Distribution and Scaling of Extensional Strain in Sedimentary Rocks

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Abstract

For this study, spatial and size distributions of normal faults and associated tensile fractures were directly measured in deformed sedimentary rocks in three extensional regions: Kimmeridge Bay and Kilve in the UK and the Maltese Islands. The collected data allow i) quantification of regional extension at different scales, ii) determination of the relative importance of large and small structures, iii) quantification of the spatial heterogeneity of brittle extension, iv) comparison of scaling laws for veins and faults belonging to the same extension event, and v) analysis of the evolution of brittle damage in space and time.

Multiple scan-lines of different length and resolution were collected in each study area to record the entire extension-related deformation. In order to quantify the heterogeneity of fracture and strain distributions, a new method of spatial analysis has been developed. The method is based on a non-parametric comparison of the cumulative frequency and extension with that for a uniform distribution and provides a measure of heterogeneity based on both the position and the displacement of individual fractures sampled along a linear traverse.

Seismically observable extension is found to scale with total extension in the three study areas, obeying a power-law relationship. The proportion of the total extension that is resolved in seismic reflection data systematically increases with increasing strain. This means that seismic data significantly underestimate the total extension at low strains but record most of the total extension in higher strain regions.

Heterogeneity analysis carried out for the three study areas shows that i) heterogeneities of the distributions of fractures and strain in an area can differ significantly, ii) heterogeneities are strongly dependent on lithology and mechanical heterogeneity, and iii) heterogeneities evolve with increasing strain.

At Kimmeridge Bay, both veins and faults display power-law scaling, but do not form part of the same distribution. Veins and faults along the Kilve-Lilstock section conform to a single power-law distribution. At the Maltese Islands fault-frequencies conform to power-law scaling, but yield a higher scaling exponent in lower-strain
zones than in higher-strain (damage) zones.

The platform carbonates at the Maltese Islands take up early extension by randomly distributed small-scale faulting. The layered and mudstone-rich rocks around Kimmeridge Bay respond to low strain by distributed ("ductile") deformation in the shales and by randomly distributed or anti-clustered veining in the stiffer carbonate beds. In inter-bedded carbonates and shales along the Kilve-Lilstock section, early extension is highly localised in narrow zones of faults and associated damage, preserving large portions of virtually unfractured rock in between.

A tensor method has been developed which permits three-dimensional strain analysis from line-data. The results of this analysis show that one-dimensional estimates of extension generally are good approximations of the maximum principal strain and that deformation in most sampled sections conforms to pure-shear, plane-strain conditions.

Fold-structures associated with normal faults are explained as due to superimposed "normal drag" within the process zone and slip-related "reverse drag" within the damage zone of a (propagating) normal fault.
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To my wife,

my family, and myself.
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Chapter 1

Introduction

Extensional basins are areas of subsidence, usually controlled by normal faults, in which sediments are deposited (e.g. Gibbs, 1984). Extension of the continental upper crust occurs at different length scales and in different tectonic settings. Depending on tectonic setting and mechanical properties of the deformed rocks, the strain (and thus the extensional structures) can be uniformly distributed or highly localized. Extensional structures commonly dominate fluid flow properties of volumes of rock as they can act as seals or channels and thus increase or decrease bulk permeability of rocks. The spatial and size distributions of populations of extensional structures are also highly relevant for estimates of crustal (or lithospheric) stretching factors, seismic moment of a region and for the prediction of number and size of structures and earthquakes beyond the limits of observation.

1.1 Aims

This study aims to directly measure the spatial and size distributions of normal faults and associated tensile fractures in sedimentary rocks formed during extension, to:

- Quantify regional extension at different scales and in particular below seismic resolution limits.
- Determine the relative importance of large and small structures in accommo-
quantify the spatial heterogeneity of brittle extension at different scales and in particular relationships between large and small structures.

• Compare the scaling laws for veins and faults belonging to the same extension event.

• Understand how brittle deformation evolves in space and time during extension of a region.

1.2 Background

1.2.1 Brittle deformation

A characteristic of rocks in the upper crust is that they generally do not behave as continua when deformed. Depending mainly on rheology, confining and pore fluid pressure, strain rate and temperature, rocks deform elastically until a certain threshold value (critical stress) is reached, beyond which deformation becomes plastic and irreversible. Based on macroscopic observation, two types of permanent deformation are distinguished in rocks (Fig. 1.1): i) Brittle behaviour, which is the breaking (failure) of the rock along discrete fractures, and ii) ductile deformation which accommodates deformation in a distributed manner without loss of cohesion (Rutter, 1986).

Laboratory experiments on rock samples deformed beyond their elastic limits and under a variety of conditions have increased our understanding of brittle deformation substantially. Fig. 1.2 is a schematic representation of brittle failure styles in triaxial tests after Griggs and Handin (1960). Rocks, deformed under extension, most commonly fail at low strains by tensile fracturing orientated perpendicular to the least compressive stress (Fig. 1.2a). In compression tests under low confining pressures (Fig. 1.2b), splitting fractures form parallel to the largest compressive stress at low strains. Increase in confining pressure results in failure along shear
Figure 1.1: Diagrammatic representation of stress-strain curves and deformation/failure modes. (a) Brittle, (b) semi-brittle, (c) semi-ductile, (d) ductile, (e) brittle, (f) ductile. (a) to (d) represent compressional conditions, (e) and (f) are extensional deformation. Modified after Price and Cosgrove (1990).
fractures at about 30° to the largest compressive stress at moderate strains. With further increase in confining pressure, shearing becomes more distributed and shear zones develop at strains of 5 to 10% (Fig. 1.2d). Very high confining pressures result in flow-like, distributed, and thus, ductile deformation (Fig. 1.2e). Some lithologies - in particular soft sedimentary rocks (e.g. mudstones, evaporites) - may show ductile behaviour and thus flow-like deformation even at low confining pressures.

Laboratory tests provide useful material properties over a range of conditions but are strongly restricted in terms of sample-size and time-scale of the experiment, which limits their significance for (large-scale) tectonic deformation processes. For example, it is almost impossible to replicate extensional shear fractures (normal faults - see following section) in this type of experiments. This makes field-studies and analogue or numerical modelling of deformation structures and processes indispensable for understanding tectonic deformation.
1.2.2 Fractures accommodating brittle extension

Fractures are surfaces along which materials (e.g. rocks) have broken and therefore surfaces across which the material has lost cohesion. Fractures are distinguished by the relative motion that has occurred across the fracture surface during fracturing and accumulation of displacement. For tensile (Mode I) fractures, the relative motion is perpendicular to the fracture walls, and thus, results in an opening of the fracture (Fig. 1.2a). For shear fractures (Fig. 1.2c) the relative motion is parallel to the surface and two types of tip displacement are distinguished. Tips with sliding motion perpendicular to the edge of the fracture are called Mode II, whereas tips with sliding motion parallel to the fracture edge are called Mode III. Thus shear fractures and faults are mixed Mode II/III.

Shear fractures that accommodate extension, either along layering or horizontally are called normal faults (e.g. Peacock et al., 2000) (Fig. 1.3), those which accommodate shortening (compression) are referred to as thrust faults (e.g. Butler, 1982). Shear fractures which accommodate significant horizontal displacement are called strike-slip faults. As the topic of this thesis is extensional strain, tensile fractures and normal faults are discussed in more detail in the following section.
**Tensile fractures - veins**

Opening-mode (Mode I) fractures are composed of two opposing surfaces that originally were bounded together to form an intact rock mass (Fig. 1.2a). The aperture of an opening-mode fracture is defined as the maximum normal distance between the two surfaces of the fracture (Pollard and Segall, 1987). Veins are formed by mineralization of Mode I fractures, and vein thicknesses can preserve the former aperture of the fractures.

