxygen and carbon from 83 samples indicate that cementation then occurred through meteoric fluid flow in the fault zone from springs at localised positions along strike. Palaeotemperature calculations of the parent water from which the calcite cement precipitated are indicative of a climate between 7°C and 10°C colder than Crete's present average annual temperature. This most likely represents the transition between a glacial and interglacial period in the Late Pleistocene. Ground penetrating radar also indicates that cemented colluvium is present in the hanging-wall subsurface below uncemented colluvium. Using these results a model for the temporal development of the fault and formation of the cemented colluvium is proposed.
5.2 Introduction

Quaternary colluvial sedimentary deposits are associated with the downslope zone of steep bedrock slopes and topographic escarpments, and have long been recognised as an important product of mass wasting (Blikra & Nemec, 1998). Colluvial processes involve rockfalls/debris falls, debris flows and channelized or unconfined water flow (Nemec and Kazanci, 1999; Ventra et al., 2013), which mobilise and then deposit the colluvial material. These processes depend strongly on the local geology and geomorphology and the climatic conditions at the time of deposition. Colluvial deposits in general have quite a small spatial extent, but with the right conditions colluvial sequences can form deposits traceable for many kilometres and be many tens of metres thick. Such conditions can occur in extensional tectonic environments where fault activity produces the volume/space for colluvial deposition. Here, large bedrock footwall mountain fronts are uplifted forming steep slopes and hanging-wall slopes are subsided. These normal faults are typical for the Aegean region and colluvial processes can deposit thick sequences in the hanging-wall covering large areas.

Colluvial sedimentation is by its nature episodic; the study of these deposits provides information about a specific time period when the colluvium was deposited, and its use as a proxy record for terrestrial climatic conditions has been shown by many authors (e.g. Blikra & Nemec, 1998; Nemec and Kazanci, 1999; Gradzinski et al., 2014). Quaternary colluvial sequences throughout the world mainly comprise non-cemented gravel deposits which have been used in applied geological studies (gravel exploitation, slope stability, etc.) for hundreds of years. However, occasionally the colluvial gravels become cemented by mineral precipitation between the gravels from near surface or groundwater fluids. Some researchers refer to cemented colluvial deposits as a type of travertine. For example, the term cemented rudite is used in the travertine classification of Pentecost (1993, 2005) and Pentecost and Viles (1994) to describe “surface cemented rudites, consisting of cemented screes, alluvium, breccia, gravel etc.” The term calcrete was also originally used to refer to cemented deposits (Lamplugh, 1902); however, calcretes are now mostly regarded as carbonates formed within soil profiles and are mostly associated with low sedimentation rates. To avoid confusion we refer to unconsolidated deposits cemented by secondary carbonate as ‘cemented colluvium’ as this is a descriptive term lacking genetic connotations.
The focus of this study is cemented colluvium which is present on the Lastros fault located within the lerapetra fault zone (IFZ) in eastern Crete, Greece. Crustal extension has led to the development of normal faults throughout Crete, and many of these faults comprise
footwall limestone bedrock scarps mainly juxtaposed against Quaternary colluvium, which has fallen from the footwall mountain above the scarp and landed on the hanging-wall. In places this colluvium has often become cemented forming large hanging-wall lobes or spatially irregular sheets of varying thickness parallel to the bedrock fault scarp. These colluvial sheets should not be confused with fault cataclasite (tectonic breccia) which also forms sheets and belts parallel to faults (e.g. Stewart and Hancock, 1988; 1990). We therefore provide characteristic criteria for a distinction to be made.

The cement within cemented colluvium has precipitated from fluid flow within the fault zone. However, the origin and nature of this fluid is unknown. Faults themselves can be both conduits and barriers for fluid movement during the seismic cycle (Brown and Bruhn, 1996; Uehara & Shimamoto, 2004; Wibberley et al. 2008). To understand the processes of fluid flow within fault zones, the cement precipitated from the fluid therefore needs to be investigated. Laboratory analyses of the stable isotope composition of carbon and oxygen within the cemented colluvium, and also ground penetrating radar (GPR), indicate the cement’s likely formation processes and give insights into how the fault has evolved throughout the Quaternary.

5.3 Geological and tectonic setting

The island of Crete is located in the southern Aegean region just north of the Hellenic Arc and Trench System (Fig. 5.1a). Late Miocene–Quaternary extension (Papanikolaou & Vassilakis, 2010) has led to a complex pattern of extensional detachment faults and high angle normal faults throughout the region. This is attributed to crustal back arc extension, interpreted as a response to the southward slab-rollback of the Hellenic margin, the southwestward expulsion of the Aegean microplate and the anticlockwise rotation of the African lithosphere relative to Eurasia (Meulenkamp et al., 1988; Reilinger et al., 2006). The southward slab-rollback is the predominant mechanism; from the Middle Miocene the central and southern Aegean domain began to extend rapidly in a north-south direction implying the rapid migration of the Crete trench relative to northern Greece (Angelier et al. 1982; Royden and Papanikolaou, 2011; Jolivet et al. 2013). Extension is occurring orientated both arc-perpendicular and arc-parallel creating normal faults roughly oriented both NNE – SSW and ESE– WNW (Fig. 5.1b). These faults cross-cut all of the bedrock geology, which comprises Alpine thrust sheets developed as a result of oceanic closure and subsequent nappe stacking. Crete comprises several thrust sheets which have been successively imbricated by other sedimentary and metamorphic units (van Hinsbergen et al., 2005; Papanikolaou & Vassilakis, 2010). In eastern Crete the stratigraphically lowest Mani unit (Fig. 5.1c) is the
autochthon basement comprising crystalline limestone. This unit was imbricated by the Western Crete unit (Fig. 5.1c) which comprises mainly Permo-Triassic phyllites and Middle to Late Triassic evaporites. The Arna unit was next to be imbricated, but these rocks have not been preserved in eastern Crete; the largest exposures of the Arna unit are in western Crete and comprise low to medium grade metamorphic phyllites and quartzites. The Tripolis unit (Fig. 5.1c) was then imbricated on the Arna unit and comprises a thick sequence of flysch, limestone, dolomite, andesite, diabase and phyllites, all of which are preserved in eastern Crete. These rocks form the volcano-sedimentary base to the shallower carbonate rocks of the Tripolis carbonate platform, exposed in both Crete and the Peloponnese (Papanikolaou & Vassilakis, 2010).

The study area is the Lastros fault located within the Ierapetra fault zone (IFZ) in eastern Crete (Fig. 5.1b,c). The IFZ consists of a roughly 25 km long zone of fault segments dipping to the WNW and ESE. Within the IFZ colluvial sequences have developed throughout the Quaternary. The largest colluvial sequences occur at the Kavousi fault where thick aprons of colluvium have been deposited, predominantly in the hanging-wall but in some locations also covering the lower reaches of the footwall (north of Kavousi town in Fig. 5.1c). The high deposition rate of this colluvium does not allow a scarp to be exposed on the Kavousi fault. The Lastros fault is another fault within the IFZ which has significant colluvial deposits. The Lastros fault is 11 km long and has a clear fault scarp for approximately 5 km, which comprises two segments separated by a step over; in the south the fault jumps back into the footwall by around 150 m (Mason et al., 2016) (Figs. 5.1c, 5.2a,b). The footwall is comprised of limestone of the Mani tectonic unit (Papanikolaou & Vassilakis, 2010) making the Kapsos Mountain ridge and is brecciated as first observed by Stewart & Hancock (1988, 1990). The fault’s northern segment comprises footwall limestone juxtaposed against hanging-wall phyllites which are overlain by a colluvial sequence. The southern segment’s hanging-wall close to the scarp comprises the same carbonate rocks as the footwall overlain by colluvium. The colluvium is mostly unconsolidated material that has fallen from the footwall mountain...
above the scarp and settled on the hanging-wall; however, in many places over the whole fault the colluvium has become cemented.

Figure 5.3. a) t-LiDAR derived DEM showing slope angle. Black arrows indicate the fault scarp. b) t-LiDAR derived hillshade DEM showing the calcite vein sampling site and GPR location. Yellow box indicates the location of c and d. Red arrows indicate the fault scarp. c) close-up hillshade view of the lobe from the t-LiDAR derived DEM showing profile location of cross section (see Fig. 5.10a). d) close-up aerial view of the lobe using Google Earth imagery from April 2013. Cemented colluvium sampling locations are shown in yellow with respective numbers indicating metres from the fault plane.
Chapter 5: Hanging-wall colluvial cementation along active normal faults

Figure 5.4. Photoplate showing the forms of cemented colluvium on the Lastros fault. a) view of the north part of the Lastros’ lower segment showing a lobe of cemented colluvium sitting proud in the hanging-wall, yellow arrows indicate the scarp; b and c) sheets of cemented colluvium of varying thickness attached to the Lastros fault’s bedrock plane; d) schematic diagram showing the structure of the tectonic breccia (cataclasite) and cemented colluvium (modified from Stewart and Hancock, 1988).

A large hanging-wall lobe of cemented colluvium at the northern end of the fault’s lower segment is approximately 110 m long and 140 m wide (Figs. 5.3a,b,c,d, 5.4a) and forms an impressive protuberance within the hanging-wall morphology. In some areas the cemented colluvium forms sheets parallel to the fault plane (Fig. 5.4b,c,d) which can be many metres thick, and in others it is completely absent allowing the bedrock scarp to be exposed. Where present the thickness of these sheets attached to the plane ranges between around 10 cm and 3 m. However, there are areas where only remnants of the cemented colluvium are left
on the fault plane and the thickness is therefore much less than originally; this is caused by subaerial erosion of the sheet after exposure. In certain locations within the colluvial sheets, calcite has grown forming veins and crystals (Fig. 5.4c). Terraces have been constructed on the hanging-wall at the base of the scarp to allow the space to be used for agriculture and utilise water runoff. This construction process has artificially exposed the scarp and cemented colluvium in some areas.

Cemented colluvium is present at some other normal faults in the Aegean region. Altunel and Hancock (1993) describe locally cemented footwall derived talus at the Pamukkale fault in western Turkey as ‘range-front travertines’ deposited from spring water. Hancock and Barka (1987) describe trails and tabular sheets of ‘brecciated colluvium’ attached to several fault planes in western Turkey. ‘Indurated breccia’ is mentioned to be present on a number of faults on Crete (Stewart and Hancock, 1991; Caputo et al. 2006; Caputo et al. 2010). The Zou fault within the Sitia fault zone (Fig. 5.1b) has a prominent cemented lobe, and sheets of varying thickness are present on the Kera, Asomatos, Spili and Kastelli faults (Fig. 5.1b). The formation process of this cemented colluvium (brecciated colluvium or indurated breccia) is, however, not discussed in the aforementioned publications and little work has been undertaken on these deposits and their relationship to individual faults.

5.4 Methods

5.4.1 Field methods

Ground Penetrating Radar (GPR) was undertaken on the hanging-wall of the fault to determine whether cemented colluvium was present in the hanging-wall subsurface below uncemented colluvium. Several GPR profiles were undertaken on an area where thick vegetation and boulders were at a minimum, allowing easier access to the hanging-wall (see Fig. 5.3b for location). This allowed good coupling of the GPR antenna with the ground surface. We used a GSSI 270 MHz antenna with a survey wheel, a SIR-3000 data recording unit, and a handheld GPS to mark the start and end of the profiles. Trace increment was set to 0.02 m to achieve a high data density; the range was set to 120 ns TWT (two way traveltime) and with a sample rate of 512 scans/sec. This achieved a penetration depth of about 6 m. Data processing was then done with ReflexW V6.1 (Sandmeier Scientific Software) which included static correction, background removal, gain adjustments and velocity adaption for time-depth conversions. The latter was based on a hyperbola analysis where such features were present. In other cases, we estimated characteristic velocities presented by Neal (2004). In order to confirm the GPR results, trial pitting was also
undertaken with hand tools. This allowed the shallow geology to be confirmed and time-depth-conversion to be carried out.

t-LiDAR (terrestrial Light Detection And Ranging) scanning was undertaken along the fault (Fig. 5.3a,b) covering the footwall and hanging-wall. This was undertaken so that the surface morphology of the footwall and hanging-wall can be visualised along with radar reflections of the hanging-wall subsurface. We used an ILRIS 3D laser scanner from Optech Inc. The t-LiDAR scanning produced a point cloud that was cleaned of isolated points (e.g. vegetation), geometrically aligned and then georeferenced using GPS data taken in the field. A raster was then produced with a 0.5 m cell size resolution using a geographical information system (GIS, ArcMap ESRI®) which creates a spatially regular distribution of the point cloud data. The GPR profiles were then georeferenced using GPS data and manual measurements. Topographic correction using the t-LiDAR data was then applied to the GPR profiles.

5.4.2 Laboratory methods

Cemented colluvium samples from the Lastros fault were extracted from the lobe structure towards the fault’s northern end (Fig. 5.3a,b). Samples were collected from locations with no vegetation cover at ~5 m intervals from the fault plane into the hanging-wall up to a distance of 59 m (Fig. 5.3d). This was undertaken by hammering the deposit until a cobble size piece was freed. A calcite vein sample was also taken from within a sheet (Fig. 5.3b). The cemented colluvium samples were cut, photographed, and stained with Alizarin red S (ARS) and potassium ferricyanide dissolved in a dilute hydrochloric acid solution in order to confirm that the composition of the cement was carbonitic (Dickson, 1966; Evamy 1963). The sedimentary descriptions of the mineralogy and texture of the cemented colluvium were based on the macroscopic and microscopic characteristics of the material. Thin sections were made of representative areas and were viewed using a standard petrographic microscope. Soft-cemented (friable) samples were first impregnated with an epoxy resin to make them harder for cutting. All samples were sectioned, glued to a glass slide (3 - 5 mm) and polished using progressively finer abrasive grit until the sample was 20-30 µm thick.

To reconstruct the occurrence, origin and movement of fluids within the fault zone, the stable isotope geochemistry of the cement was investigated. The ratios of two isotopes pairs $^{13}\text{C}/^{12}\text{C}$ and $^{18}\text{O}/^{16}\text{O}$, which are respectively expressed as the delta values $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$, are measured and used to determine the physio-chemical conditions of the source water from which the cement precipitated. The ratios relate the isotopic concentration of the sample to that of a standard, and delta value are said to be either heavier (enriched) or lighter...
(depleted) than the standard (Pentecost, 2005). The standard used for carbonates is the VPDB (Vienna Pee Dee Belemnite) - originally from the Cretaceous Pee Dee Formation (Urey et al., 1951) but has now been replaced by the artificial Vienna PDB which is a marble (Pentecost, 2005). Ratios are expressed as ‰ providing an expanded scale for the comparatively small differences observed. Fractionation, resulting from evaporation, degassing or metabolic consumption of the lighter isotopes during the (bio)geochemical cycle, induces discrete changes in $\delta^{13}C$ and $\delta^{18}O$, providing information about the physicochemical conditions of the precipitation events (Gandin and Capezzuoli, 2008).

Calcite cements form through the degassing of surfacing CO$_2$-rich groundwaters. The concentrations of $\delta^{18}O$ and especially $\delta^{13}C$ in the cement can be used to determine the source of the CO$_2$, which can be either meteoric or thermal. Meteoric CO$_2$ originates from the atmosphere and the soil-zone. Cements formed from groundwaters charged with meteoric CO$_2$ are termed metegenes and typically form from cold-water springs in regions underlain by carbonates. Thermal CO$_2$ originates from thermal processes within or even below the Earth’s crust. Thermally generated CO$_2$ dissolves in groundwater producing high concentrations of CO$_2$, which are capable of dissolving large volumes of rock. Cements formed from groundwaters charged with thermal CO$_2$ are termed thermogenes and form from hot-water springs (Pentecost, 2005). The concentration of $\delta^{18}O$ can be used to estimate past water temperatures. Groundwater exchanges oxygen atoms with the dissolved CO$_2$, and where isotopic equilibrium conditions are established, meaning there is isotope exchange between the calcite and water phases, the difference between the ratio of oxygen isotopes in the precipitated calcite and that of the parent water can be used to estimate the palaeotemperature (Pentecost, 2005).

Colluvium has a porous framework and calcite cement precipitating waters can penetrate below ground and continue to precipitate calcite cement downstream of the source spring. At emergence the source water is supersaturated with carbonate, and calcite then precipitates on the pre-existing gravels. As penetration continues downstream and below ground the supersaturation will fall until saturation is reached and no further calcite precipitation will occur (Pentecost, 2005). Calcite precipitating groundwaters have $^{18}O$ compositions similar to the local precipitation, but groundwaters are often slightly enriched in $^{18}O$ compared to rainwater due to surface evaporation, which preferentially affects the lighter $^{16}O$ isotopes (Yurtserver, 1976; Pentecost, 2005).

Powdered samples were obtained by microdrilling of the unstained cut surface using a Dremel drill. 83 powder samples were analysed to determine the composition of: 1) calcite
crystals in the sampled vein, 2) crystalline calcite cement, 3) crystalline clay rich calcite cement, 4) soft clay rich calcite cement, 5) clay rich rims around clasts, 6) calcite growing in secondary voids, and 6) limestone clasts. The size of the hole drilled will have an effect on results, as a large hole will cover a larger time period of calcite precipitation, and fractionation conditions may have changed. Therefore, the microdrilled holes were kept to <3 mm diameter to try and minimise this effect as only 50-100 mikrograms of cement powder are required per sample. To determine the stable isotopic composition of calcite samples we used a Finnigan 253 gas mass spectrometer coupled to an automatic carbonate preparation device Kiel IV at the Alfred Wegener Institute in Bremerhaven, Germany. The mass spectrometer was calibrated via international standard NBS19 to the PDB scale, and results are given in δ notation versus VPDB. The precision of δ¹⁸O and δ¹³C measurements, based on an internal laboratory standard (Solnhofen limestone) measured over a one-year period was better than ±0.08 and ±0.06 ‰, respectively.

5.5 Results

5.5.1 Cemented colluvium descriptions

The source of colluvial deposits at the Lastros fault is relatively small and consists of carbonates from the Mani unit (Fig. 5.1c, 5.2b). The lobe structure has three distinct cemented colluvium facies that can be recognised. The classification of these facies is dependent on the clay content of the cement and the clast composition and texture. The macroscopic and microscopic characteristics of these three facies are described below:

Facies 1 can be described as a matrix-supported breccia with crystalline cement containing a minor constituent of fines. This facies occurs at 9 and 40 m from the fault plane. The cemented colluvium comprises poorly sorted carbonate rock clasts. The vast majority of these clasts are crystalline limestone from the Mani unit (Papanikolaou & Vassilakis, 2010) and are light blue/grey in colour (Fig. 5.5a). There are occasional lithic fragments of impure limestone and some dolomite fragments are also present which do not stain red with ARS (Fig. 5.5b). The carbonate clasts are mainly angular and elongate in shape indicating short transport distances (Fig. 5.5a,b,c). The clasts are mainly coarse gravel size (between 2 and 6.3 cm); however, larger cobbles are occasionally present. Dolomite fragments are generally medium gravel size (between 0.63 and 2 cm). The cemented colluvium is mainly matrix supported with constituent clasts rarely touching. The matrix mainly comprises sparry calcite cement. Where there is relatively pure calcite there is an isopachous cement rim of micrite around the parent clast (Fig. 5.5d). The size of the crystals then significantly increases towards the centre of the void and the crystals become drusy and blocky (Fig. 5.5d). In some
areas the cement appears slightly light brown in colour; here clay/silt minerals are a minor constituent. Occasional voids are also present within the cement matrix.

![Figure 5.5](image)

Facies 2 can be described as a matrix-supported breccia with crystalline cement with a medium fines content. Fines are particles with sizes <0.063 mm. This facies is the most common occurring at 19, 35, 51, 57 and 59 m from the fault plane and the clast composition is similar to facies 1, with the majority of clasts comprising crystalline limestone from the Mani unit (Papanikolaou & Vassilakis, 2010). The cemented colluvium is matrix supported with no clasts touching (Fig. 5.6a,c,e). The major difference compared to facies 1 is the composition of the cement matrix. The cement matrix comprises between 30 and 50% fines. The fines comprise poorly sorted rounded peloids of clay/mud. Surrounding most parent carbonate clasts there is, in most cases, a clay/mud coating between 100 and 200 µm in thickness (Fig. 5.6b,d,g,h) which sometimes projects into the pore space. The remaining 50 to 70% comprises sparry calcite. This proportion of calcite allows the cement to have a crystalline appearance when viewed macroscopically. Sparry calcite crystals form in between peloids of clay/mud and have a drusy texture. There are still some voids visible within the cement matrix (Fig. 5.6b,d,e) indicating where the most recent calcite crystals were growing.
Figure 5.6. Facies 2: a) Cut sample 19 m from the fault plane. b) Thin section image (cross polars) from the 19 m sample showing the sparry calcite cement (Ca), clay minerals comprising ~50% of the matrix (Cl) and the clay isopachous cement rim around the parent clast (P). c) Cut sample 51 m from the fault plane. d) Thin section image (cross polars) from the 51 m sample showing the sparry calcite cement (Ca), clay minerals comprising ~50% of the matrix (Cl) and lithic fragments of impure limestone (Li). e) Cut sample 35 m from the fault plane. f) Cut sample 35 m from the fault plane stained with ARS. g) Thin section image (cross polars) from the 35 m sample showing elongate crystals of pure sparry calcite cement (Ca), voids in which the calcite crystals were growing, the fine grained isopachous cement rim around the parent clast (P), and mud/clay minerals (Cl). h) Higher magnification image from (g) showing the clay minerals (Cl) and calcite (Ca) in more detail and the clay rim around the parent clast (P).
Facies 3 can be described as a matrix-supported breccia with cement which has a high fines content. This facies occurs at 16 and 24 m from the fault plane. The clast composition is similar to facies 1 and 2, but rare fragments of siliceous material are also present (Fig. 5.7a,b). The main difference is the composition of the cement matrix. The matrix comprises between 50 and 70% poorly sorted rounded peloids of clay/mud, with the remaining 30 to 50% being sparry calcite cement (Fig. 5.7a,b,c). This high fines content makes the matrix rather friable and no longer appear crystalline when viewed macroscopically (Fig. 5.7a,c).

Figure 5.7. Facies 3: a) Cut sample 16 m from the fault plane. b) Thin section image (cross polars) from the 16 m sample showing fine sparry calcite cement and clay minerals (Cl) comprising ~60% of the matrix (Cl), lithic fragments of impure limestone (Li), dolomite fragments (D) and a siliceous fragment (Si). c) Cut sample 24 m from the fault plane. d) Cut sample 24 m from the fault plane stained with ARS; note the dolomite fragments which do not stain red.

5.5.2 Stable isotope analyses

The stable isotope composition of facies 1, 2 and 3 as well as the calcite vein from within a colluvial sheet, totalling 83 measurements, are listed in Table 1 and presented in Figure 5.8. The calcite vein shows the highest negative values of both δ\(^{13}\)C and δ\(^{18}\)O, whereas the limestone bedrock clasts from all facies show the lowest negative δ\(^{18}\)O values, with many having positive δ\(^{13}\)C values (yellow triangles in Fig. 5.8). There is no clear correlation between δ\(^{13}\)C and δ\(^{18}\)O for facies 1, 2 and 3. However, there is a clear grouping of the values. The relatively pure cement of facies 1 has δ\(^{13}\)C values ranging from -10.2 to -7.7 ‰,
and δ^{18}O values ranging from -5.4 to -4.5 ‰. The crystalline clay rich cement of facies 2 has a larger range of δ^{13}C values from -9.0 to -2.3 ‰, and also a larger range of δ^{18}O values from -5.8 to -4.3 ‰. The soft clay rich cement of facies 3 has δ^{13}C values ranging from -10.2 to -7.8 ‰, and δ^{18}O values ranging from -5.4 to -5.0 ‰. The sampled secondary void areas in facies 1 and 3 generally show similar isotope values to the cement. However, for facies 2 there is a large scatter in both δ^{13}C and δ^{18}O values, which range from -8.1 to 1.4 ‰ and -5.5 to -3.0 ‰ respectively.

Figure 5.8. Stable isotope composition of the colluvial cement, calcite vein and limestone clasts. The calcite vein shows the highest negative values of both δ^{13}C and δ^{18}O, whereas the limestone bedrock clasts have the lowest negative δ^{18}O values and mainly positive δ^{13}C values.
### Table 5.1: Stable isotope composition of carbonate samples

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<th>Sample description</th>
<th>Facies</th>
<th>$\delta^{13}$C (‰ V-PDB)</th>
<th>$\delta^{18}$O (‰ V-PDB)</th>
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### Chapter 5: Hanging-wall colluvial cementation along active normal faults

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<td>1.428</td>
<td>-2.97</td>
</tr>
<tr>
<td>LL62</td>
<td>57</td>
<td>Crystalline clay rich calcite cement</td>
<td>2</td>
<td>-2.273</td>
<td>-4.482</td>
</tr>
<tr>
<td>LL63</td>
<td>57</td>
<td>Crystalline clay rich calcite cement</td>
<td>2</td>
<td>-6.089</td>
<td>-4.804</td>
</tr>
<tr>
<td>LL64</td>
<td>57</td>
<td>Crystalline clay rich calcite cement</td>
<td>2</td>
<td>-6.399</td>
<td>-4.913</td>
</tr>
<tr>
<td>LL65</td>
<td>57</td>
<td>Limestone bedrock clast</td>
<td>2</td>
<td>1.076</td>
<td>-2.471</td>
</tr>
<tr>
<td>LL66</td>
<td>57</td>
<td>Limestone bedrock clast</td>
<td>2</td>
<td>1.031</td>
<td>-2.605</td>
</tr>
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<td>LL67</td>
<td>57</td>
<td>Limestone bedrock clast</td>
<td>2</td>
<td>1.439</td>
<td>-2.545</td>
</tr>
<tr>
<td>LL68</td>
<td>59</td>
<td>Crystalline clay rich calcite cement</td>
<td>2</td>
<td>-7.616</td>
<td>-5.538</td>
</tr>
<tr>
<td>LL69</td>
<td>59</td>
<td>Edge of void within calcite cement</td>
<td>2</td>
<td>-6.41</td>
<td>-5.508</td>
</tr>
<tr>
<td>LL70</td>
<td>59</td>
<td>Crystalline clay rich calcite cement</td>
<td>2</td>
<td>-6.084</td>
<td>-5.812</td>
</tr>
<tr>
<td>LL71</td>
<td>59</td>
<td>Crystalline clay rich calcite cement</td>
<td>2</td>
<td>-6.19</td>
<td>-5.206</td>
</tr>
<tr>
<td>LL72</td>
<td>59</td>
<td>Crystalline clay rich calcite cement</td>
<td>2</td>
<td>-7.181</td>
<td>-5.278</td>
</tr>
<tr>
<td>LL73</td>
<td>59</td>
<td>Crystalline clay rich calcite cement</td>
<td>2</td>
<td>-7.13</td>
<td>-5.303</td>
</tr>
<tr>
<td>LL74</td>
<td>59</td>
<td>Limestone bedrock clast</td>
<td>2</td>
<td>2.048</td>
<td>-2.069</td>
</tr>
<tr>
<td>LL75</td>
<td>59</td>
<td>Limestone bedrock clast</td>
<td>2</td>
<td>1.685</td>
<td>-2.417</td>
</tr>
<tr>
<td>LV1</td>
<td>0.2</td>
<td>Calcite crystal in vein</td>
<td>CV</td>
<td>-10.822</td>
<td>-6.38</td>
</tr>
<tr>
<td>LV2</td>
<td>0.2</td>
<td>Calcite crystal in vein</td>
<td>CV</td>
<td>-10.308</td>
<td>-5.63</td>
</tr>
<tr>
<td>LV3</td>
<td>0.2</td>
<td>Calcite crystal in vein</td>
<td>CV</td>
<td>-10.368</td>
<td>-5.616</td>
</tr>
<tr>
<td>LV4</td>
<td>0.2</td>
<td>Calcite crystal in vein</td>
<td>CV</td>
<td>-11.403</td>
<td>-6.316</td>
</tr>
<tr>
<td>LV5</td>
<td>0.2</td>
<td>Calcite crystal in vein</td>
<td>CV</td>
<td>-11.363</td>
<td>-5.845</td>
</tr>
<tr>
<td>LV6</td>
<td>0.2</td>
<td>Calcite crystal in vein</td>
<td>CV</td>
<td>-11.343</td>
<td>-6.106</td>
</tr>
<tr>
<td>LV7</td>
<td>0.2</td>
<td>Calcite crystal in vein</td>
<td>CV</td>
<td>-11.204</td>
<td>-5.829</td>
</tr>
<tr>
<td>LV8</td>
<td>0.2</td>
<td>Calcite crystal in vein</td>
<td>CV</td>
<td>-10.907</td>
<td>-5.549</td>
</tr>
</tbody>
</table>
5.5.3 Ground penetrating radar

Three GPR profiles undertaken perpendicular to the fault dip are presented having been combined with the t-LiDAR derived DEM (Fig. 5.9a,b; see Fig 5.3b for location). All three profiles are between 30 and 35 m long and are spaced between 10 and 20 m apart. All GPR profiles have distinct radar facies (Neal, 2004). In particular six are identified, five of which are repeated in each profile (Fig. 5.9a,b). Radar Facies 1 (RF1) can be described as having planar shaped parallel reflectors with a horizontal dip in relation to the hanging-wall slope surface. Due to the geometry of these reflectors, which end abruptly only a few metres along each profile, RF1 is interpreted as the subsurface continuation of the horizontally bedded limestone exposed in the footwall. Radar Facies 2 (RF2) is described as having planer to sinuous shaped, parallel to sub-parallel reflectors which are moderately continuous and have a horizontal dip in relation to the slope surface. RF2 is located in the top 10 to 30 cm of each profile and represents recent soil of the vegetated surface and was confirmed through trial pitting. Radar facies 3 (RF3) has discontinuous and sinuous shaped reflections which are sub-parallel to oblique. Some hyperbolae are present indicating the presence of boulders. RF3 is located in all profiles beneath the recent soil layer in the first 20 m of each profile and is interpreted as anthropogenically affected colluvial deposits containing boulders of either limestone or cemented colluvium. Radar Facies 4 (RF4) is perhaps the most interesting facies. It is present in all profiles up to around 20 m horizontal distance from the bedrock fault scarp. Only in the southernmost profile 1 is RF4 not present close to the fault plane; here RF4 begins approximately 8 m from the fault. This radar facies has low reflectance properties compared to the surrounding facies, which indicates that it is a significantly harder material. The reflections that are present can be described as planer to sinuous, parallel to sub-parallel and discontinuous.