Opening-mode fractures are commonly grouped in two categories. Fractures, which occur independently of layer boundaries either in massive rocks or where the fracture height is much smaller than the layer thickness, are called unconfined fractures. The second category consists of fractures that are restricted by, and terminate at, layer boundaries and thus are called confined fractures (Bai et al., 2000; Brooks et al., 1995).

Elastic crack theory is the conventional method used for describing the propagation and resulting displacements of tensile fractures in brittle rock (Vermilye and Scholz, 1995). The required idealization of rock as a homogeneous, isotropic, linear elastic material has proved satisfactory for explaining many, but not all aspects of fracture behavior (e.g. Delaney and Pollard, 1981; Pollard and Segall, 1987). Displacement profiles for idealized elastic cracks are elliptical with maximum displacement gradients at the crack tips. Elastic theory predicts that opening displacement for Mode I fractures scales linearly with fracture length (Pollard and Segall, 1987). Heterogeneity in rocks, deviations from ideal elastic behaviour and interactions between different fractures may cause deviations from the predicted aspect ratios.

Field studies of fracture frequency versus length show a wide variety of possible distributions such as power law (Segall and Pollard, 1983; Odling, 1997; Odling et al., 1999; Renshaw, 1999; van Djik et al., 2000), negative exponential (e.g. Priest and Hudson, 1981) and log-normal (Gillespie et al., 2001). The causes for these different distributions are not well understood to date and are an issue that is addressed in this thesis.
Normal faults

Normal faults (Fig. 1.3) are discrete surfaces along which the hanging wall (rocks above the fault) is displaced downwards relative to the footwall (rocks below the fault). Most of the displacement is accommodated along one or several fault planes (surfaces on which slip occurs). For dip-slip movement, the separation of two originally adjacent points in the hanging wall and footwall is called fault-displacement or slip. The horizontal and vertical components of the dip-separation measured in a vertical cross-section perpendicular to the strike of a fault are called heave and throw respectively (Fig. 1.3).

Normal faults can be “penny shaped” (unrestricted), “thumb-nail shaped” (reach the surface) or vertically restricted (confined due to rheological layering). “Blind normal faults” (penny shaped faults) do not intersect the Earth’s surface and are entirely contained within the rock volume that they displace. They therefore post-date the deposition of the host-rock. In sedimentary sequences this means that the surrounding sedimentary units are older than the fault and originally had uniform sedimentary bed thickness across the fault. Thumb-nail shaped faults reach the surface and often are synchronous with sediment deposition, causing wedge-shaped thickness changes in syn-tectonic sediments across these “growth-faults”. Vertically restricted faults typically occur within brittle layers, which are underlain and overlain by more ductile layers that arrest fault-propagation.

Many blind, planar normal faults have an approximately elliptical tip line with systematically decreasing throw from a maximum near the centre to zero at the tip (e.g. Barnett et al., 1987; Walsh and Watterson, 1989; Childs et al., 2003). For single faults: $D = kL$ Where $D$ is the maximum displacement, $L$ is the maximum length and $k$ is typically around $3 \times 10^{-2}$. Faults usually occur in populations, and interact with each other and layering, which causes deviations from this simple relationship (Scholz, 2002). Mechanical restrictions to the faulted volume also strongly affect the $D/L$ ratio.

Major normal faults (e.g. graben or basin bounding faults) often display listric ge-
ometries with a dip-angle that decreases with depth, producing a shallow-dipping detachment (Shelton, 1984; Suppe, 1985). The detachment can be located in soft sedimentary rocks (e.g. mudstones or evaporates) or it is formed by a ductile shear zone in the lower crust or upper mantle (Gibbs, 1984; Shelton, 1984; Galloway, 1986).

**Damage zones**

Depending on confining pressure, displacement and lithology, faults commonly develop a zone of crushed material between sliding surfaces, called the fault core that may significantly reduce the frictional resistance to displacement (e.g. Mooney et al., 2007) (Fig. 1.4). Usually deformation is not restricted to the fault core but is associated with a zone of fractures adjacent to the fault. This deformed rock volume around a fault is called the damage zone (Chester and Logan, 1986; Koestler et al., 1994) and is caused by at least two superimposed processes: i) stress concentrations in the tip zone (process zone) of and at linkage zones between faults during fault growth and ii) kinematic damage occurring during accumulation of displacement on the fault (Kim et al., 2003, 2004).

Field observations in sandstones (Knott, 1994) indicate that the width of damage zones increases with increasing fault displacement and that this relationship is stepped with thickness increasing sharply above certain threshold values of displacement. This relationship has been interpreted as due to discontinuous growth of the fault zone. Knott et al. (1996) observed that damage zones around normal faults are widest in the extensional fields (close to upper tip in the hanging wall and close to the lower tip in the footwall) and that they are roughly twice the fault throw in the extensional and half the fault throw in the contractional field. This agrees with observations of Shipton and Cowie (2001) for normal faults from the Navajo Sandstone with damage zone width being about 2.5 times larger than the total fault throw. For argillaceous rocks (mudstones) they suggest that deformation zones around faults should be narrower than in sandstones as argillaceous rocks are
likely to cause strain softening. Micarelli et al. (2005) find a power-law scaling for the fault-displacement damage-zone-width relationship in high-porosity carbonates with a scaling exponent of about 0.4. Janssen et al. (2002) compiled data of fault-length to damage-zone-width relationships from 6 different studies and find power-law scaling, with a scaling-exponent of about 1, over more than six orders of fault-length magnitude.

1.2.3 Fracture spacing

Several different distributions of fracture spacing have been observed in field studies: i) negative exponential (Priest and Hudson, 1976; Hudson and Priest, 1979; Sen and Kazi, 1984; Rives et al., 1992; Gillespie et al., 1993) ii) log-normal (Sen and Kazi, 1984; Narr and Suppe, 1991; Rives et al., 1992; Gillespie et al., 1993, 2001) iii) normal (Rives et al., 1992) and iv) power-law (e.g. Ouillon et al., 1996).

Randomly spaced fractures can develop in rocks with few heterogeneities and will lead to a negative exponential distribution of fracture spacing. Such distributions are also widely reported from traverses through rocks with multiple sets of differ-
ently oriented fractures (e.g. Bonnet et al., 2001). Opening-mode (Mode I) fractures in layered materials - such as many sedimentary rocks - are often confined by mechanical layer boundaries and the fracture spacing is commonly proportional to the thickness of the fractured layer (Narr and Suppe, 1991; Gross, 1993; Bai et al., 2000; Soliva and Bededicto, 2005; Soliva et al., 2006). This imposes a specific length-scale, around which the spacings are generally normally or log-normally distributed. Experimental results show that fracture spacing decreases with increasing strain as additional fractures are formed in between earlier ones, a process called sequential infilling (Gross, 1993; Ackermann et al., 2001). Eventually the fracture spacing reaches saturation which means the fractures are spaced so closely that no more fractures can infill and further increase in strain is accommodated by further opening of existing fractures (Wu and Pollard, 1995). Again this will tend to produce (log-)normal spacing distributions.