Two trial pits were excavated on the northernmost GPR profile in order to see RF4 and its upper boundary. In both trial pits at 0.8 m our hand tools could not dig any further due to cementation of the colluvial gravels; we therefore interpret RF4 as cemented colluvium. There are two other radar facies present in figure 5.9: Radar Facies 5 (RF5) is located beneath RF4 in both profile 3 (northernmost) and profile 2 (central profile). This is most likely a continuation of the cemented colluvium at depth. RF5 is also located up to 8 m from the fault plane in profile 3 and is most likely cemented. Radar Facies 6 (RF6) is located further into the hanging-wall. It has planer to sinuous, sub-parallel, moderately continuous reflections and much higher reflectance properties than RF4 and RF5. We interpret RF6 as uncemented colluvium. The cemented colluvium (RF4 and RF5) stops quite abruptly around 20 m from the fault plane, this is particularly evident in profiles 2 and 3.
Figure 5.9. a) Un-interpreted and interpreted GPR profiles at the Lastros fault (see Fig. 5.2b for location). b) Close-up view of interpreted GPR profile 3.
5.6 Discussion

5.6.1 Cemented colluvium identification criteria

As first identified by Stewart and Hancock (1988, 1990), the footwall of the Lastros fault is brecciated, and this can be observed in areas where no cemented colluvium is attached to the fault. Where only thin sheets of breccia are located on the bedrock fault plane it needs to be determined whether this breccia has been formed through the faulting process (cataclasite), or it is actually the remnants of cemented colluvium attached to the fault plane. For a clear differentiation to be made Table 2 shows the criteria that must be considered. Tectonic breccia (cataclasite) comes in two forms, compact breccia sheets and incohesive breccia belts (Fig. 5.4d). To distinguish the latter from cemented colluvium is relatively simple because incohesive breccia belts often have no fine grained matrix and the clast boundary matching is moderate to high (Stewart and Hancock, 1988); furthermore phyllosilicate-rich pressure solution seams (Bullock et al., 2014) are indicative of incohesive breccia belts and are not present in cemented colluvium (Table 2). To differentiate between compact breccia sheets and cemented colluvium, the internal structure, matrix composition, clast sizes and clast angularity are the best criteria to consider. The presence of foliation (Bussolotto et al. 2007) is indicative of compact breccia sheets as foliation is not present in cemented colluvium. The matrix composition of cemented colluvium generally has a higher clay/mud content than the matrix of compact breccia sheets; however, an up to 2 mm thick clay-rich gouge layer can sometimes coat the primary slip surface of compact breccia sheets (Bullock et al. 2014). Furthermore, the cemented colluvium matrix can also have low clay content as observed in facies 1. Clast size and angularity should therefore also be considered. Compact breccia sheets have maximum clast sizes of 2.5 cm which are moderate to subrounded (Stewart and Hancock, 1988), whereas cemented colluvium has much larger clast sizes which are more angular, indicating the short transport distances and physical weathering of the footwall limestone.
Table 2: Criteria for differentiating fault breccia (cataclasite) and cemented colluvium on the Lastros fault.

<table>
<thead>
<tr>
<th></th>
<th>Tectonic breccia (cataclasite)</th>
<th>Cemented colluvium</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Compact breccia</td>
<td>Incohesive breccia</td>
</tr>
<tr>
<td><strong>Form</strong></td>
<td>Sheet</td>
<td>Belt</td>
</tr>
<tr>
<td><strong>Thickness</strong></td>
<td>Centimetres</td>
<td>Metres to decimetres</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Matrix composition</strong></td>
<td>Granulated carbonate fragments in a calcite cement</td>
<td>Often no matrix present</td>
</tr>
<tr>
<td></td>
<td>A thin (up to 2 mm) clay-rich gouge layer containing sub-mm clasts of calcite and limestone can coat the primary slip surface</td>
<td>High proportion of voids are located in the fractures between clasts</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Proportion of fine grained matrix</strong></td>
<td>High</td>
<td>Low</td>
</tr>
<tr>
<td><strong>Clast lithology</strong></td>
<td>Carbonate clasts</td>
<td>Carbonate clasts</td>
</tr>
<tr>
<td><strong>Clast packing</strong></td>
<td>Mainly matrix supported</td>
<td>Clast supported</td>
</tr>
<tr>
<td><strong>Clast sorting</strong></td>
<td>Poor</td>
<td>Moderate, preferred orientation of fractures</td>
</tr>
<tr>
<td><strong>Clast angularity</strong></td>
<td>Moderate to subrounded</td>
<td>Subrounded to angular</td>
</tr>
<tr>
<td><strong>Clast size</strong></td>
<td>0.1 - 2.5 cm</td>
<td>0.5 - 10.5 cm</td>
</tr>
<tr>
<td><strong>Clast boundary matching</strong></td>
<td>Low</td>
<td>Moderate – high</td>
</tr>
<tr>
<td><strong>Location and structure</strong></td>
<td>Well cemented 3-50 cm thick sheets occurring adjacent to primary slip plane beneath fault gouge (where present)</td>
<td>1-3 m wide breccia belts located beneath compact breccia sheets</td>
</tr>
<tr>
<td></td>
<td>Sheets can be foliated containing minor shear planes</td>
<td>Subsidiary slip surfaces can be present which lack gouges or cataclasites</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

- Tectonic breccia attributes are mainly from Stewart and Hancock (1988, 1990) who studied normal faults in carbonate rocks in the Aegean; further tectonic breccia characteristics are from Bussolotto et al. (2007) and Bullock et al. (2014) who both studied normal faults in carbonate rocks in Italy.
5.6.2 Diagenetic interpretation of cement and lobe internal structure

The cement from facies 1 has a minor constituent of mud/clay and the predominantly clast supported nature of the facies is indicative of a rockfall deposit which later became cemented due to the precipitation of calcite below ground. The minor constituent of mud/clay that is present in facies 1 most likely comes from pedogenic clays and silts, which formed after rockfall deposition and were washed into an originally porous gravel skeleton. The higher content of clay/mud in facies 2 and 3 are representative of debris flow deposits (e.g. Sanders et al. 2010) where fines were incorporated into the cemented colluvium gravel matrix by the re-working of pedogenic clays and silts during mobilisation and deposition (Flügel, 2010). After deposition on the hanging-wall, calcite cement then precipitated in the pore spaces of the matrix.

The macroscopic and microscopic results from facies 1, 2 and 3 show that there is no significant difference between the calcite within the cement. Sparry calcite is present in all three facies and it is only the size of the crystals that differ; with increasing clay content in facies 2 and 3 the calcite crystal sizes drop significantly reflecting the lack of void space for them to grow. Stable isotope values of $\delta^{13}$C and $\delta^{18}$O are also very similar for all three facies (Table 1; Table 3) and the values indicate that the source water has a meteoric origin and the deposit can be classified as a meteogene. The average $\delta^{13}$C value for all three facies is -7.95 ‰ (Table 3) which falls within the typical range for meteogenes; theremogenes have significantly higher $\delta^{13}$C values averaging at 3.89 ‰ (Pentecost, 2005). Dissolved inorganic carbon in groundwater is derived from two sources: from dissolution of organic carbon in the soil zone, and from dissolution of inorganic carbon from carbonate aquifer rocks. Shallow groundwater leaving the soil zone is highly depleted in $\delta^{13}$C values (around -25 ‰; Deines et al., 1974), whereas deeper groundwater has $\delta^{13}$C values around 0 ‰. The average $\delta^{13}$C value of -7.95 ‰ for all three facies suggests that either the source spring water flowed through a portion of soil before precipitation of the calcite cement, or rainwater has percolated through the soil zone and mixed with groundwater before calcite precipitation. In either case this indicates that the cement formed from the dissolution of limestone in meteoric groundwaters charged with soil zone CO$_2$. The gravel clasts within the cemented colluvium show no visible signs of dissolution, which implies that the calcium carbonate within the parent water was not derived from the host sediment. The calcite vein from within the colluvial sheet has lower $\delta^{13}$C and $\delta^{18}$O values than all the lobe facies (Table 3). However, the values still fall within the typical range for meteogenes (Pentecost, 2005).

The form of calcite crystal can be used to help identify the hydrological environment in which they precipitated. Drusy calcite crystals increasing in size towards the centre of the void are
observed in all three facies and are typical of saturated zones. This would imply a phreatic environment, but saturated conditions can also occur in the vadose zone (Flügel, 2010) where there is a high water supply. There is both primary and secondary porosity within the analysed cements. Primary porosity is observed microscopically as remnant pores/voids which were not completely filled by calcite. Sparry calcite crystals often surround these primary pores (Figs. 5.6d,g; 5.7b), which are indicative of the cessation of saturated conditions. Secondary porosity is observed as larger pores/voids which can be observed macroscopically, which represent vadose eluviation and/or dissolution of the matrix (Figs. 5.5a,c; 5.6a,c,e; 5.7a,c).

When $\delta^{18}O$ concentrations of the facies cement are compared with contemporary spring water near Aghios Nikolaos (Fig. 5.1b) there is some enrichment within the facies cement. Leontiadis et al. (1988) analysed 19 samples of spring water and determined a mean $\delta^{18}O$ concentration of -7.23 ‰. The average $\delta^{18}O$ for all three cement facies is -5.01 ‰. This indicates that some evaporation of the groundwater is likely to have taken place prior to calcite precipitation. Evaporation is caused by surface and soil zone exposure and preferentially affects the lighter $^{16}O$ isotopes leading to enrichment in $\delta^{18}O$. Furthermore, the majority of the samples by Leontiadis et al. (1988) were taken from higher altitudes than the lobe. There is progressive depletion of $\delta^{18}O$ in rainwater at higher altitudes leading to more negative values compared with lower altitudes (Dotsika et al. 2010). The temperature of the water will also have an effect on the $\delta^{18}O$ concentrations, and oxygen isotope fractionation in speleothems has been shown to cause an increased oxygen isotope value with decreasing temperatures (Mühlinghaus et al., 2009). Palaeotemperature calculations of the parent water that formed for lobe cement are presented below in the section Palaeotemperature.

The three different facies within the lobe are representative of the mass wasting process that formed them, with facies 1 representing a rockfall deposits and facies 2 and 3 representing debris flows. The facies would have originally been deposited in layers close to parallel with the hanging-wall surface. The dip angle of the lobe facies is hard to determine due to the cemented nature of the deposits and the high density of vegetation covering the surface. However, there are many hanging-wall catchment gullies on the Lastros fault (Fig. 5.3a,b) where the dip of the colluvial layers can be accurately measured; this angle can also be observed in the GPR profiles (Fig. 5.9a,b). The average dip of the hanging-wall colluvium is 24° and assuming this average for these facies layers allows the internal structure to be visualised. Figure 5.10a shows a cross section through the full extent of the lobe and figure 5.10b shows the sampled area in detail. The facies layers are all >1 m in thickness with the
stratigraphically lowest facies 2 deposit considerably thicker. It should be noted that a further division of these layers would be likely with continuous sampling along the profile.

The above observations, in combination with localised position of the cemented colluvium lobe with little lateral extension, clearly indicate that the cementation occurred in vadose zone conditions with the meteoric source water coming from a spring at the fault plane. The spring formed locally saturated zones in the vadose environment. This localised outflow allows cement to be precipitated at the source and immediately downstream of it, reflecting the channelling of flow in the vadose zone (Halley and Harris, 1979) and formation of the lobe.

Figure 5.10. a) Cross section through the lobe (see Fig. 5.3c for location). b) Cross section of sampled area showing the locations of the three facies; numbers represent samples taken at distances from the fault plane.
The lobe could be classified as a type of travertine. Altunel and Hancock (1993) describe similar hanging-wall morphologies from the Pamukkale fault in western Turkey. The authors name locally cemented footwall derived talus as ‘range-front travertines’ deposited from spring water. They describe these morphological structures as heavily denuded and provide a provisional U/Th date of 66,000 ± 5,600 years. Martinez-Diaz and Hernandez-Enrile (2001) also describe ‘range front travertines’ at the Alhama de Murcia fault in the Betic Cordillera, Spain. Here the fault allows spring-water to issue out onto the slope cementing alluvial deposits. These structures in both Turkey and Spain are, however, formed from cement precipitation in thermal waters making them thermogenes.

Table 5.3. Average stable isotope values for facies 1, 2 and 3

<table>
<thead>
<tr>
<th>Facies</th>
<th>No.</th>
<th>δ13C (% VPDB)</th>
<th>δ18O (% VPDB)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Facies 1</td>
<td>11</td>
<td>-8.73</td>
<td>-5.00</td>
</tr>
<tr>
<td>Facies 2</td>
<td>22</td>
<td>-7.06</td>
<td>-4.95</td>
</tr>
<tr>
<td>Facies 3</td>
<td>10</td>
<td>-9.07</td>
<td>-5.16</td>
</tr>
<tr>
<td>All facies</td>
<td>43</td>
<td>-7.95</td>
<td>-5.01</td>
</tr>
<tr>
<td>Calcite vein</td>
<td>8</td>
<td>-10.96</td>
<td>-5.91</td>
</tr>
</tbody>
</table>

Values from secondary void edges and rims around dolomite clasts are excluded

5.6.3 Palaeotemperature

The temperature of the fluid from which the calcite cement precipitated can be calculated to infer what the climatic conditions were like at the time of precipitation. These palaeotemperature calculations assume isotopic equilibrium conditions. This means there is isotope exchange between the calcite and water phases (Pentecost, 2005). Equilibrium isotope fractionation effects are temperature dependant and affect mainly the oxygen isotope composition, usually through degassing; fractionation effects of carbon isotopes are rather minor and only weakly temperature dependant (Letch, 2014). For the lobe cement, excluding secondary void areas, there is a poor correlation between the δ13C and δ18O values (Fig. 5.8 solid squares) indicating that degassing is not a major factor (Hendy, 1971). Furthermore, when δ18O values are plotted against distance to the fault plane (Fig. 5.11) there are only minimal changes up to 40 m. An increase in δ18O at 51 m may be indicative of δ18O-depleted CO2 degassing; therefore up to 40 m from the fault plane we can assume that the cement precipitated under isotopic equilibrium conditions and parent water palaeotemperature calculations can therefore be undertaken.

There are several different equations in the literature that can be used to calculate the temperature of the precipitating fluid. We present the results using two equations:
1) Hays and Grossman (1991) used meteoric cements to formulate the following equation:

\[ t \ (° C) = 15.7 - 4.36(\delta_c - \delta_w) + 0.12(\delta_c - \delta_w)^2 \]

where \( \delta_c \) is the oxygen isotope composition of the calcite measured in VPDB, and \( \delta_w \) is the oxygen isotope composition of water measured in VSMOW (Vienna Standard Mean Ocean Water).

2) Kim and O’Neil (1997) used non-marine carbonates based on the earlier work of O’Neil et al (1969) and Freidman and O’Neil (1977) to determine the following equation:

\[ 10^3 \ln \alpha_{c-w} = (18.03 \times 10^3) \div T - 32.42 \]

where \( T \) is temperature measured in degrees Kelvin, and \( \alpha_{c-w} \) is the oxygen equilibrium fractionation factor between calcite and water defined by:

\[ \alpha_{c-w} = \frac{1000 + \delta^{18}O_c}{1000 + \delta^{18}O_w} \]

where \( \delta^{18}O_c \) and \( \delta^{18}O_w \) are the oxygen isotopic compositions of the calcite and water both measured in VSMOW.

Equations 1 and 2 both assume knowledge of the oxygen isotopic composition of the parent water from which the calcite precipitated, but this is unknown. However, investigations by Leontiadis et al. (1988) determined that the \( \delta^{18}O \) concentration from 19 samples of spring water near Aghios Nikolaos (Fig. 5.1b) have a mean value of -7.23 ‰. It is likely that the groundwater during the Pleistocene interglacials had a similar composition, and as Aghios Nikolaos is in close proximity to the Lastros fault one can accept this mean value as representative.

The results of these calculations based on the above values are shown in Table 4. Secondary voids areas have been excluded as the possibility of secondary precipitation renders their interpretation rather ambiguous. Using equation 1 the average crystallisation temperature for all cement samples is 7°C and there is little difference between the three facies (Table 4). Using equation 2 the average temperature of all cements drops to 4°C. Groundwater temperatures are generally equal to the average annual air temperature above
the land surface. The average air temperature for Crete at 2 m above ground was 17.5°C between 1961 and 1990 (BoG, 2011). As the altitude of the cement samples are from 460 to 480 m, the average air temperature must be corrected for altitude for a comparison to be made with water precipitation temperature. Assuming a lapse rate of 0.7°C for every 100 m in altitude (Flocas et al., 1983) the current average air temperature at the altitude of the cement samples is 14°C. Therefore, the parent water palaeotemperature results indicate a climate between 7 and 10°C cooler than present day. However, there is a possibility that the actual δ¹⁸O of the parent water was lower than the assumed value of -7.23 ‰ used in the calculations. When the δ¹⁸O of the parent water is decreased by 0.5 ‰ there is a 2°C increase in palaeotemperature and this possibility cannot be excluded.

Cement from facies 1 was tested for possible U-Th dating. Unfortunately the mud/clay content within facies 1, although low, contained too much thorium to allow an accurate date to be determined. Therefore, a Pleistocene marine isotope stage (MIS) cannot be ascribed to the cement. Interglacial conditions are suggested by the relatively low values of δ¹³C (Table 1; Table 3) indicating groundwater charged with biogenic CO₂ (Baker et al., 1997; Gradinski et al., 2014), and the presence of the mud/clay within the cement matrix implying the presence of soil cover during colluvial deposition and cementation.

On Crete periglacial processes are ongoing on the highest peaks (Hughes et al., 2006) but no glaciers exist today. However, there is evidence of glaciers on Crete in the Pleistocene (Nemec and Postma, 1993; Fabre and Maire, 1983) and they were the southernmost glacial landforms in Europe (Poser, 1957; Boenzi et al., 1982; Fabre and Maire, 1983). Kuhlemann et al. (2008) suggest that the air temperature around Crete during the last glacial maximum (23-19 ka) was 6.5 to 7° lower than today, but the sea temperature was only between 3 and 4° lower. The relatively warmer sea temperature compared to the air would have promoted the convection of moist air and high precipitation rates. The highest peak on Crete is over 2400 m in height and located within the White Mountains in west of the island (Fig. 5.1b). Alluvial fans were formed in this region due to ice cap melting and discharge of large amounts of meltwater. Five alternate periods of fan growth and abandonment are considered to be related to periods of deglaciation and re-glaciation (Nemec and Postma, 1993). Pope et al. (2016) undertook a sedimentological and palaeoclimatic study of the Sphakia alluvial fan in this region carrying out OSL and U-Th dating on several segments of the fan. The U-Th dates from carbonate cements within the representative fan sequence are 123.9 ± 7.4, 75.4 ± 2.9, 70.9 ± 1.0 and 72.3 ± 6.1 ka; one sample of pedogenic calcrete was dated at 59.0 ± 2.3. High groundwater levels were therefore causing cementation at these times, mostly between 70 and 75 ka which was the transition from MIS 5a (warm climate) to MIS 4 (cold
climate). Pope et al. (2016) also show that alluviation of the fan was occurring during most climatic settings (interglacial, interstadial and stadial episodes) throughout the last interglacial/glacial cycle due to high sediment supply from the catchment in the White Mountains (Fig. 5.1b). Further glacial features have been found on Mount Ida in the centre of the island (Fig. 5.1b). Here a cirque and associated moraines at an altitude of 1945 m have been identified (Fabre and Maire, 1983).

Table 5.4. Crystallisation temperatures of calcite cement

<table>
<thead>
<tr>
<th>Sample no.</th>
<th>Distance from fault (m)</th>
<th>Facies</th>
<th>$\delta^{18}O$ (% VSMOW)</th>
<th>Calc. 1 T(°C)</th>
<th>Calc. 2 T(°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>LL01</td>
<td>9</td>
<td>1</td>
<td>25.309</td>
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- Calc. 1 is from Hays and Grossman (1991)
- Calc. 2 is from Kim and O’Neil (1997)
- $\delta^{18}O$ concentration of parent water is assumed to be -7.23 ‰ (Leontiadis et al. 1988) for both calculations. If varied by ±0.5 ‰ there is a ±2°C change in the derived palaeotemperatures for both calculations.
The Kapsos Mountain comprising the footwall block of the Lastros fault has a maximum altitude of 980 m. It is therefore too low to have once hosted a permanent ice cap during Pleistocene glacial times and there is no geomorphological evidence for this. However, the parent water palaeotemperature calculations (Table 4) suggest there would have likely been a snow cap on the mountain in winter months. The lobe cement precipitated at a maximum altitude of 480 m and the top of the mountain is 500 m; therefore, average annual temperatures on top of the mountain at the time of cement precipitation were between 0°C (Calc. 1) and -3°C (Calc. 2) assuming lapse rate of 0.7°C for an increase of 100 m in altitude (Flocas et al., 1983). High precipitation rates throughout the year and seasonal melting of the winter snow and would have led to the high groundwater levels needed for spring activity at the fault plane and subsequent cementation. This is likely to have occurred during a transitional period soon after a glacial maximum, most likely following a Heinrich Event (meltwater influx) (Dansgaard et al., 1993) in the Late Pleistocene. However, as Pope et al. (2016) have shown cementation is also occurring on Crete during the transition between warm to cold climates where there are high altitude catchment areas. Seasonal changes in palaeoclimate have also been observed after the LGM (Last Glacial Maximum) by Kontakiotis (2016) who analysed planktonic foraminifera and their stable isotope concentrations from sea cores in the northern and southern Aegean Sea. Kontakiotis (2016) determined that the seasonal changes in palaeoclimate were most pronounced during the transition from LGM to deglaciation between 15.1 and 12.9 ka. This was caused by sea surface temperature variability in the Aegean, where humid air from Africa increased precipitation and river runoff. It is therefore likely that seasonal hanging-wall colluvial cementation was occurring following and perhaps preceding most deglaciation stages in the Late Pleistocene, with the cold temperatures controlling cement precipitation. Cemented colluvium is therefore likely to be present deeper in the hanging-wall having been formed at earlier deglaciation stages and then subsequently buried.
5.7 Cemented colluvium formation and fault evolution

The sheets and lobes of cemented colluvium are clearly visible in the fault zone morphology. However, the GPR data (Fig. 5.9a,b) shows that there is cemented colluvium present in the hanging-wall subsurface below uncemented colluvium. In all three GPR profiles the cemented colluvium (RF4 and RF5 in Fig. 5.9a,b) stops around 20 m downslope of the scarp; this is particularly evident in profiles 2 and 3 where there is a sharp boundary between the cemented and uncemented colluvium. This boundary either represents the downslope extent of vadose zone seasonal cementation formation, or the presence of a secondary fault in the hanging-wall. There is no topographic expression for a secondary fault, but this may be due to the construction of ancient anthropogenic terraces which affected the uppermost 1 m of the hanging-wall. Nevertheless, the presence of cemented colluvium in the subsurface below uncemented colluvium shows that throughout the Pleistocene the surface springs at the fault plane were not restricted to the areas where cemented colluvial deposits are currently exposed at the surface. Groundwater was seasonally high enough to form large cemented areas along strike, most likely through multiple springs. Fault activity then most likely disrupts the groundwater flow path (e.g. Michetti et al., 2007) causing the cessation of some springs; colluvial deposition then continues burying the cemented deposits.

Figure 5.11. $\delta^{18}O$ values of cement plotted against distance to the bedrock fault. Secondary void areas and rims are excluded.
The proposed model for the temporal development of the fault and formation of the cemented colluvium is presented in figure 5.12. After fault initiation (Fig. 5.12a) the hanging-wall block was continually downthrown during successive ruptures throughout the Pleistocene and the volume this created was filled with colluvium (Fig. 5.12b). Freeze thaw physical weathering in the glacial periods throughout the Late Pleistocene helped break and dislodge the brecciated footwall limestone (Tucker et al., 2011). Mostly gravel sized limestone fragments in rockfalls and debris flows were then deposited on the hanging-wall (Fig 5.12b). The rate of colluvium deposition was higher than the fault’s slip rate and no bedrock scarp was exposed. During Late Pleistocene transitions towards interglacial periods soil cover developed and seasonal snowmelt and precipitation raised groundwater levels and pore pressures. Due to the preferential pathway of the fault, groundwater then began issuing out into the hanging-wall at localised positions along strike (Fig. 5.12c). The brecciated footwall limestone may have also contributed to preferential fluid flow. At these spring positions calcite cement then precipitated in-between the colluvial gravels. During gravel cementation the colluvium also fused to the bedrock footwall. In Late Pleistocene interglacial periods (Fig. 5.12d) the increased temperature caused the water table to drop and spring activity ceased. Fault ruptures continued and the sedimentation rate of hanging-wall colluvium was lower than the fault’s slip rate allowing the bedrock fault plane to be exposed (Fig 5.12d). Soil cover reduced due the development of semi-arid conditions. During each rupture the cemented colluvium is uplifted along with the footwall exhuming the areas of hanging-wall where cementation has occurred. Repetition of glacial to interglacial phases (Fig. 5.12b – 5.12d) are then likely to have occurred. In the Holocene (Fig. 5.12e) fault ruptures continued to expose the fault scarp and cemented areas causing them to grow in height. Farming during historical times (Fig. 5.12f) then further exhumed the lower parts of the fault scarp by terrace construction to utilise water runoff. This also increases the height of colluvial sheets and increases the exposed size of the cemented lobe.
Chapter 5: Hanging-wall colluvial cementation along active normal faults

Figure 5.12. Schematic diagrams showing phases of fault evolution and formation of the cemented colluvium. a) Carbonate rock of the Mani geotectonic unit is fractured during the initiation of faulting. b) Footwall derived colluvium is deposited on the hanging-wall by rockfall (RF) and debris flows (DF) predominantly during glacial conditions. No fault scarp is exposed; c) Seasonal groundwater percolates through brecciated bedrock and the fault springing at localised positions along the footwall—hanging-wall boundary during a transitional period soon after a glacial maximum. Calcite precipitates between the hanging-wall colluvial gravel clasts at these positions. Erosion of the footwall and sedimentation on the hanging-wall reduces and the fault scarp begins to be exposed by recurrent earthquakes; d) Higher temperatures and a dry climate during interglacial conditions lower the groundwater table reducing spring activity and cement precipitation. The fault scarp continues to be exposed by recurrent earthquakes. Cemented colluvial sheets and lobes have become fused to the footwall and are also consistently uplifted during these extension events. Possible repetitions of phases b to d then occur; e) During the Holocene the fault scarp and cemented colluvial sheets and lobes grow from continued exposure by recurrent earthquakes; f) Man made terraces are constructed in the hanging-wall below the fault scarp during historical times further exposing the bedrock scarp.
Chapter 5: Hanging-wall colluvial cementation along active normal faults

5.8 Conclusions

Cemented colluvium located on the hanging-wall of the Lastros fault forms sheets and lobes often attached to the fault plane. The cemented colluvium can be described as a breccia with predominantly limestone clasts within a carbonate cement matrix which contains varying amounts of mud/clay. This mud/clay content in combination with clast size and angularity can be used to distinguish between cemented colluvium, derived from rockfalls and debris flows, and fault cataclasite. The stable isotope values of carbon and oxygen from the cement show that the cement formed from the dissolution of limestone in meteoric groundwaters charged with soil zone CO₂. The deposit can therefore be classified as a meteogene. The localised positions of the cemented colluvium along the fault are indicative of individual springs issuing water into the hanging-wall. Palaeotemperature calculations of the parent water from which the cement precipitated indicate a climate between 7°C and 10°C colder than today’s average annual temperature on Crete when corrected for altitude. This is indicative of a transitional period between glacial and interglacial conditions, most likely in the Late Pleistocene. Cemented colluvium is also present in the shallow subsurface of the hanging-wall below uncemented colluvium indicating an earlier phase of spring activity and high groundwater levels. The distribution of cemented colluvium and the stable isotope values can be used as a proxy for climatic conditions and landscape evolution.
Chapter 6: The Kaparelli fault trench

This chapter presents the results of a palaeoseismological trench log undertaken in 2014 on the Kaparelli fault, eastern Corinth Gulf. This work was undertaken for the development of a new palaeoseismological method to visualise the trench wall stratigraphy in 3D. For further information Schneiderwind et al. (2016) should be consulted.


6.1 Rationale

A new palaeoseismological method developed by Schneiderwind et al. (2016) at RWTH Aachen University uses photographs, t-LiDAR and GPR on palaeoseismological trench walls to visualise the trench wall stratigraphy in 3D to a depth of approx. 3 metres. This method was developed on the Sfaka fault road cut trenches (see Chapter 4), but the method needed to be verified at another site. Kokkalas et al. (2007) undertook a trenching investigation at the Kaparelli fault, which last ruptured during the 1981 Corinth earthquake series with a magnitude (Ms) of 6.4. The trenches that Kokkalas et al. (2007) excavated are still open which gave us the opportunity to verify the 3D visualisation method.

So that the GPR and t-LiDAR can be accurately interpreted, a good trench log is needed which accurately maps all the features contained on the trench wall. Using the trench log figure presented in Kokkalas et al. (2007) this was not possible; the logged stratigraphic detail was not sufficient for our purposes. Therefore, a complete re-logging of Trench 1 was undertaken. This included cleaning of the trench walls from weathering effects, attaching a 1 x 1 m string line grid, and sketching and describing the stratigraphy. Figure 6.1 shows a comparison of the two logs.

6.2 Trench stratigraphy

Figures 6.2 and 6.3 show the photomosaic and trench log respectively. Stratigraphic labelling was undertaken using the following system: The capital letter of the single units indicate the material (fines, sand, gravel), the number represents different units and the lower case letter (a, b, etc.) indicate a slight difference between the unit with the same capital letter and number. For example unit F7a is very similar or identical to unit F7, but has larger siltstone nodules. The trench was excavated along a slope with an angle between 13-23° through the 1981 surface rupture of the Kaparelli fault exposing a 32 m long, 3 m wide and 2 – 4 m deep...
section. The fault in the trench wall is exposed for 3.7 m and dips 70-80° to the south, creating a 2 m wide fault zone (Fig. 6.4).

The footwall (m 27-32) in northern part of the trench comprises a multi-coloured pebbly-cobbly gravel deposit (unit G14) with clasts ranging from a few mm up to 50 cm in a cemented silty matrix (10-20%). The clasts are sub-angular to well-rounded ophiolites and limestones. The deposit contains two subunits (G14a/b) with a different clasts size than the main unit G14, which are possibly sheared and cut by two smaller north dipping faults. One of the north dipping faults is 1.28 m long and dips 60° to the north, and the other one is 84 cm long with a shallower dip of 40° to the north. In the southern part of the footwall (m 27) a gradual boundary between the gravel bed G14 and a brownish yellow clayey silt (unit F1) occurs. The gravel and silt beds steeply dip to the south around 70°. The brownish silt bed (unit F1) is overlain by a pink 10-33 cm thick silt with rare siltstone clasts, that is deformed up.

Figure 6.1. Comparison of the trench logs: a) Kokkalas et al. (2007); b) our re-logging in 2015; gravel features are shown in various shades of blue and have a radial nature (see Fig. 6.3 for larger image).