**Spatial relationships between faults and fractures**

Very few studies have examined the spatial and scaling relationships between faults and veins (Gross et al., 1997). Veins and normal faults formed during the same tectonic event commonly display a (strong) spatial dependency (Putz and Sanderson, in press). Veins are often clustered within damage zones around faults accommodating, at least part of, the wallrock damage (Fig. 1.4). Such fractures can either result from tip- and linkage-damage as the fault develops or be formed in response to displacement accumulation on a fault (kinematic damage) (Peacock, 2001).

### 1.2.4 Fracture populations

Scaling attributes of fracture systems promise a means to statistically predict the occurrence of structures at a certain scale if data at other scales are available. In earthquake hazard assessment, the main issue is the validity of the Gutenberg-Richter law for predicting the probability of occurrence of large earthquakes. In the hydrocarbon industry, such scaling laws provide a key to predicting the nature
of sub-seismic fracturing that significantly influences the reservoir and cap rock quality. In groundwater applications or for waste disposal, fluid flow is particularly sensitive to the properties and scaling of fractures systems. In civil engineering and mining, many material properties are related to fracturing (e.g. strength, stiffness, permeability, blastability).

Innumerable studies dealing with scaling of fracture populations have been carried out (e.g. Evans, 1990; Yielding et al., 1992; Jackson and Sanderson, 1992; Needham et al., 1996; Johnston and McCaffrey, 1996; Bonnet et al., 2001). Most of these studies, however, are based on data-sets with limited scale ranges of rarely more than two orders of magnitude and/or are biased due to incomplete sampling within the observed scale range (e.g. Pickering et al., 1995).

The majority of studies have sought to establish a power-law relationship between the frequency \(N\) and displacements \(d\) of faults, such that:

\[
N = Cd^{-E}
\]  

(1.1)

Where \(E\) is a scaling exponent, often referred to as the 'fractal dimension', and \(C\) is a constant, defining the frequency at unit displacement.

Several authors (Scholz and Cowie, 1990; Marrett and Allmendinger, 1992; Pickering et al., 1995) have shown that the extrapolation of frequencies, and hence, the estimation of strain, requires accurate and unbiased estimation of \(E\). This is difficult, if not impossible, to achieve with samples of limited scale range.

In this study a radically different approach is taken that involves directly measuring extension on a wide range of scales based on a hierarchical sampling strategy, rather than estimating it from extrapolation of frequency distributions.

### 1.2.5 Evolution of (extensional) fracture populations

The potential change in the character of fracture populations during basin formation is an interesting problem. Brittle fault populations may develop with increasing extension through one or several of the following processes: i) nucleation of new faults
by rock fracture, ii) propagation of existing faults, iii) linkage (coalescence) of existing faults to form bigger structures, and iv) cessation of activity ("death") of faults. Some data, such as that from earthquake seismology present a snapshot in geological time, showing only the faults that are active during one or several seismic events. Palaeoseismological studies have shown that significant variations in seismicity can occur through time (e.g. Crone and Haller, 1991; DePaolo et al., 1991). Field studies (including those using seismic reflection data) provide data-sets that are integrated over time. Earthquake studies suggest that only a small fraction of the total seismic moment in an active area is contributed by small faults (Jackson and McKenzie, 1988). Field studies, on the other hand, have found that small-scale faulting may contribute significantly to the total strain (Walsh et al., 1991; Marrett and Allmendinger, 1992; Pickering et al., 1994).

Numerical and analogue models suggest that active fault populations should evolve from power-law to exponential as brittle strain within a region increases (Spyropoulos et al., 1999; Ackermann et al., 2001, Hardacre and Cowie, 2003). The models show that fracture populations evolve continuously as a function of strain and fracture density. After a critical strain is reached, cracks begin to nucleate and their numbers increase rapidly. In this regime, dominated by nucleation and the growth of individual fractures, the population generally exhibits power-law length distributions. As the fracture density (and extension) increase further, fracture (fault) interactions become increasingly important, the nucleation rate decreases as more regions become stress shadowed, and linkage (coalescence) becomes more common, resulting in a decrease in the number of fractures. In this coalescence regime the population evolves towards a power-law length distribution with lower exponent or a negative exponential length distribution.

Finally, at high strain, the population evolves towards a saturated state, characterized by system-sized cracks with spacings proportional to the layer thickness (Scholz, 2002). This progression agrees with the observations of natural systems (Gupta and Scholz, 2000; Walsh et al., 2003) where the “system-size” may be a mechanical con-
finement such as the thickness of the Upper Crust or the thickness of the sediment infill of a basin. Depending of the scale of the confinement the saturated state and thus the transition from power-law (scale invariant) to a log-normal or negative exponential (scale dependent) distribution will be reached at different strain values. Based on these modelling results, higher-strain regions should display deviations from power-law to exponential size-frequency distributions. However, as the inactive faults follow a power-law distribution and are more numerous, the fault-population as a whole (active + inactive faults) may either still display an approximate power-law scaling, or, may show a kink in a size-frequency log-log plot with a smaller slope (exponent) for smaller (earlier) faults and a higher slope for (larger) active faults. Such transitions in scaling relationships have been observed in nature at different strains and confinements (e.g Fossen and Rornes, 1996; Hunsdale and Sanderson, 1998; Wilson, 2001). In this study it is examined how different lithologies (or sedimentary sequences) respond to extension and how the deformation patterns evolve with increasing strain.

1.2.6 Brittle extension (strain)

As discussed above, rocks in the upper crust do not behave as continua when deformed. Extension is taken up on discrete structures via slip (faults), opening (tensile fractures) or precipitation (veins). Thus, the deformation is a product of the relative motion of largely undeformed blocks of rock on discrete surfaces comparable to “large-scale grain-boundary sliding” (Fig. 1.5a). This localisation of deformation along discrete structures causes many problems in describing and quantifying the amount of accommodated strain in a volume of rock. The “discrete strain” (e.g. slip along faults, aperture of veins) is what can be observed in the field and, depending on the resolution, also from geophysical data. As the term ‘strain’ should be used for continuous deformations only, the concept of “fault-strain” has been introduced (Jamison, 1989) to describe the deformation caused by a number of faults. Most of this thesis deals with the distribution of extension in the direction of max-
Figure 1.5: Fault-strain. (a) A marker-layer (yellow line) is displaced and extended by 4 normal faults. (b) Restored geometry of the marker layer (yellow line) before extension. The difference in length between \( L \) and \( L' \) is about 3%.

In other words, three-dimensional strain is examined along sections parallel to the maximum principal extension and thus treated as one-dimensional (longitudinal) strain (Fig. 1.5). This elongation \( E \) is defined as the change in length of a line from initial length \( L \) to final length \( L' \).

\[
E = \frac{L' - L}{L} \quad (1.2)
\]

Several studies have dealt with methods to estimate fault-strain from field data (Cladouhos and Allmendinger, 1993; Fossen and Tikoff, 1993; Gauthier and Angelier, 1985; Horsman and Tikoff, 2005; Jamison, 1989; Little, 1996; Molnar, 1983; Peacock and Sanderson, 1993) or earthquake data (Molnar, 1983; Marret and Allmendinger, 1990). However, the developed techniques either require information that usually is not obtainable from field-data (such as volume of the faulted region or area of the fault surfaces), or the geometric models are over-simplified and are thus not very applicable.