The footwall (m 27-32) in northern part of the trench comprises a multi-coloured pebbly-cobbly gravel deposit (unit G14) with clasts ranging from a few mm up to 50 cm in a cemented silty matrix (10-20%). The clasts are sub-angular to well-rounded ophiolites and limestones. The deposit contains two subunits (G14a/b) with a different clasts size than the main unit G14, which are possibly sheared and cut by two smaller north dipping faults. One of the north dipping faults is 1.28 m long and dips 60° to the north, and the other one is 84 cm long with a shallower dip of 40° to the north. In the southern part of the footwall (m 27) a gradual boundary between the gravel bed G14 and a brownish yellow clayey silt (unit F1) occurs. The gravel and silt beds steeply dip to the south around 70°. The brownish silt bed (unit F1) is overlain by a pink 10-33 cm thick silt with rare siltstone clasts, that is deformed up.
and cut by the fault. The offset is up to 57 cm between the upper layer boundaries and around 37 cm between the lower ones.

The hanging-wall (m ~ 25-0) comprises loess type silt with close to vertically oriented gravel structures of various sizes. This silt deposit contains whitish siltstone nodules and calcrite with increasing frequency downslope. The gravel structures basically comprise gravel in a silty sometimes sandy matrix with clasts ranging from few mm up to 13 cm. In the northern part of the hanging-wall close to the fault zone, the gravels are poorly cemented and they become more cemented further south. The gravel structures are up to 1 m in width and some horizontally sheared (m 20 and 16-15). Some of these gravel structures break through the silt deposit up to the recent topsoil (unit TS1) and some of them do not reach the topsoil. At the southern end of the trench these gravel structures are almost perpendicular to the surface, but closer to the fault zone the angle decreases (Fig. 6.3). Several gravel structures include voids from either animal burrows or where the loose material has been eroded away since initial excavation by Kokkalas et al. (2007).

The fault dips 70-80° to the south creating a 2-3 m wide and 3.5 m high fault zone (m 27-25; Fig. 6.4) that was reactivated in 1981. It shows several cracks and fissures that were filled with surrounding and overlying material. The fault cuts through a pink, 10-33 cm thick silt layer (unit F2), that is interrupted by a fissure filled with a brownish yellow slightly clayey silt layer (unit F3) that is bedded above. This fissure/crack fill is wedge shaped, about 108 cm length with a maximum width of 50 cm tapering downwards. The displacement of the upper boundaries of unit F2 is around 57 cm and 37 cm between the lower ones. An old topsoil up to 22 cm in thickness (unit TS2) overlies fill unit F3, and is warped down towards the fault and buried around 1 m below the surface. On top of TS2 there is an assemblage of pale brown silt (units F4, F5) with few rootlets and up to 20% clasts that seems to be partly dragged along the fault. These are overlain by the colluvial wedge (unit CW) and material of a remnant soil (unit RT1) buried after the 1981 earthquake. About 1.5-2 m southwards from the fault a small north dipping (60°) antithetic fault is present. At meter 24 m this fault displaces a small sand unit (unit S9) by approximately 5 cm and several small parallel cracks are visible downslope. Interestingly the direction of shear, observed through offset units S9 and F7, indicates compression rather than extension (m 24, Figs. 6.2 and 6.3).

The uppermost layer exposed in the trench is a 5-25 cm thick strong brown clayey silty recent topsoil (unit TS1) containing clasts ranging from gravel to cobble (max. 15 cm) and abundant roots. The topsoil ranges over the full length of the trench and contains many rootlets from the overlying vegetation. See Appendix B for full descriptions of all soils within the trench.
Figure 6.2. Kaparelli trench photomosaic produced for Schneiderwind et al. (2016).
Chapter 6: The Kaparelli fault trench

Figure 6.3. Kaparelli trench log produced for Schneiderwind et al. (2016). See Appendix B for unit descriptions.
6.3 Palaeoseismic interpretation of gravel structures

The fault zone stratigraphy (Fig. 6.4) is broadly similar to that described by Kokkalas et al. (2007). However, Kokkalas et al. (2007) describes the many coarse grained structures located within the hanging-wall of the trench as tension fissures, which formed by the filling of fissures with gravel during individual earthquakes. For this to happen the fissure must be open and gravel must be flowing of the slope surface ready to fill them. However, some of these fill features do not reach the surface (Fig. 6.3). The soil in which these features are present is Loess which formed in the Pleistocene, and the hanging-wall slope surface is currently an erosional one with little soil development; soil only began forming with agricultural use of the land. Therefore, if these features are tension fissures and did not break the surface they would have to be very old and have formed during the Loess deposition in the Pleistocene. However, Kokkalas et al. (2007) provides young dates for these features of around 5000 BC. These features also show a radial pattern when viewed in the trench log which led us to think that they may have a liquefaction origin and may be clastic dykes injected from below. This radial pattern could not be observed in the Kokkalas et al. (2007) figures as the lowest 10 m of the trench log is not presented (Fig 6.1a). We
therefore took some samples of the gravel features for classification analyses and to compare with liquefaction gradation curves.

The grading curves shown in figure 6.5 indicate that the fill material is potentially liquefiable. Only the larger gravels sizes fall beyond the ‘potentially liquefiable’ curve. Each of the curves are quite similar to each other, indicating a common source bed. Liquefaction can occur in much coarser material as happened in the Borah Peak earthquake (Mₕ 7.3) in 1983, Idaho US (Fig. 6.5). Similar structures that are interpreted as clastic dykes caused by liquefaction have been described by Dechen et al. (2013) in a Quaternary loess type silt deposits in the Fangshan District in China. Here an exceptionally strong upward-directed flow caused by high hydrostatic pressure that built up due to impermeable silt material which was overlying a braided river gravel deposit. The injection caused clasts up to 30 cm in diameter to be preserved within the clastic dyke (Dechen et al. 2013).

Figure 6.5. Particle size distribution curves from the Kaparelli trench. See figure 6.3 for sample locations. Most liquefiable and potentially liquefiable curves as well as the curve from the Borah Peak earthquake are taken from McCalpin (2009).

One of the main factors for determining whether a deposit has a liquefaction origin is the magnitude that is needed to inject such coarse grains into a thick silt layer. Obermeier (2009) proposes that for large liquefaction-induced dykes the threshold magnitude is around Mₕ ~7. Observed magnitudes for gravel dykes caused by liquefaction are of high magnitude (> 7.0 Mₕ) (McCalpin, 2009; Dechen et al., 2013). The Kaparelli fault is relatively small and has an upper threshold magnitude of approximately Mₕ 6.4 based on empirical correlations (Wells...
and Coppersmith, 1994), but there are several larger faults/fault zones in the vicinity of the trench site which may have caused the ground acceleration necessary for gravel dyke injection; e.g. the Erithres-Dafnes Fault Zone is located only a few km to the southeast of the trench site and comprises at least 20 km of fault segments, or the South Alkyonides Fault System which is located only around 12 km to the south of the site and comprises over 35 km of fault segments. If the segments in these fault zones where to rupture together they are capable of producing high magnitude earthquakes. Furthermore, within the Gulf itself there are a huge amount of connected faults which are likely able to produce earthquakes capable of creating the ground shaking required for liquefaction.

This re-logging needed for the Schneiderwind et al. (2016) publication highlights the need for clear publication of findings. Although the larger structures in the Kokkalas et al. (2007) and our re-logging are comparable, the lack of detail and the poor resolution of the figures does not allow the reader to objectively examine their results. Furthermore, a possible liquefaction origin is not even mentioned in the publication. One reason for this is the assumption that bedrock limestone was reached at the bottom of the trench during excavation. This is only because the excavator could not dig any further; no limestone was observed. However, the fluvial gravels at the northern end of the trench have become well cemented and this cementation would significantly increase excavation difficulty. From our limited investigations, a liquefaction origin may be the source. A tension fracture formation process can effectively be ruled out for some of the gravel structures as they do not all extend completely through the loess and break the surface (Fig. 6.3). Further work is, however, needed on these gravel structures to definitively determine their origin.
Chapter 7: Discussion and conclusions

This chapter brings together the main conclusions of the research showing how the objectives set out in the scope have been met and discussing any uncertainties. A seismotectonic overview of the broader study areas (Peloponnese and Crete) is presented and suggestions are made for future research. The Kaparelli fault trench results are not further discussed here. For information on the outcome of the Kaparelli trench research see Schneiderwind et al. (2016).

7.1 Western Peloponnese

7.1.1 Fault investigations

The collection of field data on kinematic slip direction and fault geometry, combined with GIS analyses of footwall morphology, enabled a more accurate determination of fault segment lengths. The study found that both the Lapithas faults have kinematic indicators which vary in orientation along strike. Previous work in Greece and Italy have shown that many bedrock faults have kinematic indicators which have a mean dip-slip orientation at the centre of the fault, and a strike slip component towards the fault's tip (e.g. Roberts, 1996). This assumption of dip-slip movement and the centre of the fault and an increasing strike slip component towards the tips was used in conjunction with measurements of observable throw in GIS to accurately map the onshore lengths of the northern and southern Lapithas faults. Onshore lengths of 18 and 5 km for the northern and southern Lapithas faults were determined respectively. Using published bathymetry data, an offshore continuation of the northern Lapithas fault is likely, forming another segment of a fault array. The southern Lapithas fault most likely does not extend offshore as there is no clear evidence in the bathymetry data. These conclusions agree with those of Papanikolaou et al. (2007) who infer a 15 km offshore continuation of the northern Lapithas fault, and no offshore continuation for the southern Lapithas fault (Fig. 7.1).

Based on empirical calculations (Wells and Coppersmith, 1994) and when viewed separately, the onshore and offshore segments of the northern Lapithas fault could produce maximum earthquake magnitudes ($M_w$) of 6.5 and 6.4 respectively. A worst case scenario multi-segment rupture is capable of producing a maximum earthquake magnitude ($M_w$) of 6.9. As the southern Lapithas fault does not continue offshore, the total fault length is ca. 5 km and could produce a maximum magnitude ($M_w$) of 5.8.
During field data collection what was immediately noticeable was that preserved striations and clear corrugations were only present where anthropogenic activity had taken place, such as quarry excavations of hanging-wall colluvium or terrace excavations for agriculture. At all other locations clear kinematic indicators were not present. There are weathering structures covering the limestone scarp such as karren, solution pits and rills, which develop relatively quickly once the scarp is exposed. This then obscures and erodes the kinematic indicators that are mm-cm in depth. For further investigations on bedrock faults in the region, excavations in the immediate hanging-wall may be required to determine kinematics.

Figure 7.1. Synthetic geological map of the Kyparissiakos Gulf together with the offshore data. Probable correlations of the onshore with the offshore faults are shown with dashed lines. F-1FZ: Filiatra-1 fault zone, F-2FZ: Filiatra-2 fault zone, KFZ: Kyparissia fault zone, KNFZ: Kalo Nero fault zone, NFZ: Neda fault zone, LFZ: Lepreo fault zone, ZFZ: Zaharo fault zone, LaFZ: Lapihas fault zone, AFZ: Alfios fault zone, EFZ: Epitalio fault zone, VFZ: Vounargo fault zone (from Papanikolaou et al. 2007)
Figure 7.1 shows many of the faults in the western Peloponnese. The Lapithas faults are good examples of bedrock faults in the region, but there are also many more capable faults which all have potential for large earthquakes. Many of the coastal faults extend offshore into the Kyparissiakos Gulf making them significant structures capable of large earthquakes. However, the kinematics of the other bedrock faults in the region are largely unknown and exact fault lengths have not been determined. Systematic palaeoseismological investigations are needed on all of these faults to determine their activity and the threat they pose to local inhabitants (see section 7.1.4).

7.1.2 Archaeological damage and historical documentation

Seismic hazard assessment is predominantly based on historical and instrumental seismicity. However, the normal faults throughout the Aegean have low slip rates and long recurrence intervals; they may not have ruptured in the instrumental period and are often absent from historical catalogues. Earthquake recurrence intervals on individual faults are commonly several hundred to several thousand years, and therefore the time period covered by the historical record in many cases is shorter than the earthquake recurrence interval. However, many archaeological sites in Greece have evidence for earthquake damage and this can be studied to determine where the destructive seismic waves originated. Subsequent palaeoseismological investigations can then be undertaken on likely causative faults. This data can be used to improve the accuracy of seismic hazard assessments.

There is evidence for possible earthquake damage at both Samicum and Ancient Olympia. However, there are other explanations for the observed damage and these should not be ruled out until the damage is investigated further. The systematic cataloguing of the effects is needed to determine if structural damage has a similar orientation as this would be indicative of being produced from seismic waves. Using archaeological damage in earthquake studies has been successfully undertaken by many authors. Silva et al. (2009), for example, showed that studying simple statistical analysis of orientation data from the ancient Roman city of Baelo Claudia, Spain, could help determine the directivity of ground deformation. Furthermore, where event horizons are present and preserved within the archaeological remains, radio carbon dating ($^{14}$C) can be carried out on carbon rich materials. Grützner et al. (2010) successfully used this technique to date two paleoearthquakes in the Isis Temple in Baelo Claudia, Spain. However, there are many ambiguities within archaeoseismology and care must be taken not to misinterpret observed damage as EAE, as the damage could have been produced by other causes. Archaeological sites are all prone to natural disruptive processes that can mimic the expression of seismic rupture or shaking. Moreover, there are the vagaries of uncertain human action, such as questionable construction quality to the
potential for manmade destruction. EAE investigations, therefore, need to be complimented by other techniques such as paleoseismology in order to minimise the potential for misinterpretation.

What our research on Ancient Olympia has shown is that not all literature sources can be trusted. Just because a peer reviewed paper or book states an archaeological fact does not mean it is true. As stated in Chapter 3, the fallen columns of the Temple of Zeus have been attributed to earthquakes in either AD 522 and/or AD 551, and these dates are cited in countless publications (e.g. Stiros and Jones, 1996). However, the original source documents show that this interpretation is based on a rather ambiguous theory of Boetticher (1883). Furthermore, when historical earthquake records are consulted there are no reported damaging earthquakes in these years which can be attributed to the destruction of the temple. The original source material should therefore always be consulted and reviewed, preferably by a native speaker of the language used in the original document so that there are no misinterpretations caused by translation.

7.1.3 Seismotectonic overview

Definitive fault lengths and maximum possible earthquake magnitudes have been determined for northern and southern Lapithas faults, and this information can be used to increase the accuracy of seismic hazard assessments in the area. However, the whole of the Peloponnese can classified as having a high seismic hazard (Fig. 1.7) due to the numerous normal faults in the broader region and the subducting slab of the African plate. An indication of the seismic activity within the Peloponnese and offshore into the Ionian Sea is presented in Figure 7.2. Microseismicity up to M = 5 recorded by the SEAHELLARC amphibious seismic array (Papoulia et al., 2014) shows that most seismic activity is occurring offshore and is associated with thrusts relating to the the Hellenic trench (Fig. 7.2a). However, shallow onshore activity is also occurring which is associated with crustal extension. Fault plane solutions of selected earthquakes to the north of the Northern Lapithas fault (Fig. 7.2b) show overall NW-SE oriented extension. These may be associated with the Northern Lapithas fault or faults within the Olympia Basin (Fig. 7.1). Extension continues further to the south of the Kyparissiakos Gulf; however, more E-W oriented extension is occurring associated with the N-S striking normal faults west of Kalamata. This transition from Domain I in the north to Domain II in the south of the Kyparissiakos Gulf occurs at a similar latitude as the deepest recorded earthquakes by SEAHELLARC (Fig. 7.2; Papoulia et al., 2014). Here there are many hypocentres below 60 km compared with in the north around Zakynthos Island where all hypocentres are shallower than 30 km depth. This change in hypocentre depth is related
to the geometry of the subducting slab, which in turn may be the cause of the change in extension orientation.

Figure 7.2. a) Microseismicity recorded by the SEAHELLARC amphibious seismic array using a 3D independent velocity model defined from 2D active seismic experiments and 3D gravity modelling. b) Fault plane solutions for events with focal depths less than 15 km (yellow) and between 16 and 30 km (green) (modified from Papoulia et al., 2014).
The most recent damaging earthquakes that struck the Peloponnese occurred in Pirgos in 1993 and Movri in 2008 (see section 3.3). The Pirgos earthquake has been attributed to oblique normal slip on the Alfiossa fault (Koukouvelas et al., 1996) within the Alfios fault zone (Fig. 7.1). This shallow crustal extension occurs in the same domain (Fig. 1.4) as the Lapithas Mountain faults and is associated with the same deformation regime. The Movri earthquake, however, was caused by dextral slip along a high angle blind fault which caused a complex pattern of surface ruptures (Koukouvelas et al., 2010) and can be attributed to a slab fault caused by segmentation of subducting slab beneath the Peloponnese (Sachpazi et al., 2015). Sachpazi et al. (2015) undertook a study using P and S teleseismic waves from a dense seismological network to produce a high resolution 3D image of the subducting slab beneath the Peloponnese. The authors show that the subducting slab is segmented down to 100 km depth by NE-SW tending slab faults creating a series of slab panels (Fig. 7.3). The Movri earthquake occurred just above one of these slab faults and ruptured the lower crust and Moho of the upper plate, and is the likely cause of the complex surface ruptures observed. Nine slab faults are observed beneath the Peloponnese (Fig. 7.3) and they are likely the source of future destructive events. Furthermore, slab segmentation is also
shown to be the cause of intermediate depth slab earthquakes and the width of the slab panels (Fig. 7.3) may also limit the size of shallow megathrust earthquakes confining them to smaller magnitudes (Sachpazi et al., 2015). This 3D imaging shows the complexity of crustal deformation in the region. Surface faults in the study area mainly trend E-W and are associated with N-S extension, the same as is observed in the Gulf of Corinth (Fig. 1.4). However, due to subducting slab segmentation slightly deeper yet very destructive strike slip earthquakes can also be expected in the region.

7.1.4 Future research

Accurate determination of fault lengths and potential earthquake sizes is still needed for the majority of faults in the western Peloponnese, and the kinematics of these faults are largely unknown. A further and major requirement for these individual faults is the determination of earthquake dates. Only one palaeoseismological investigation has been carried out to date in the broader region (Zygouri et al., 2015). Accurate slip rates are unknown for the vast majority of faults. For the mountain front faults comprising bedrock footwalls and colluvial and/or marine sediments in the hanging-walls, such as the Lapithas faults, exposure dating with cosmogenic nuclides is likely the best method to obtain an accurate slip rate. Where faults are located in or cross sedimentary basins such as the Olympia basin (Fig. 7.1), trenching is the best method to apply. Here, trench site selection should focus on areas with sedimentation on both the footwall and hanging-wall. The three neotectonic trends (Domains I, II and III) of normal faults throughout the Aegean (see Section 1.1.4) warrant further investigation. It could be that one of these orientations is currently less or more active than the others. Comprehensive earthquake dates and slip rates on the individual faults are needed to determine this.

If potential EAE are found at archaeological sites in the region, a thorough check of historic reports of damage should be undertaken using the original source documentation. If a seismic source cannot be ruled out, archaeoseismological studies can determine the direction of seismic waves and then a short list of potential source faults can be determined. Palaeoseismological investigation can then be undertaken on these faults dating the most recent earthquakes and finding the causative fault.
7.2 Eastern Crete

7.2.1 Geomorphological investigations

Geomorphological analysis on the Lastros fault using t-LiDAR enabled an accurate digital representation of the scarp to be analysed over a large spatial area (1.3 km). The results show that previous slip rate estimations are too high. These postglacial slip rate estimates are the result of scarp height measurements in areas influenced by natural erosion and/or anthropogenic activity. A 100 m long section of scarp shows little influence from erosion or sedimentation. The natural variability of scarp height in this 100 m section is 8.3 to 10.7 m, which leads to corresponding slip rate range of 0.61 ± 0.14 mm/a to 0.79 ± 0.18 mm/a using a postglacial (15 ± 3 ka) date for first exhumation. The average scarp height of 9.4 m in this section is deemed representative and leads to a slip rate of 0.69 ± 0.15 mm/a.

There are some ambiguities to overcome when constructing profiles, both in the field and using the t-LiDAR derived DEM. As stated by Papanikolaou et al. (2005) and Papanikolaou & Roberts (2007), there is a possible error of up to 20% when constructing profiles because of local variations in footwall and hanging-wall morphology. Therefore, many profiles should be undertaken in representative areas (little erosion or sedimentation) and an average taken. Best practice would be to undertake both manual profiles in the field and DEM derived profiles. This will allow the digital data to be calibrated. In the 100 m long area deemed representative for slip rate calculations (Fig. 4.4 iii) there must be some erosion and sedimentation occurring. This is because when using 10 m wide polygons in GIS to calculate the scarp height we have scarp height variation which is 12% of the mean (range is 8.3 to 10.7 m). If narrower polygons were used there would likely be a larger variation in scarp height as small morphological changes will also be measured, but the mean scarp height value would remain the same. The most representative polygon width to use to analyse scarp height variation will depend on the goals of the research: to study the effect of erosion on scarp morphology narrower polygons are suggested - detailed analysis of the brecciated bedrock would also be required for this as fracture orientation have a large effect on footwall erosion; to study long-term slip rates 10 m wide polygons produce accurate results.

The Lastros fault scarp forms two segments: the northern and southern segments. The t-LiDAR scan only covered the northern segment; the southern segment remains unanalysed and there is a possibility of larger scarp heights since it located at the centre of the fault (Fig. 4.3a). The southern segment could not be scanned because of the oblique scanning angle caused by the steepness of the hanging-wall. Therefore, a UAV (unmanned aerial vehicle)
survey using photographs and structure from motion software would be most appropriate at this location.

The biggest uncertainty regarding the fault scarp analysis is the date of first exhumation, as this date determines the slip rate of the fault. An error has been presented in this work based on the $15 \pm 3$ ka theory for first exhumation, which is mainly derived from work in the Italian Apennine Mountains. There is confidence of this $15 \pm 3$ ka figure among the scientists working in the Apennines because there is now a wealth of fault scarp exposure data using cosmogenic nuclides. However, the climate of the Italian Apennines and Crete are quite different. Crete is over 700 km further south than the Apennines and the end of the Last Glacial Maximum (LGM) is likely to have occurred at a different time. Also, cosmogenic nuclide investigations on the Sparta fault in the Peloponnese (Benedetti et al., 2002) show that the top of the scarp has an exposure date of ca. 13 ka, and for the Kaparelli fault in the Gulf of Corinth (Benedetti et al., 2003) the top of the scarp has a date of ca. 20 ka. As the upper degraded part of the scarp is not sampled, post LGM first exhumation is likely older then these dates. The large difference between the exhumation dates of these faults shows that even in areas of similar climatic conditions there may still be significant differences in dates of first exhumation. Therefore, cosmogenic nuclide investigations are needed on Crete to determine whether the $15 \pm 3$ ka figure used in this research is representative, and even reduce the error.

**7.2.2 Palaeoseismological investigations**

Radiocarbon dating of fissure fills in trenches 1 and 2 on the Sfaka fault determined four earthquake events having maximum ages of $16055 \pm 215$ BP, $13168 \pm 107$ BP, $9484 \pm 63$ BP and $6102 \pm 113$ BP for events $w$, $x$, $y$ and $z$ respectively. Recurrence intervals therefore range between 2887 and 3684 years. However, these earthquake ages not well constrained. The samples used were bulk soil which allow for uncertainties in the determined ages. There is an inherited $^{14}$C signature within the soil which filled the fissures after the surface rupturing earthquake; therefore the dates presented are maximum ages. Also, the presence of roots within the fissure fills may have introduced younger materials into these deposits (Walker, 2005) which would then cause an underestimation of the $^{14}$C age. However, if both of these processes (inheritance and the root influence) have occurred there should be a much larger $2\sigma$ error produced in the calibrated ages. The errors are in fact quite small, between 63 and 215 yrs. Therefore, it is likely that the influence of the roots is negligible and the presented dates are in fact maximum ages.
According to the investigations at trenches 1 and 2 there have been no surface rupturing events since $6102 \pm 113$ BP. With the very regular recurrence interval between 2887 and 3684 years, there should have been at least one more recent event. Due to the anthropogenic nature of the trenches, it is possible that during their excavation the uppermost layers and more recent earthquake evidence has been removed. This may also be the reason why there is only evidence of event $z$ in trench 1. Furthermore, as there is a postglacial scarp at the trench site, younger events may have been recorded on the bedrock fault plane and not the trenches. However, this does not negate the possibility that we are overdue for an earthquake on the Sfaka fault, or that the fault has undergone a period of quiescence since $6102 \pm 113$ BP.

Further trenching investigations are required to better constrain the earthquake dates. Charcoal is the best material to date, but with the dating of fissure fills the charcoal may have been produced significantly earlier than the surface rupture and subsequent fissure filling. There are also few suitable areas along the Sfaka fault where sedimentation is occurring; significant colluvial deposits are only present for around 700 m of the fault (Fig. 4.3). As the Sfaka fault has an impressive fault scarp of many metres along the majority of its length, exposure dating through cosmogenic nuclides is likely the best approach to constrain earthquake dates.

Retrodeformation analyses of trenches 1 and 2 at the Sfaka fault indicate vertical displacements of 0.2, 0.1, 0.25 and 0.4 m for events $w$, $x$, $y$ and $z$ respectively. As with all retrodeformation analyses there are some uncertainties and assumptions which should be considered when interpreting the results. The vertically aligned clasts observed at the westernmost extent of the fault gouge implies that this is the location where slip occurred and this is what was modelled. However, it is possible that some slip was accommodated within the gouge, and/or some slip occurred on the bedrock fault plane itself. If this is the case the vertical displacements calculated through retrodeformation analyses are too low. Therefore, the vertical displacements should be considered minima and conservative estimates. One of the main assumptions with subaerial colluvium is that it was deposited at the angle of repose for the material (McCalpin, 2009). In trenches 1 and 2 at the Sfaka fault, the colluvium is not only scarp derived but predominantly from the hanging-wall to the south at elevated positions. Therefore, the angle of repose would not form a non-horizontal layer perpendicular to fault strike into the hanging-wall. It is more likely that colluvium deposition formed close to horizontal layers and this was modelled in the retrodeformation analyses. However, this is an assumption and non-horizontal deposition, either dipping towards or away from the fault,
cannot be ruled out. Therefore, interpretations using non-horizontal colluvial deposition will result in different vertical offsets.

### 7.2.3 Cemented colluvium

During initial fieldwork in 2013, cemented colluvium was observed at many faults throughout Crete. The Zou fault in the Sitea fault zone (Fig. 5.1b) has two prominent cemented lobes and sheets of varying thickness attached to the bedrock fault plane. Cemented colluvial sheets were also observed on the Kera, Kastelli, Asomatos and Spili faults. From our pioneering work on the cemented colluvium on the Lastros fault some interesting observations and conclusions can be made regarding the cemented colluvium’s formation process and the palaeoenvironmental conditions during cement precipitation.

The footwall block of the Lastros fault is brecciated forming two types of cataclasite (Stewart and Hancock, 1988; 1990): compact breccia sheets and incohesive breccia belts. The breccia cataclasite and cemented colluvial sheets can often look quite similar; therefore to distinguish between cataclasite breccia and cemented colluvium several criteria should be used: the internal structure, matrix composition, clast sizes and clast angularity. The presence of foliation is indicative of compact breccia sheets as foliation is not present in cemented colluvium. The matrix composition of cemented colluvium generally has a higher clay/mud content than the matrix of compact breccia sheets, although an up to 2 mm thick clay-rich gouge layer can sometimes coat their primary slip surface. Compact breccia sheets have maximum clast sizes of 2.5 cm which are moderate to subrounded, whereas cemented colluvium can have much larger clast sizes which are more angular, indicating the short transport distances and physical weathering of the footwall limestone.

Cemented colluvial sheets do not dramatically affect scarp heights, but lobes protect the fault scarp from exhumation severely limiting scarp growth. As shown in Figs. 4.4c and 4.5-P7, the scarp behind the lobe at the Lastros fault is only ~1 m in height. Either side of the lobe the scarp is over 15 m in height. Therefore, through the cementation process the hanging-wall colluvium has become fused to the footwall bedrock; when a scarp building earthquake occurs the lobe is uplifted along with the footwall, whilst the surrounding uncremented colluvium within the hanging-wall is downthrown.

Macro and microscopic descriptions of the lobe cemented colluvium show that it was originally deposited through rockfalls and debris flows, and cementation then occurred in vadose zone conditions; the source water was from a spring at the fault plane forming locally saturated zones in the vadose environment. The stable isotope analysis on the cement show
that it precipitated from meteoric water, and palaeotemperature calculations corrected for altitude show that the climatic conditions at the time of calcite precipitation were around 7 to 10°C cooler than present day. The mud/clay content within the cemented colluvium, although low, was high enough not to allow an accurate precipitation date to be determined through U-Th. Therefore, a Pleistocene marine isotope stage cannot be ascribed to the cement. However, the palaeotemperature calculations allow us to rule out a Holocene age; they strongly suggest conditions soon after a Pleistocene glacial maximum, most likely after a Heinrich event (Dansgaard et al., 1993). Deep sea coring in the North Atlantic (Bond and Lotti, 1995) indicate six Heinrich events have occurred since ~60 ka, with around ~10,000 years elapsing between events. As the palaeoenvironmental conditions were likely quite similar during these Heinrich events, hanging-wall cementation was most likely occurring soon after all of them due to raised groundwater levels (high precipitation and seasonal snowmelt). Through further studies on the cemented colluvium at faults including absolute dating, a detailed palaeoenvironmental history of study areas can be determined. At present there is a lack palaeoenvironmental studies on Crete and stable isotope studies are an effective way of reconstructing past palaeoenvironments. However, a major factor that must be assumed to calculate parent water precipitation temperatures is the oxygen isotopic composition of the parent water. This can obviously never be known for certain. Best estimates are from nearby locations with similar geological conditions. This assumption of the parent water oxygen isotopic composition does allow for misinterpretations to be made, and absolute dating techniques such as U-Th would allow comparisons with the palaeoenvironmental record to be conclusively undertaken. Samples for U-Th dating should contain pure calcium carbonate so that accurate dates can be determined.

The extent of the cemented colluvium is more than is observed at the surface. GPR profiles (Fig. 5.8a,b) undertaken on the hanging-wall show that cemented colluvium is present below uncemented colluvium. This was also verified through trial pitting. Therefore, throughout the Pleistocene the surface springs at the fault plane were not restricted to the areas where cemented colluvial deposits are currently exposed at the surface. Groundwater was seasonally high enough to form large cemented areas along strike, most likely through multiple springs. After cessation of some springs, colluvial deposition continued burying the cemented deposits.

### 7.2.4 Historical documentation

During the literature review into historical earthquakes in eastern Crete a misleading account was found in Ambrasseys (2009) regarding an earthquake that occurred in November 1815. Ambrasseys (2009) states that, according to Turner (1820), the November 1815 event
caused bad damage in two towns, Ierapetra in Crete and Bodrum in Turkey. These two towns are over 250 km apart and a large subduction event would be required to cause the damage. However, Turner (1820) actually states that he visited some ruins in Bodrum in 1815 which were most likely destroyed by earthquakes but there is no reference to when the destruction happened; definitely not during or shortly before his visit. On Turner's visit to the ruins, his Greek guide said that an earthquake was felt 20 days before, which is most likely the same event that badly damaged Crete. Therefore, the November 1815 event could be due to shallow normal faulting in Crete as it caused damage in Ierapetra and was only felt in Bodrum. This again highlights the need to always check original sources of documentation as misinterpretations can happen, especially when the source material is not in the interpreter's native language.