### 1.2.7 Extensional sedimentary basins

As mentioned earlier, the basic definition of a sedimentary basin is that it is an area of subsidence in which sediments are deposited, with subsidence commonly being
A wide variety of models have been proposed for how sedimentary basins originate and develop, however most have limited predictive power and rely on unobserved processes to operate in the crust and upper mantle. Two classic models for lithospheric extension have been widely accepted, one by McKenzie (1978) and one by Wernicke (1982). The McKenzie model (1978) is based on pure shear which results in symmetric basins (Fig. 1.6a) whereas the Wernicke model (1982) is based on simple shear which results in asymmetric basins (Fig. 1.6b).

**McKenzie Model**

In the pure shear model proposed qualitatively by Falvey (1974) and quantitatively by McKenzie (1978) there is an initial stretching of the lower crust and upper crust. The strain is distributed evenly across the continental lithosphere. Crustal thinning and tectonic subsidence is accommodated by ductile pure shear extension of the lower crust and lithospheric mantle and by brittle normal faulting in the upper
crust. For isostatic equilibrium to be maintained the thermal boundary forming the base of the lithosphere needs to rise (Fig. 1.6a). The initial “tectonic subsidence” is then followed by a much longer period of thermal subsidence as the base of the lithosphere falls in response to cooling.

**Wernicke Model**

In the simple shear model proposed by Wernicke (1982) a low-angle detachment fault penetrates through the lithosphere. The detachment allows for extension by domino-style block rotation in the upper plate. Extension is accommodated by ductile shearing in the lower crust and by brittle faulting in the upper crust. This model permits a lateral offset between basins caused by thermal subsidence and basins produced by tectonic faulting (Fig. 1.6b).

**Stretching factor**

The crustal stretching factor ($\beta$) is defined as the ratio of original (undeformed) thickness of the crust ($y_0$) to thickness of the crust after extension ($y'$).

$$\beta = \frac{y_0}{y'}$$  \hspace{1cm} (1.3)

In the original uniform stretching model proposed by McKenzie (1978) crustal and lithospheric stretching is equal and thus the stretching factors are the same.

**Implications of McKenzie’s uniform stretching model**

McKenzie’s uniform stretching model is an essentially one-dimensional consideration of lithosphere-scale pure-shear deformation. It considers a lithostatic column that is made up of an upper portion with lower (crustal) density and a lower portion with higher (sub-crustal lithosphere) density. Uniform stretching of the entire column (the entire lithosphere) results in two responses: i) the thinning of the crust, largely accommodated by brittle deformation, causes a permanent subsidence and ii) the ductile (or viscous) thinning of the mantle portion of the lithosphere which
may cause transient elevation changes (uplift or subsidence) due to (Airy) isostatic compensation to maintain isostatic equilibrium.

Assuming that the stretching occurs fast (i.e. in less than 20Ma), radioactive heat production can be ignored, isostatic compensation is maintained throughout, and the continental surface is initially at sea level, the initial (faulting related) subsidence is given by

\[ y_S = \frac{1 - 1/\beta}{\rho_m - \rho_s} \left[ \rho_m y_L - \rho_C y_C - (y_L - y_C) \rho_{SC} \right] \] (1.4)

Where \( y_S \) is the subsidence, \( y_C \) and \( y_L \) are the undeformed thicknesses of the crust and lithosphere respectively, \( \rho_m \), \( \rho_C \) and \( \rho_{SC} \) are the average densities of the mantle, crust and subcrustal lithosphere, respectively, \( \rho_s \) is the average bulk density of sediment and water filling the basin, and \( \beta \) is the stretching factor. Equation 1.4 allows predictions of initial subsidence of an extensional region based on the crust/lithosphere thickness ratio. The subsidence is positive for values of the crust/lithosphere thickness greater than 0.12. This implies that in regions with initially “normal” crustal thickness, say 20 - 40 km, initial subsidence with no uplift will occur for any stretching factor.

**Estimation of the stretching factor**

Knowledge of the crustal or lithospheric stretching factors (which are identical in the original McKenzie model) is essential for predictions of geothermal gradients, heat-flow history of sediment fill and for assessing the total seismic moment of an extensional region. Several methods have been proposed to estimate the total extension that has occurred in a basin. Stretching factors can be derived from:

1. thermal subsidence history. This method is directly based on a prediction of McKenzie’s model which is that initial uplift or subsidence of the extended region depends only on the ratio of crustal to lithospheric thickness and the stretching factor \( \beta \).
2. observed crustal thickness changes in deep seismic refraction and wide angle reflection data (e.g. Emter, 1971; Reston, 2007).

3. tectono-stratigraphic modeling of basin development through time using a rheologic model of the lithosphere (e.g. Odinsen et al., 2000).

4. inversion of strain rate history from subsidence data (e.g. White, 1993).

1.2.8 Thesis in context

Any dataset is restricted, both, in dimensions (extent) and resolution which means that estimates of the extension will always underestimate the total extension of the observed region. A recurring observation in studies of extensional basins is that the amount of extension visible on normal faults (e.g. from seismic reflection profiles) is significantly less than the amount of extension indicated by crustal thickness and thermal subsidence (Blundell, 1991; Wagreich and Decker, 2001). It has been suggested that small-scale structures may account for this discrepancy (Walsh et al., 1991; Marrett and Allmendinger, 1992). Scaling laws allow the prediction of fault frequencies beyond the observed scale range of observed structures and have been used for estimating total brittle strains (e.g. Scholz and Cowie, 1990; Marrett and Allmendinger, 1991; Walsh and Watterson, 1992).

However, the applied scaling laws are usually inferred from data of limited fault-displacement scale-range (typically less than two orders of magnitude) and it is difficult to assess their validity in particular for small faults and tensile fractures which are below the resolution of seismic data.

Spatial distributions of veins and faults have traditionally been studied separately, with the result that spatial and size relationships between normal faults and associated veins have received little attention (Putz-Perrier and Sanderson, 2008).

Heterogeneity of extension related damage and the evolution of this heterogeneity in space and time has been even more neglected. This is surprising given that many important physical properties of rocks (e.g. permeability and conductivity) are closely
related to the distribution of extension.

In this thesis extensions are directly measured on normal faults and veins over a wide range of scales (>5 orders of displacement magnitude) to examine the relative contributions of larger and smaller structures to the total extension of a region. The rationale has been to extend traditional, exposure-scale, field observations to measurement at the “map” or “seismic” scale. In addition the spatial distributions and size-distributions of extensional structures in sedimentary sequences are examined. Thus, this study fills a gap, both in terms of the size-range of observed structures and in methodology, in the measurement of heterogeneity of brittle strain.

1.3 Objectives

- Choose suitable study areas that allow normal faults and tensile fractures to be recorded over a wide range of scales.

- Integrate collected field-data and available high-resolution data of different length-scales and resolutions to compile representative samples of fracturation in the studied regions.

- Develop a method for heterogeneity analysis of one-dimensional spatial data.

- Analyse the data-sets in terms of distribution and scaling of the observed structures and their accommodated extension.

- Establish models for the temporal and spatial evolution of the examined fracture populations.

- Develop a three-dimensional tensor method for analysis of brittle strain and validate the one-dimensional approach.

- Develop models for fold structures associated with normal faulting.
1.4 Layout of thesis

Chapter 1 provides an introduction to this thesis and gives an overview of the state of research on brittle extension and the aims of the thesis within this context. In particular the characteristics of tensile fractures and normal faults are reviewed before discussing fracture populations in terms of the extension they accommodate, their spatial distribution and how these properties may evolve in space and time. Finally basic models for basin formation and lithospheric stretching are reviewed.