### 7.2.5 Seismotectonic overview

The Lastros fault is the controlling fault in the graben with the Sfaka fault having an antithetic relationship, and the two faults meet around 2.4 km depth. The dated events on the Sfaka fault therefore most likely represent coseismic activity on the Lastros fault: on some occasions when the Lastros fault ruptures the Sfaka fault slips at the same time. There is the possibility that the Sfaka fault fissure fills actually represent only accommodation events: the Sfaka fault may move sympathetically depending on the need for volume accommodation in the hanging-wall block of the Lastros fault. If this is the case, using the average recurrence intervals for the Sfaka and Lastros faults, 3318 years (Fig. 4.10) and 437 years (section 4.4.1) respectively, an accommodation event may occur after the Lastros fault has ruptured 7-8 times which produces around 1 m of extension.

The Lastros fault has accumulated around 300 m of total throw (Fig. 4.11). Using a constant slip rate of $0.69 \pm 0.15$ mm/a, which equates to a throw rate of $0.62 \pm 0.14$ mm/a, the Lastros fault only began to develop ca. 435 ± 120 ka, making it particularly young. For the same time period the Ierapetra fault displays relatively low slip rates based on displaced marine terraces (Gaki-Papanastassiou et al., 2009). Furthermore, the Ierapetra fault does not have a clear exposed scarp along its length; only near Ha Gorge is there a scarp which has been largely exhumed by the gorge itself. Therefore, it is likely that the Ierapetra fault has a current throw rate less than 0.3 mm/a, as faults with higher throw rates have preserved scarps.

The relatively young age of the Lastros and Sfaka faults (ca. 435 ± 120 ka) may be the results of a changing stress regime. The oblique slip Ierapetra fault has been the controlling fault within the IFZ since its formation in the Late Miocene, but in the Mid-Late Pleistocene the stress regime then changed to more pure extension (ENE-WSW) which caused the
development of the Lastros and Sfaka faults. If this is the case the Lastros and Sfaka faults currently pose the highest seismic threat in the IFZ.

Figure 7.4. Mapview of Crete and the Hellenic Trench showing microseismicity between July 2003 to June 2004. Event location magnitudes are between 1 to 5 (modified after Becker et al. 2009).

Figure 7.4 shows microseismicity (magnitudes 1 - 5) for the whole Crete during a 12 month survey in 2004/2005 (see Becker et al. 2009). Approximately half of the earthquakes recorded are deep (>20 km) and occurred to the south of Crete. These are therefore associated with the shallow dipping subduction zone. The other half of the recorded earthquakes are shallow (<20 km) with a significant proportion being very shallow (<10 km). Unfortunately, no structural data is presented so no focal mechanisms can be calculated; however, these shallow earthquakes are most likely associated with shallow crustal extension. It can be seen in figure 7.4 that many shallow (yellow) and very shallow (orange) earthquakes occurred in and around the IFZ during the 12 month monitoring period. Some of these occurred offshore to the north where historical earthquakes have been reported (see Chapter 4). This gives evidence to a possible offshore continuation for the faults of the IFZ. To help determine this, an offshore bathymetry survey would help to map any sub-sea
features indicative of offshore faulting. Furthermore, E-W offshore seismic profiles would help map fault locations. If this is undertaken it should not just be confined to the IFZ but other fault zones on Crete where offshore continuation is possible.

As there are two general orientations of normal faults on Crete, ~E-W and ~N-S (Fig. 4.1), there has been much debate as to which of these orientations represent the current stress regime in the shallow crust; no surface rupturing earthquakes have occurred on Crete for at least 200 years (see section 4.2.2). Our research in eastern Crete shows that the faults in the IFZ are active with ENE-WSW extension being dominant since the Mid-Late Pleistocene. Further evidence for ENE-WSW extension being the dominant regional stress regime comes from instrumental seismicity data. Kiratzi (2013) presents selected focal mechanism data for shallow (<25 km) earthquakes over Mw 4.0 (Fig. 7.5) that have occurred along the Hellenic arc. Figure 7.5 shows a series of pure normal faulting earthquakes and normal faulting earthquakes with considerable strike slip components.

Whatever the faulting style, the T-axis trend is ~ESE-WNW indicating that the normal faults with orientations close to N-S are the active ones, with the E-W trending normal faults being considerably older (Early Pleistocene or Pliocene) (Kiratzi, 2013). This is also confirmed by
Mountrakis et al. (2012) who state that for northwestern Crete the ESE-WNW trending faults are older and now inactive as they do not affect Quaternary deposits and are overprinted by N-S trending faults. Therefore, in terms of fault studies for seismic hazard analysis, future work in the region should concentrate on the N-S oriented faults (Fig. 4.1) as they likely have a higher potential for generating earthquakes. However, the E-W trending faults such as the Spili and Kastelli faults (Fig. 4.1) have postglacial scarps and are active; further research determining slip-rates and palaeoearthquake dates are also needed for these faults. The proximity of the island to the Hellenic trench renders the whole island with a high seismic hazard (Fig. 1.7) and regional uplift is the cause of the normal faulting throughout the island. However, the way normal faults have interacted with the trench system and other normal faults throughout the Quaternary is largely unknown. It is likely that some of the fault zones are less active than others, as we have suggested for the IFZ, and stress transfer may leave some fault zones inactive for long periods.

7.2.6 Future research

Future research possibilities on Crete are summarised as follows:

The southern segment of the Lastros fault should be analysed and scarp heights compared to the northern segment. A UAV (unmanned aerial vehicle) survey and photogrammetry using structure from motion software is required for this. The hanging-wall of this southern segment has little evidence of erosion from anthropogenic activity, and therefore only natural erosional effects need to be considered when choosing representative areas for slip rate calculations.

Cosmogenic nuclide dating on all faults with exposed fault scarps, including both the Lastros and Sfaka faults. This would help to determine exactly when postglacial scarp exhumation began. The most recent earthquakes can be dated and slip rates can be calculated. High resolution DEMs, e.g. from t-LiDAR, can be used to help define sampling areas along strike not influenced by sedimentation and erosion. The cosmogenic nuclide investigations should be complemented by trenching studies in areas where there are high sedimentation rates.

Offshore bathymetry studies are needed to determine accurate fault lengths. Many of the likely more active ~N-S oriented faults are located close to shorelines and possibly extend offshore. Larger fault lengths increase the maximum potential earthquake size, and this knowledge would be highly beneficial for the accuracy of seismic hazard assessments. Offshore studies should include seismic profiles perpendicular to fault zone strike.

Coastline notches should be accurately mapped using t-LiDAR and/or photogrammetry techniques. Changes in notch height over a small spatial extent may represent recent normal
faulting which affected the coastline, and therefore future research can concentrate on these more recently active onshore faults.

Cemented colluvium should be studied on all faults where it is present. This should include both stable isotope analyses and absolute dating techniques, which would allow the Quaternary palaeoenvironment to be accurately constructed throughout the whole island. The data for the ~N-S and ~E-W trending faults should be compared. Significant differences in similar geological settings would show that fault activity is influencing the groundwater, either by causing conduits or barriers for groundwater flow.


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Chapter 9: Acknowledgements

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I have spent over 3 months in Greece doing fieldwork for this project and excursions for BSc and MSc students. There are many great stories to tell and the wonderful hospitality of the Greek people will always be remembered. I have lost count of how many times we have been offered fruit and raki by land owners and farmers at the end of long day in the field. There are too many people to thank individually, but for assistance in fieldwork I would like to thank Yasar Manβ, Alex Woywode, Lauretta Kearger and Tobias Baumeister who spent a long time in the field with me. I would also like to thank my co-authors who have helped me a lot throughout this research, both in the field and the office; in particular the GIS elements which I have had a lot of assistance with. They are Sascha Schneiderwind, Thomas Wiatr, Aggelos Pallikarakis and Silke Mechernich.

I also want to thank my parents, friends, lecturers and professors who have helped me get to this position in my career. The biggest thanks go to my wife Teresa who has given me constant encouragement and support throughout.
Appendix
Appendix A: Structural measurements of the Lastros & Sfaka faults

Table A1: Compass measurements at the Lastros - Sfaka Graben

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</tbody>
</table>
### Appendix B: Trench soil descriptions

#### Table B1: Sfaka Trench 1 (South) soil descriptions

<table>
<thead>
<tr>
<th>Unit</th>
<th>Material</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>CU1</td>
<td><strong>Cemented Colluvium</strong></td>
<td>Sub-angular to sub-rounded limestone gravel clasts in a relatively hard, reddish carbonate matrix containing some pedogenic iron.</td>
</tr>
<tr>
<td>CU2</td>
<td><strong>Colluvium</strong></td>
<td>10YR 7/4 Red – light yellowish brown very clayey gravel. Gravels comprise angular limestone and range from 0.5 – 20cm in diameter. Occasional limestone cobbles.</td>
</tr>
<tr>
<td>CU3</td>
<td><strong>Colluvium</strong></td>
<td>5YR 5/6 Light brown very clayey silty gravel. Gravels comprise of angular limestone and range from 1 – 10cm in diameter.</td>
</tr>
<tr>
<td>CU4</td>
<td><strong>Colluvium</strong></td>
<td>10YR 7/4 Light red – yellowish brown gravely clay with occasional cobbles. Gravels comprise of angular limestone and range from 3 – 10cm in diameter</td>
</tr>
<tr>
<td>CU5</td>
<td><strong>Colluvium</strong></td>
<td>10YR 7/4 Light yellowish brown very silty clayey gravel. Gravels comprise of angular limestone and range from 3 – 20cm in diameter.</td>
</tr>
<tr>
<td>CFU1</td>
<td><strong>Crack Fill</strong></td>
<td>5YR 4/4 Light brown to reddish brown very gravelly silty clay with occasional cobbles (up to 15 cm in diameter) and containing roots and rootlets. Many small clasts vertically aligned.</td>
</tr>
<tr>
<td>CFU2</td>
<td><strong>Crack Fill</strong></td>
<td>10YR 6/6 Light brown to brown gravelly clay containing rare cobbles (up to 12 cm in diameter).</td>
</tr>
<tr>
<td>FG</td>
<td><strong>Fault gouge</strong></td>
<td>60% fine-grained cohesive matrix and around 40% clasts. The matrix comprises light yellow clayey silt (7.5 YR 7/6) and the clasts are comprised of gravels and cobbles of footwall limestone. Voids exposed when cleaning.</td>
</tr>
<tr>
<td>RT</td>
<td><strong>Recent topsoil</strong></td>
<td>5 5 YR 5/2 Dark brown gravelly silty clay with abundant roots and rootlets.</td>
</tr>
</tbody>
</table>

#### Table B2: Sfaka Trench 2 (North) soil descriptions

<table>
<thead>
<tr>
<th>Unit</th>
<th>Material</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>CL1</td>
<td><strong>Cemented Colluvium</strong></td>
<td>Sub-angular to sub-rounded limestone gravel clasts in a relatively hard, reddish carbonate matrix containing some pedogenic iron.</td>
</tr>
<tr>
<td>CL2</td>
<td><strong>Colluvium</strong></td>
<td>10YR 6/6 Light brown to brown clayey silty gravel with occasional cobbles. Gravels and cobbles are angular limestone and range in size from 2 cm to 30 cm in diameter. Gravels have no clear observable orientation in the west but are dipping towards bedrock fault plane in the east.</td>
</tr>
<tr>
<td>CL3</td>
<td><strong>Colluvium</strong></td>
<td>10YR 7/4 Light brown very silty clayey gravel with occasional cobbles Gravels and cobbles comprise of angular limestone and range from 3 – 15cm in diameter.</td>
</tr>
<tr>
<td>CL4</td>
<td><strong>Colluvium</strong></td>
<td>5YR 5/6 Light brown very clayey silty gravel. Gravels comprise of angular limestone and range from 1 – 15cm in diameter.</td>
</tr>
</tbody>
</table>
### Unit Material Description

<table>
<thead>
<tr>
<th>Unit</th>
<th>Material</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>CL5</td>
<td>Colluvium</td>
<td>10YR 7/4 Light brown silty clayey gravel with occasional cobbles. Gravels and cobbles comprise of angular limestone and range from 2 – 12 cm in diameter.</td>
</tr>
<tr>
<td>CL6</td>
<td>Colluvium</td>
<td>10YR 6/6 Light brown to brown clayey silty gravel with occasional cobbles. Gravels and cobbles are angular limestone and range in size from 2 cm to 15 cm in diameter.</td>
</tr>
<tr>
<td>CFL1</td>
<td>Crack Fill</td>
<td>5YR 5/6 Dark reddish brown gravelly clay with numerous rootlets. Gravels are angular limestone and range from 1 – 8 cm in diameter.</td>
</tr>
<tr>
<td>CFL2</td>
<td>Crack Fill</td>
<td>5YR 4/4 Light brownish red gravelly sandy clay. Gravels are angular limestone and range from 1 – 5 cm in diameter.</td>
</tr>
<tr>
<td>CFL3</td>
<td>Crack Fill</td>
<td>5YR 5/6 Light brownish red gravelly sandy silty clay. Gravels are angular limestone and range from 1 – 10 cm in diameter.</td>
</tr>
<tr>
<td>FG</td>
<td>Fault gouge</td>
<td>60% fine-grained cohesive matrix and around 40% clasts. The matrix comprises light yellow (7.5 YR 7/6) clayey silt and the clasts are comprised of gravels and cobbles of footwall limestone</td>
</tr>
<tr>
<td>RT</td>
<td>Recent topsoil</td>
<td>Dark brown gravelly silty clay with abundant roots and rootlets.</td>
</tr>
</tbody>
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### Table B3: Kaparelli Trench soil descriptions