Chapter 2 provides an overview of the methods used and developed for collecting and analysing field-data. These are a “hierarchical sampling strategy”, statistical methods for the quantification of spatial heterogeneity of fracture spacing and strain, and methods for estimating cumulative fracture frequency and extension from multiple line samples of a fracture population.

Chapters 3 to 5 are three extensive field studies, carried out in Southern England and on the Maltese Islands, and make up the core of the thesis. These chapters examine extensional systems in sedimentary rocks in terms of i) their spatial distributions of fractures and strain and ii) the relative importance of sub-seismic scale fractures for accommodating regional extension. In Chapter 3 relationships between normal faults and veins are examined within a mudstone-dominated extensional region at the Dorset coast. Chapter 4 focuses on faulting and veining in a higher-strain section in inter-bedded carbonates and shales within the Bristol Channel Basin. Chapter 5 is a study of faulting in the Maltese Graben System, with focus on differences in fault-pattern and extension between higher and lower strain zones.

In Chapter 6, a tensor-method for the three-dimensional analysis of brittle extension based on line-samples is presented and applied to examples from the field-studies. This allows testing and validation of the essentially one-dimensional approach for estimating the maximum extension that is used in Chapters 2 to 5. Chapter 7 addresses drag-folding associated with normal faulting, based on field observations and supported by a simple numerical model.

Chapter 8 discusses and summarizes the findings of this thesis and how they fit into
the larger picture. A brief outlook on further research opportunities related to the results of these studies is given before some global conclusions on the results of the thesis are drawn.
Chapter 2

Methods

2.1 Introduction

This chapter gives an overview of the methods used for the collection and analysis of field-data presented in Chapters 3 to 5. The aim of the field-studies was to examine extensional systems in terms of the spatial distributions of fractures and strain, and in terms of the relative importance of small-scale fractures compared to large-scale faults in accommodating the regional extension.

The key to achieve these aims was to compile data-sets that are representative in terms of frequency and spatial distribution of fractures and faults over the entire observed scale-range (\(<\text{mm to }>100 \text{ m displacements}\)).

Commercial (mostly hydrocarbon related) seismic reflection profiles, the most widely used type of data for studying fault-populations, have a lower-resolution limit of 10 m to 20 m and cannot directly resolve fractures with smaller displacements, which are the main target of this study. Well-core data provide high resolution information of small fractures but the samples are usually too small and localised to be of much use in the analysis of regional deformation. Only outcrop-studies provide the resolution and spatial extent needed to examine regional fracture and strain distributions below seismic resolution. Outcrops suitable for this study need to (Fig. 2.1):

- Provide several kilometres of continuous outcrop.
Figure 2.1: Hierarchical sampling of regional deformation. (a) shows a cliff-section of several kilometres length. Two marker-beds are displaced by a set of normal faults. (b) Detailed view of a damage zone with tensile fractures (veins). (c) Detailed view of a “background strain” section with regularly spaced veins. (d) Detailed view of a folded marker-bed. The two marker-beds in (a) can be traced across the entire outcrop, and thus can be used to estimate the extension accommodated by faults. (b) to (d) provide higher resolution samples of bed-scale damage.

- Intersect the studied structures at high angles.
- Contain marker layers which can be traced over long distances and permit accurate measurements of fault-displacements.
- Have a well-known stratigraphy which allows correlation of marker beds across faults with displacements > outcrop height.
- Permit high-resolution scan-line type sampling of fractures in marker-beds.

Thus, the first step was to identify suitable study areas which fulfil the above criteria. A perfect outcrop would allow continuous sampling of extensional structures with mm to 100 m displacements over distances of 10s of kilometres. However, even the most exceptional outcrops are limited, both in extent and resolution. For this reason, a sampling-strategy was needed that allows integrating of data of different length-scales and resolution to attain a representative data-set of regional extension.
and heterogeneity of deformation at different scales.

## 2.2 Hierarchical sampling

Good study areas, like the ones chosen in Chapters 3 to 5, provide continuous cliff-sections of several kilometre lengths. Marker-beds within these sections can be traced over long distances and allow fault-displacements $\geq 0.1$ m to be determined along entire sections (Fig. 2.1a). Within these, it is possible to measure fracture apertures and displacements within single beds at $<\text{mm}$ resolution over several 10s of metres and to locally analyse micro-fracturation from thin-sections and deformed markers (Fig. 2.1b to d). Where available, these field-data can be supplemented by high-resolution seismic data or high-quality geological maps to cover even larger areas and faults.

### 2.2.1 Line samples

To sample regional deformation in a consistent manner over the entire observed scale range, and from different sources (thin sections, outcrops, geological maps, seismic data), one-dimensional samples have been used for this study. Scan-line type data (Priest, 1993) have many advantages:

- They are simple to implement in a consistent way, with clear specifications of resolution limits and known sample lengths.

- The limitations and sample biases are well understood and in many cases can be corrected for (Terzaghi, 1965; Priest, 1993; Peacock and Sanderson, 1993).

- They record both the position and size of displacements (Gillespie et al., 1993).

- Line samples can be used to measure displacement gradients (i.e. variation of strain along traverse) and fracture spacing.

- They can be used to sample at any length-scale.
• They allow an objective comparison of extension and fracture distributions in different lithologies and structural locations.

• Natural exposures are essentially 1-D (linear), i.e. the exposed trace of beds in cliffs and the narrow outcrops on wave-cut platforms.

• Line sampling along particular beds is analogous to standard methods used to sample fault populations mapped in seismic data.

• Access to 2-D samples is limited (i.e. outcrops) and/or restricted in terms of resolution (e.g. geological maps, seismic data). Where available, these data-sets can be sub-sampled by (multiple) traverse lines to derive line-data.

• Access to 3-D volumes is limited and commonly restricted in terms of extent and resolution. Where available, these can be sub-sampled by (multiple) traverse lines to generate line-data.

Thus, one-dimensional scan-line-type information can be extracted from any geological data-set and provide a means of sampling extensional structures in a consistent manner over the entire displacement-scale range observed in sedimentary basins.

An “ideal study area” of sub-seismic-scale fractures provides four types of data-sets of different extents and resolutions (Fig. 2.1):

1. Regional geological cross-sections (*map-scale fault-lines*),

2. Long continuous outcrops such as coastal cliffs (*cliff-scale fault-lines*),

3. Outcrops which are shorter but provide higher resolution (*bed-scale vein-lines*), and

4. *Thin-sections* and deformed markers (e.g. fossils) which locally sub-sample the *bed-scale vein-lines*.

*Map-scale sections* can be derived from high-quality geological maps (e.g. scale 1:10,000), or from high resolution seismic reflection data. These lines have a typical
length of $\geq 10$ km, and usually record all faults with throws $\geq 5 - 10$ m. Due to their regional extent they provide a good sample of the regional fault-strain at resolutions within and below commercial seismic resolution limits (Fig. 2.1a).

Cliff-scale sections are collected along continuous exposures of $\geq 1$ km lengths and typically resolve all faults with throws $\geq 0.1$ m. Ideally these sections should be located within the map-scale fault-lines to allow integration of the data. The main function of these lines is to sample the sub-seismic scale fault-strain, and to identify damage- and background-strain zones (Fig. 2.1a).

Bed-scale sections are high-resolution ($\geq 0.1$mm displacements) sub-samples of cliff-scale fault-lines with a typical length of 10 to 100 m. These sections are key for sampling the vein-scale component of brittle deformation and have to be well-placed, both in terms of number and position, to assure a representative sample of damage- and background-strain zones (Fig. 2.1a to d).

Thin sections are produced from samples taken at regular intervals from beds which are traced in bed-scale sections. They provide very high resolution ($\ll$mm) and are essential to assess the importance micro-fracturation. Similarly, deformed markers such as ammonites, found along bed-scale vein-lines can be used to estimate micro-strain (Fig. 2.1c,d).

2.2.2 Data collection

For each fault (fracture) encountered along the line, its distance from the origin of the line, its heave (or thickness for veins), and its dip and strike are recorded (Fig. 2.2, Table 2.1). Where bedding is not horizontal, bed-dips are measured at regular intervals along the line. In cliff-sections it is usually possible to trace marker-horizons over great distances if bedding is sub-horizontal (Fig. 2.2b). Displacement on a fault offsets the marker and can be directly measured. This allows even small displacements to be accurately determined.

Sample lines should be orientated at high angles to the mean fault trend observed in the region under investigation (Fig. 2.2a). Errors due to oblique sampling (i.e.
Table 2.1: Spreadsheet-type data-form, recording the measurements taken along the cliff-scale sample line shown in Fig.2.2. Numbers in red are measurements (and positions) recorded in the field, black numbers are derived from the measurements using simple geometric relationships.

Displacements

The largest faults in this study are observed in map-scale and cliff-scale sections. For these data-sets, throws (vertical component of dip-separation) are estimated from stratigraphic separation that is defined as the stratigraphic thickness between horizons on either side of a fault that were brought next to each other due to the displacement on the fault. Detailed stratigraphic logs allow correlation of marker-beds on either side of a fault and to determine their vertical distance from each other with high accuracies (Table 2.2). For smaller faults in cliff-scale sections, fault-heaves (horizontal component of dip-separation) are directly measured with tape-measures. Tensile fractures (veins) are commonly orientated at high-angles to bedding (sub-vertical). Their aperture (thickness) can be directly measured with a hand-lens (i.e. 10x magnifier with a scale graduated to 0.1 mm).
Figure 2.2: Example of a "cliff-scale fault-line" crossing 10 normal faults over a distance of 1500 m (dashed line). An alternative oblique sample line (dot and dash line) is shown and discussed in the text. (a) Map view of the section. Solid lines are normal faults with tick marks indicating the down-thrown (hanging wall) side. Number gives the dip-direction (azimuth) of the fault. (b) Cross-section along the sample line in (a). Faults are shown as black solid lines. Arrows indicate the slip direction; apparent dip-angles are given. The green line represents a marker-bed which is displaced by the faults. The sample-line traces this marker-bed. (c) Cumulative heave and number plot for the displaced marker-bed.
Table 2.2: Measurement-error estimates for the sample-line data over the entire observed scale range. The errors are given as absolute values (m) and as proportion (%) of the length-scale of observation. (a) Errors for position-measurements are generally low, i.e. smaller than 1% over the entire scale range. (b) Errors for displacement measurements are somewhat higher due to the smaller length-scales of fault displacements and vein thicknesses. For the largest faults considered in this study (displacements >100 m) the error can go up to 5%. However, for the majority of structures examined, the error is within 2% of the measured displacements.

The displacement measurement error for the largest observed structures (>100 m displacement) lies below 5% (Table 2.2). Most of the sampled faults show displacements <100 m and the error for these structures is estimated with <2%. For the smallest fractures, measured directly in the field, the error is somewhat higher due to their small displacements (apertures) which are close to the resolution limits. However, the measurement error for these structures still lies below 10%. Vein-apertures in thin sections can be very accurately determined and errors for these measurements are below 1% (Table 2.2).

If the mean fault-trend is perpendicular to the trend of the sample line (Fig. 2.2a) the heave measurements can be directly corrected for the deviation between fault azimuth (dip-direction) and line-trend (Fig. 2.2a).

\[ \text{heave}_{\text{CORR}} = \text{heave}_{\text{OBS}} \times \cos \alpha_{\text{DEV}} \]  

(2.1)

Where \( \text{heave}_{\text{CORR}} \) is the corrected heave parallel to the sample-line trend, \( \text{heave}_{\text{OBS}} \) is the observed heave.
is the observed (real) heave and $\alpha_{DEV}$ is the acute angle between sample-line trend and fault-azimuth (Fig. 2.2).

**Position of faults**

Positions of faults along *map-scale sections* were directly taken from the map at the intersection points between sample-line and fault (Fig. 2.2a), or derived from differential GPS positioning where high-resolution seismic data are used. For long *cliff-scale sections*, fault-positions were determined with a hand-held, EGNOS enabled GPS receiver. EGNOS stands for the *European Geostationary Navigation Overlay Service* that augments the US GPS and Russian GLONASS military positioning systems by combining them with three geostationary satellites and a network of ground stations. By comparing the signals received from the GPS and GLONASS systems with the actual positions of its ground stations, EGNOS provides a corrected signal with a theoretical accuracy of 2 metres, compared to about 20 metres for GPS and GLONASS alone. Even though the estimate of 2 metres is too optimistic, experience has shown that the accuracy of the EGNOS signal is usually at least 8 metres which makes it very suitable for map-scale fault-sampling.

For shorter (<1 km) *cliff-scale sections* and for *bed-scale sections* tape measures were used to determine fault and vein positions within the traverse. In addition EGNOS positions of start and end of each section were recorded.

The measurement error for fault positions is estimated to be <1% of the line-length over the entire observed scale range (Table 2.2).

If the sample-line is orientated oblique to the mean fault trend, it is necessary to correct the measured distances (positions of faults). This can easily be done by projecting the intersections of faults and sample-line on a line that is orientated perpendicular to the mean fault trend (Fig. 2.2a):

$$\text{dist}_{CORR} = \text{dist}_{OBS} \times \cos \beta_{DEV}$$  \hspace{1cm} (2.2)

Where $\text{dist}_{CORR}$ is the corrected distance (position of fault), $\text{dist}_{OBS}$ is the observed
(measured) distance and $\beta_{DEV}$ is the acute angle between the trend of the sample line and a line perpendicular to the fault trend.

To assure the representativeness of the collected data for the fracture and strain distributions of the study area, it is important to assess the heterogeneity of the deformation in a region and to systematically sample higher and lower strain zones at each resolution-level.

### 2.3 Fracture distribution

#### 2.3.1 Cumulative plots

The collected data are best visualised by plotting cumulative fracture-number and heave against the corrected distance (Fig. 2.2c). Cumulative data-sets and plots preserve both spatial information and magnitude of brittle deformation and thus all essential information needed for heterogeneity analysis. For this reason, arranging the collected data into cumulative format provides a good starting point for further analysis.

Throughout this thesis the term 'heave' is used to include the aperture of the veins. Cumulative numbers versus corrected-distance plots represent the fracture frequency. Low gradients on these graphs indicate low frequencies whilst steep slopes represent closely spaced fractures. Constant gradients indicate homogenously distributed fractures whilst varying gradients indicate heterogeneously spaced fractures.