<table>
<thead>
<tr>
<th>Unit</th>
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</thead>
<tbody>
<tr>
<td>CW</td>
<td>Topsoil, colluvial wedge of 1981 earthquake</td>
<td>10 YR 6/4 light yellowish brown topsoil with 30 % clasts, well rounded, range from 5 mm - 15 cm, colluvial wedge deposit from most recent 1981 earthquake, sharp boundaries</td>
</tr>
<tr>
<td>RT1</td>
<td>Silt, slightly clayish with gravels</td>
<td>10 YR 6/4 light yellowish brown, slightly clayish silt, contains 5-10% small gravels, clasts are slightly angular to rounded, many roots</td>
</tr>
<tr>
<td>TS2</td>
<td>Topsoil prior to 1981 earthquake</td>
<td>7.5 YR 5/4, brown clayey silt topsoil containing 15% gravelly clasts, many rootlets with layer, gravels range from 5 mm - 6 cm, distinct boundary to surrounding units - Topsoil prior to 1981 earthquake</td>
</tr>
<tr>
<td>TS1</td>
<td>Recent topsoil</td>
<td>7.5 YR 4/6, strong brown clayey silty, gravelly recent topsoil, clasts range from gravels to cobbles (max. 15 cm) and abundant roots</td>
</tr>
<tr>
<td>S9</td>
<td>Sand - gravelly</td>
<td>Small multi-coloured sand to gravel deposit, similar to unit G1, deposit is cut by antithetic fault, it is likely associated with unit G1 below as they are connected by the antithetic fault, with depth in trench wall the connection should be seen</td>
</tr>
<tr>
<td>F1</td>
<td>Clayey silt</td>
<td>10 YR 6/6 brownish yellow, clayey silt with 20% clasts, clasts range from 4mm -14 cm, well rounded, well sorted, soft matrix, bed is 10- 43 cm thick, boundary between lower fluvial gravels, boundary is gradual</td>
</tr>
<tr>
<td>Unit</td>
<td>Material</td>
<td>Description</td>
</tr>
<tr>
<td>-------</td>
<td>-------------------</td>
<td>-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------</td>
</tr>
<tr>
<td>F2</td>
<td>Silt</td>
<td>7.5 YR 8/3 pink silt with rare siltstone clasts, clasts up to 7 cm, bed is 10-35 cm thick, sharp boundary between layer and upper layers, fault cuts the upper and lower boundary producing an offset of 57 cm between the upper layer boundaries and around 37 cm between the lower boundaries</td>
</tr>
<tr>
<td>F3</td>
<td>Slightly clayey silt</td>
<td>10 YR 6/6 brownish yellow, slightly clayey silt with occasional white siltstone clasts up to 5 %, thickness around 16-52 cm, sharp boundary between lower pink silt, upper boundary fault, part of the material forms a wedge shaped fissure/crack fill of 108 cm length and a maximum width of 50 cm</td>
</tr>
<tr>
<td>F3a</td>
<td>Slightly clayey silt</td>
<td>As F3, except fewer white siltstone nodules and drier to the touch</td>
</tr>
<tr>
<td>F4</td>
<td>Silt</td>
<td>10 YR 6/3, pale brown silt containing occasional white/pink siltstone nodules up to 1.5 cm in length, rare small gravel, size clasts of mudstone, sandstone also present, few roots</td>
</tr>
<tr>
<td>F5</td>
<td>Clayey Silt</td>
<td>10 YR 6/3, pale brown clayey silt containing 20% sub-angular to well-rounded clasts up to 5 cm, angularness is due to the clasts being broken, sub-layers of pure clayey silt 5-6 cm long within the unit, few rootlets</td>
</tr>
<tr>
<td>F6</td>
<td>Silt</td>
<td>7.5 YR 7/6, reddish yellow silt containing occasional white siltstone nodules, same silt deposit as unit above (F6a) except as it is drier it has a lighter appearance, white siltstone nodules range from 3 mm - 4 cm</td>
</tr>
<tr>
<td>F6a</td>
<td>Silt</td>
<td>10 YR 7/4, very pale brown silt, occasional white siltstone nodules ranging from 3 mm - 5 cm, no clear pattern to nodule arrangement, unit is fractured upslope and sharp contact downslope to another gravel dyke structure and an antithetic (small) fault.</td>
</tr>
<tr>
<td>F6b</td>
<td>Silt</td>
<td>Same as F6 but less white silt nodules</td>
</tr>
<tr>
<td>F6c</td>
<td>Silt</td>
<td>Same as F6 but white silt nodule amount is slightly increased, also rare gravels are present (1-2 cm) particularly in the lower 2 m</td>
</tr>
<tr>
<td>F6d</td>
<td>Silt</td>
<td>Silt as F6c with 3-5 cm siltstone layer close to contact with unit G6, to right of siltstone layer is an area of most sandy gravelly clay/silt which has the appearance of a topsoil, but this could be due to the moisture</td>
</tr>
<tr>
<td>F7</td>
<td>Gravelly Silt</td>
<td>10 YR 6/3, pale brown gravelly silt, transition is gradual from the pure sand to gravel below and contact is less sharp to the surrounding F6, towards the top of F7 the gravel component becomes less and less - possible upper extent of gravel dyke structure</td>
</tr>
<tr>
<td>F7a</td>
<td>Gravelly Silt</td>
<td>Same as F7 but with larger siltstone nodules, largest is 8 cm but average is approximately 1-2 cm</td>
</tr>
<tr>
<td>F7b</td>
<td>Gravelly Silt</td>
<td>Silt matrix same as F7 and F7a but approximately 10% gravel clast are present (from dyke like structure), percentage increase with proximity to dyke (upslope) and decreases downslope</td>
</tr>
<tr>
<td>F8</td>
<td>Silt, partly gravel</td>
<td>Area of silt similar to F15 with abundant gravels, possibly connected to G7 within wall, gravels (20%) are up to 2.5 cm but mainly 0.5 cm and angular to rounded, sharp edge to left (north)</td>
</tr>
</tbody>
</table>

**Appendix B**
<table>
<thead>
<tr>
<th>Unit</th>
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<th>Description</th>
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<tbody>
<tr>
<td>F10</td>
<td>Silt</td>
<td>7.5 YR 6/6 brownish yellow silt containing rare white siltstone nodules, silt has aligned forming 3-4 layers (0.5-1 cm thick) of desiccated very weak siltstone which aligns perfectly with the contact to G2, perhaps the individual &quot;layers&quot; are shear planes caused by injection</td>
</tr>
<tr>
<td>F11</td>
<td>Gravelly Silt</td>
<td>Possible edge or splay of gravel dyke, at its highest it is 50% gravel and 50% silt matrix as F7, gravel size is 2 mm to 4 cm, contact is gradual on both sides with gravels being located up to 15 cm from main deposit,</td>
</tr>
<tr>
<td>F11a</td>
<td>Gravelly Silt</td>
<td>Same as F11 but larger maximum gravel size</td>
</tr>
<tr>
<td>F12</td>
<td>Silt</td>
<td>7.5 YR 7/6 reddish yellow silt, containing 10% clasts from adjacent G4a gravel deposits, contact is sharp downslope and less so upslope, shearing of all units about 20 -35 cm</td>
</tr>
<tr>
<td>F13</td>
<td>Fracture</td>
<td>Possible fracture with 1 cm wide root growing 80 cm bgl. darker appearance than surrounding silt possibly indicating preferred pathway for water, also siltstone nodules are aligned</td>
</tr>
<tr>
<td>F14</td>
<td>Silt</td>
<td>Desiccated silt containing occasional fine gravel up to 15 cm</td>
</tr>
<tr>
<td>F15</td>
<td>Silt</td>
<td>10 YR 7/4 very pale brown silt containing many siltstone nodules, particular in the uppermost 80 cm between 16 m and 14.5 m, nodules reach max. 1 cm (visible), unit is very similar to unit F6a</td>
</tr>
<tr>
<td>F15a</td>
<td>Silt</td>
<td>Silt as F15 with abundant gravels, perhaps gravels increase in trench wall, gravels form 20% of material and are sub-angular to rounded, contacts are graded to the surrounding soil</td>
</tr>
<tr>
<td>F16</td>
<td>Silt</td>
<td>7.5 YR 6/6 reddish yellow, incorporated block of pure silt, containing only white siltstone nodules, sharp contacts</td>
</tr>
<tr>
<td>F17</td>
<td>Silt</td>
<td>7 YR 6/6 reddish yellow silt, many siltstones, damper than surrounding silt</td>
</tr>
<tr>
<td>F18</td>
<td>Sandy silt</td>
<td>10 YR 7/6 sandy silt, sand is fine, some rootlets implying preferential path for water, some small sand pockets, gradual contact to surrounding soil</td>
</tr>
<tr>
<td>F19</td>
<td>Clayey silt</td>
<td>7.5 YR 5/6 strong brown clayey silt, with abundant gravels, gravels are sub-angular to rounded but mainly sub-angular, range up to 10 cm (average 3 cm), many siltstone nodules</td>
</tr>
<tr>
<td>G1</td>
<td>Sand and gravel</td>
<td>Multi-coloured sand and gravel deposit with gradual transition into material above (F7), gravels are up to 6 cm in length and sand matrix is medium to coarse, matrix component is very small, sharp contact to the surrounding unit F6, possible extend of gravel dyke structure</td>
</tr>
<tr>
<td>Unit</td>
<td>Material</td>
<td>Description</td>
</tr>
<tr>
<td>------</td>
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</tr>
<tr>
<td>G2</td>
<td>Gravel, sandy matrix</td>
<td>Multi-coloured gravel in a very minor sandy matrix, gravels are sub-angular to rounded and range from 2 mm up to 13 cm, average is 2-3 cm, poorly sorted, well graded, alignment is similar to contact with unit F6. For composition of gravel see sample within 1.5 m of the ground a silty matrix becomes more apparent but the contact is gradual, silt appears to be from the surrounding units (F6, F7a/b), contact to unit F6 is sharp, contact to unit F7a is more gradual with more and more silty matrix until the gravel is no more, this takes place over approximately 30 cm, see F7b description, gravel is very loose and crumbles with a light touch, SKAP-4 was taken from this unit, containing 50 % fine to medium gravel, 32% sand and 18 % fines</td>
</tr>
<tr>
<td>G3</td>
<td>Gravel/sand in silty matrix</td>
<td>Multi-coloured gravel (95%) with a bit sand 5% in silty matrix of surrounding soil unit F7a, ratio is approximately 50/50, Contact is quite sharp upslope and downslope and reaches the bottom of topsoil, In the upper 1m The ratio of gravel/matrix is 35/65, at 2 m from surface the gravel is sheared, approximately 25-30 cm, almost horizontal</td>
</tr>
<tr>
<td>G4</td>
<td>Gravel, silty matrix</td>
<td>Multi-coloured gravel deposit in a silty matrix (matrix is approx. 10-15%c of deposit), gravels range from few mm to ?6 cm (average 1-2cm), uppermost 30 cm close to the surface has increasing silty matrix, ?to that at 0.5 m b??? It is 65% matrix and 35% gravel, deposit does not reach the surface, contact is sharp on both sides</td>
</tr>
<tr>
<td>G4a</td>
<td>Gravel, silty matrix</td>
<td>Multi-coloured gravel with minor matrix (up to 5 %), deposit is very loose and crumbles with a touch, gravels range from coarse sand size to 4 cm, (average 0.5 - 2 cm), same as unit G4 with less matrix with depth, matrix is increasing upwards, the two splays surround a silt unit F12, all of which unit G3, G4, G4a, F12 have been horizontally sheared, void near ground, probably animal burrow, SKAP-6 was taken from this unit, containing 56 % fine to medium gravel, 34% sand and 20% fines</td>
</tr>
<tr>
<td>G5</td>
<td>Gravel, silty matrix</td>
<td>Multi-coloured (dark) gravels 15 % in a silty matrix comprising the surrounding unit F6c, some gravels are very weak and can be crumbled in fingers, others comprise solid pebbles of rock, desiccation of the silt gives the appearance of layering but is not, gravels are sub-angular-rounded</td>
</tr>
<tr>
<td>G6</td>
<td>Gravel, sandy matrix</td>
<td>Multi-coloured sand and gravel venting structure, sand 10%, gravel 90%, towards the sides the sand matrix becomes increasingly silty reaching 50/50 next to the sharp edge of the adjacent silt, gravels are sub-angular to well rounded and range from a few mm up to 6 cm, the largest clasts are in the centre of the structure and approx. 2 m below ground level, the deposit is extremely loose and collapses when vertical without touching it, at base of exposure a 40 cm high and 10-15 cm wide cemented “neck” is present, the approx. 20 % matrix of this “neck” comprises sand and silt, SKAP-10 was taken from this unit comprising 57% fine to medium gravel, 29 % sand and 14 % fines</td>
</tr>
<tr>
<td>G7</td>
<td>Gravel, sandy-silty matrix</td>
<td>Multi-coloured gravel with sandy and silty matrix (20%), gravels range from 3 mm to 6 cm (average 1-2cm) and are angular to well rounded, in parts it is partly cemented, sharp contact to surrounding material and sheared along a 1.5 m long horizontal shear plane, SKAP-11 was taken from this unit containing 52% fine to medium gravel, 28 % sand and 20 % fines</td>
</tr>
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<td>Description</td>
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<tr>
<td>G8</td>
<td>Gravel, silty matrix</td>
<td>Desiccated gravel and dark silt, forming another venting like structure, grave contain Is less than adjacent venting like structures, silt is 10 YR 4/6 dark yellowish brown and makes up 40 % - matrix supported, gravels are angular to sub-angular, vent definitely reached the surface as the appears to have been some mixing of the topsoil and vent material</td>
</tr>
<tr>
<td>G8a</td>
<td>Gravel, silty matrix</td>
<td>Exactly the same as G8, sharp contact at base, appears to breach the surface</td>
</tr>
<tr>
<td>G9</td>
<td>Gravel, sandy silt matrix supported</td>
<td>Matrix supported very loose gravel, gravels are sub-rounded to rounded, range from 2 mm to 3 cm (average 0.5-1 cm), matrix is sandy silt, silt is from adjacent material, gravel grain size diminishes at top, contact on both sides is not sharp</td>
</tr>
<tr>
<td>G9a</td>
<td>Gravel, sandy silt matrix</td>
<td>As G9 but smaller gravels, max. 2-3 cm (average 0.5 cm) sily sand matrix approx. 30%, sharp edges</td>
</tr>
<tr>
<td>G10</td>
<td>Gravel, sandy matrix</td>
<td>Very loose gravel deposit, matrix (40%) is mainly sand and some silt, still matrix supported, gravels are sub-angular to well rounded, size ranges from few mm - 3 cm and grain size reduces with height, erosion of this material has left an undulating trench wall, first sized blocks appear partly cemented, but can be crushed by hand, gravels are multi-coloured by mainly dark colours, both strands reach the surface, left side contain less silt at base of topsoil. sharp contacts on all sides</td>
</tr>
<tr>
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<td>Gravel, sandy matrix</td>
<td>As G10 but larger gravel sizes (up to 6 cm), matrix increases at the top, it appears to breach the surface</td>
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<td>G11</td>
<td>Gravel, silty matrix</td>
<td>Multi-coloured gravel (darkish with 60 % silt matrix, silt is moist near the surface, silt is 7.5 YR 4/6 yellowish red, gravel are max. 3 cm sub angular-rounded, transition to topsoil is gradual indicating that this vent most likely reached the surface</td>
</tr>
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<td>G12</td>
<td>Gravel, sandy matrix</td>
<td>Multi-coloured cemented gravel, cementing is not strong (can be easily broken with hammer), in general gravels fine-up and range from few mm to 9 cm, clast support with sandy matrix approx. 10%, sharp contact to north, uppermost 50 cm has 40 % sily matrix</td>
</tr>
<tr>
<td>G12a</td>
<td>Gravel, sandy matrix</td>
<td>As unit G12 but matrix is more reddish and gravels are under 3 cm, contact to south is less sharp</td>
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<td>G12b</td>
<td>Gravel, sandy matrix</td>
<td>Cemented gravel as G12, very sharp contact north and south, gravels are not present at the base, gravels comprise cherts, limestone and siltstone, cementation is slightly more than in G12</td>
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<td>G13</td>
<td>Gravel, sandy matrix</td>
<td>Multi-coloured fine gravel, clast sizes range from few mm up to 4 cm, partly cemented and in a sandy matrix, clasts are angular to sub-rounded, structure reaches the surface</td>
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<td>G13a</td>
<td>Gravel, sandy matrix</td>
<td>Same as G13 but more cemented into fist sized pieces</td>
</tr>
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<td>G14</td>
<td>Gravel, silty matrix</td>
<td>Multi-coloured cobble to bouldery gravel, sub-angular to well rounded ranging from few mm to approx. 40 cm in a yellowish brownish silty matrix (15 %), containing two basement limestone blocks at meter 29</td>
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<td>G14a</td>
<td>Gravel, silty matrix</td>
<td>Multi-coloured gravel, rounded to well rounded ranging from few cm to approx. 15 cm in a whitish silty matrix (approx., 10 %), much smaller grain size than G14</td>
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<td>G14b</td>
<td>Gravel, silty matrix</td>
<td>Multi-coloured gravel, sub-angular to well rounded ranging from few mm to approx. 4 cm in a whitish silty matrix (approx., 10 %), much smaller grain size than G14a, small grain sizes are accumulated along shear plane (or possible fault plane?)</td>
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### Appendix C: Stable Isotope data

**Table C1:** Stable isotope data from the lobe of cemented colluvium and calcite vein at the Lastros fault

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