As the ratio of heave/distance is the one-dimensional (longitudinal) extension, a cumulative heave diagram (Fig. 2.2c) represents the distribution of strain (extension) along the sampled section. Low gradients on the cumulative heave graphs represent lower extensions, whilst steep slopes represent higher extensions. Constant gradients of the cumulative heave graphs indicate continuously distributed (homogeneous) deformation whilst large steps and gradient changes indicate localized (heterogeneous) deformation.
2.3.2 Fracture spacing

Coefficient of Variation

The most common method for characterizing the spacing of fracture populations is the determination of the *Coefficient of Variation*, which is the ratio of the standard deviation ($s$) to the mean ($m$) of the spacing values. For regularly spaced fractures the standard deviation is small ($s \ll m$), hence $C_V \rightarrow 0$; for highly clustered fractures $s$ is large ($s > m$) and $C_V > 1$; for randomly located fractures from a uniform distribution, the spacing values have a negative exponential distribution (see for example Priest, 1993) with $s \approx m$, and hence $C_V \approx 1$. For small samples it is better to use $C_V^* = C_V \frac{n+1}{n-1}^{\frac{1}{2}}$, where $n$ is the sample size (Gillespie et al., 2001; Gillespie, 2003). This method has been applied to characterise the spatial evolution of modelled fracture and fault populations (Ackermann et al., 2001; Gillespie et al., 2001) and is used in this thesis to discriminate regular ($C_V < 1$), random ($C_V \approx 1$) and clustered ($C_V > 1$) distributions.

Kuiper’s Test

This heterogeneity-test is based on a non-parametric comparison of the cumulative number of fractures along a traverse with that for a uniform fracture-distribution (Kolmogorov-Smirnov Test). In other words, it is a measure of the deviation of the observed cumulative graph from a straight line (uniform distribution). The method also provides a statistic that may be used to test a cumulative data-set for significant departures from a uniform distribution and thus for the statistical significance of the derived heterogeneity value. Unlike the Coefficient of Variation, this method can also be used to quantify extension-heterogeneities and thus is explained in detail by means of two theoretical examples in section 2.4.1 (Strain heterogeneity).
2.3.3 Fracture frequency

Because of the different length-scales of map-scale, cliff-scale and bed-scale line-samples, care has to be taken when interpreting the data in terms of regional fracture frequencies and scaling.

Due to the regional extent of map-scale and long cliff-scale sections these data can be regarded as being representative on a regional scale. Thus, fault-frequencies per heave-interval can be directly determined by “binning” the collected fault-displacements (e.g. in bins of half-order-of-magnitude) from map-scale and cliff-scale lines.

Shorter cliff-scale and bed-scale sections provide local samples of the regional deformation and need to be treated with more care. However, provided, that a sufficient number of these lines is collected to represent this regional heterogeneity, it is possible to estimate total fracture frequencies. Usually it is possible to distinguish different ”deformation types” in a study-area. Most commonly these will be “damage-zone” with high fracture frequencies (fractures/m) compared to “background-zones” with lower frequencies (Fig. 2.3). The lines can be grouped depending on the deformation types they represent and the average frequencies within each heave-interval for each type can be determined. Based on map-scale and cliff-scale observations it is possible to estimate the regional extent of higher strain (damage) zones and lower strain (background strain) zones. Multiplying the regional extents with the derived average fracture frequencies for each heave interval and deformation type gives an estimate of the regional fracture numbers and frequencies.

The most common method of representing these interval-frequencies is by plotting them in cumulative frequency or cumulative number plots (Fig. 2.4). By using logarithmic bins and plotting the cumulative values in log-log space, it is possible to test the sample for power-law size-distribution (straight line on a cumulative log-log plot) and for deviations from self-similar distributions. Plotting cumulative binned fracture-frequencies for damage and background-strain zones separately can be useful to examine systematic differences in fracture size-distributions between different
deformation types.

2.4 Strain distribution

2.4.1 Strain heterogeneity

Heterogeneity of fracturing depends on two components: The spatial distribution of the extensional structures (joints, veins and faults) and the amount of displacement (aperture or heave) on each of these structures. Although the Coefficient of Variation is useful for analysing fracture spacing, it cannot easily be adapted for the investigation of strain distributions as it does not consider the size of the displacement on the fracture. For this reason a method is needed that allows the analysis of the spatial distribution of displacements. The following section introduces a workflow based on Kuiper’s Method (Kuiper, 1960) that allows quantification of spatial heterogeneity of any cumulative distribution (such as fracture frequency or strain).

Examples of different strain distributions

To discuss the problem of quantifying spatial heterogeneity of extension in detail two theoretical examples of brittle deformation are presented (Fig. 2.3) and used to explain the workflow for heterogeneity analysis.

Fig. 2.3a and b show two examples of the same population of extensional structures that extend a layer of rock by the same amount but with different spatial organization. The populations consist of the same three groups of structures: (i) thin veins accommodating 1.05% extension, (ii) damage zones, comprising thick veins and pull-aparts that accommodate another 3.6% extension, and (iii) faults that accommodate a further 6% extension. In both examples the damage zones are localised around the faults but in Fig. 2.3a the faults and veins are uniformly distributed, whereas in Fig. 2.3b they are strongly clustered. The veins and faults have heaves that obey a power-law distribution (Fig. 2.4); the filled circles in Fig. 2.4 represent the scale-range of structures shown in Fig. 2.3.
Figure 2.3: Examples of spatial strain heterogeneity. (a1) and (b1) show layers that have undergone brittle extension by 10.65% accommodated by brittle structures (veins, pull-apart structures and normal faults) with a displacement-size distribution that obeys a power-law as shown in Fig.2.4. (a2), (b2), (a3) and (b3) show the same extended layers after removal of the largest-displacement structures.
Figure 2.4: Plot of cumulative numbers of fractures (>heave) versus heave for the examples in Fig. 2.3. The size-distribution obeys a power-law over the observed scale range (filled circles). The empty circles extrapolate the power-law beyond the scale-range of observation.

**Cumulative plots**

As discussed in the previous section, a simple and efficient way for presenting the spatial and displacement data on fractures recorded along scan-lines, is to plot cumulative number and heave against distance (Fig. 2.5). This represents both the position and magnitude (heave) of all the data and hence includes the most relevant information on the spatial distribution of, and the extension accommodated by, the sampled structures. The cumulative graphs (Fig. 2.5) for the two examples of Fig. 2.3 share common starting and ending points and accommodate the same total number of fractures and the same extension over the observed interval. However, the fractures and the total extension in Fig. 2.3a are distributed over a distance of about 100 m whilst in Fig. 2.3b they are accommodated within a zone of only 30 m.
Figure 2.5: (a) to (c) Cumulative number versus distance plots for the examples in Fig.2.3. (d) to (e) Cumulative heave versus distance plots for the examples in Fig.2.3. Solid lines: examples a, dashed lines: examples b. From left to right subsequent removal of the largest structures in steps of one-order-of-magnitude from the initial data-sets.
Kuiper’s method

A uniform spatial distribution of fractures or extension is represented on a cumulative plot by a straight line (Fig. 2.6, dashed line). The extent to which the observed data (Fig. 2.6a, solid line) conform to a uniform distribution can be tested using non-parametric Kolmogorov-Smirnov tests (K-S tests). The simplest of these goodness-of-fit tests is based on the maximum deviation \( D_{\text{max}} \) of the observed values from the cumulative curve of the hypothesised distribution (Conover, 1980). The value of \( D_{\text{max}} \) is strongly dependent on its position (distance) on the cumulative plot as can be seen in Fig. 2.6a to c.

For this reason Kuiper (1960) developed a variant of the K-S test that utilises \( D^+ \) and \( D^- \) which represent the maximum deviation above and below the proposed cumulative distribution function (Fig. 2.6a). Kuiper’s test uses the quantity \( V = |D^+| + |D^-| \) and is as sensitive in the tails as near the median of the cumulative curve. This means that the value of the test result is largely independent of the starting point of the scan-line in relation to the greatest concentration of fractures or extension.

To allow comparison of cumulative frequency and heave data over different lengths and scale ranges, the quantity \( V \) needs to be normalized by dividing by the cumulative total \( T \), \( V' = \frac{V}{T} \), where \( T = \text{number of fractures} \) or \( T = \sum(\text{heaves}) \), for frequency and strain analysis respectively. Fig. 2.6a demonstrates how \( V' \) is determined from a cumulative graph. This can either be done graphically by measuring the largest deviation above \( (D^+) \) and below \( (D^-) \) the straight-line uniform distribution, or in a spread-sheet by calculating the theoretical cumulative value at each point along the scan-line and subtracting it from the observed value.

To examine the spatial relationships between the larger and smaller structures (heaves/apertures) in a data-set, this procedure can be repeated several times for each scan-line with sequential removal of the largest structures from the data-set. This is particularly useful for analysis of strain heterogeneity as the largest structures cause the largest steps in cumulative heave graphs and thus may dominate the
analysis. The derived heterogeneities for fracture-frequency \( (V'_F) \) and strain \( (V'_S) \) are best plotted against the maximum heave included in each data-set to show the relationships between large and small structures (Fig. 2.7). Throughout this thesis the subscripts “S” and “F” stand for “strain” (heave/aperture) and “frequency” (number) data respectively.

**Heterogeneity at different scales**

Fig. 2.7 shows the resulting \( V' \) values for the six cumulative-heave graphs (Fig. 2.5) derived from the distributions shown in Fig. 2.3a and b. This shows the heterogeneity at different scales, the largest structures were removed from both data-sets in one-order-of-magnitude steps, and \( V'_F \) and \( V'_S \) determined for each step to analyse the scale-dependency of fracture distribution and extension-heterogeneity in the two examples (Fig. 2.7). The corresponding sketches (with the largest structures removed) are shown in Fig. 2.3a2 - a3 and b2 - b3.

Fig. 2.7a shows a clear separation of data-sets a and b. Example a (Fig. 2.7) shows uniform fracture distribution \((0.04 \leq V'_F \leq 0.13)\) over the entire scale-range whilst example b shows high heterogeneity \((0.67 \leq V'_F \leq 0.69)\) of fracture spacing. Removing the largest structures from the data-set (from \( a_1 \) to \( a_3 \) and from \( b_1 \) to \( b_3 \)) does not have a strong impact on the derived heterogeneities of fracture spacing but it has a considerable influence on the strain-heterogeneity (Fig. 2.7b).

Comparison of the two complete data-sets (Fig. 2.7b) shows that the strain heterogeneity in \( a_1 \) \( (V'_S = 0.42) \) is lower than in \( b_1 \) \( (V'_S = 0.75) \). Sequential removal of the largest structures rapidly reduces the value of \( V'_S \) in case a, but has only a minor effect in case b. Thus, small-scale extension in case \( a_3 \) is homogenously distributed \((V'_S = 0.02)\) whilst in case \( b_3 \) it is heterogeneous \((V'_S = 0.70)\). In case a, the strain is moderately heterogeneous at the fault-scale \((V'_S = 0.42)\), but homogeneous for the thin veins \((V'_S = 0.02)\), and is hence scale-dependent. The strain-heterogeneity in case b is scale-invariant, being high over the entire observed scale-range \((0.7 \leq V'_S \leq 0.75)\).
Figure 2.6: (a) Shows a cumulative graph over distance (solid line). $(D^+)$ and $(D^-)$ are the maximum positive and negative deviations of the cumulative graph from the uniform distribution (dashed line). (b) and (c) show the same cumulative graph as (a) but with different starting points. $D_{\text{max}}$ is the maximum deviation of the cumulative graph (solid line) from the uniform distribution (dashed line).
The described workflow based on Kuiper’s test is a robust method to quantify heterogeneities in cumulative data-sets. The significance of the results however depends on the sample-size. This is quite obvious, given that a sample with $n = 1$ would give the highest possible heterogeneity of $V' = 1$ because the cumulative curve would consist of only one large step.

To establish whether the determined heterogeneities are statistically significant, Stephens (1965) proposed a parameter $V^* = n^{\frac{1}{2}} V'$ for Kuiper’s test and tabulated critical values for rejection of the null hypothesis of the uniform distribution (see also Mardia, 1972). This indicates whether the heterogeneity ($V'$) is significantly different from a uniform distribution given the sample size ($n$). For large $n$, say $> 50$, the critical value is $V^* \approx 1.7$ at the 0.05 (5%) level. Data-sets that are not significantly different from a uniform distribution are either homogeneous or have too small a sample-size to allow rejection of this hypothesis.

In Fig. 2.8 the determined values of $V'_F$ and $V'_S$ for the examples a and b from Fig. 2.3 are plotted against sample-size; the three curves in the diagram represent the critical values for rejection of the null hypothesis of the uniform distribution at confidence levels of 90%, 95% and 99.5% respectively (Stephens, 1965). It can be seen...
Figure 2.8: Plot of heterogeneity measures ($V'$) with respect to the sample size $n$ for the examples from Fig. 2.3. $V'_F$: spacing heterogeneity, $V'_S$: strain heterogeneity. The three curves labelled 0.995, 0.95 and 0.90 are the critical values for Kuiper’s test (Stephens, 1965) for the probabilities with 90%, 95% and 99.5% confidence that a data-set is significantly different from a uniform distribution. Values below the lines are not significantly heterogeneous whilst values above the curves are significantly heterogeneous.

that both fracture and strain distributions of the complete data-set $b_1$ are significantly different from a uniform distribution and thus are significantly heterogeneous. Removing of the largest structures reduces the heterogeneity of fracture spacing so that in $b_3$ fractures are uniformly distributed whilst strain heterogeneity remains high. The fracture distribution in example a shows no significant heterogeneity and thus can be considered as uniform whilst the strain-distribution is significantly heterogeneous.
2.5 Estimation of total brittle extension of a region

The presented methods for sampling and heterogeneity analysis so far are based on the observation of heave distributions along sample-lines, and thus, are an essentially one-dimensional consideration of extensional strain. Chapter 6 expands the strain analysis to two and three dimensions and Chapter 7 discusses other aspects of extension related deformation. At this point however, the focus remains on one-dimensional (longitudinal) extension which can directly be determined from line-data recorded in the field.

Elongation \( E \) is defined as the change in length of a straight line from an initial length \( L \) to a final length \( L' \) (Fig. 1.5).

\[
E = \frac{L' - L}{L}
\]  

(2.3)

The same equation can be expressed in terms of observed (final) length \( L' \) and cumulative heave \( H \) (corrected for oblique sampling):

\[
E = \frac{\sum H_i}{L' - \sum H_i}
\]  

(2.4)

Relative contributions of structures of a certain size

The brittle extension of a region consists of the extension accommodated by faults and tensile fractures (veins). To quantify the relative importance of smaller and larger structures to the total extension it is useful to determine the extension accommodated by structures within structure-size intervals (e.g. half-order-of-magnitude heave-intervals). Relative contributions of faults of a certain size to the total brittle extension can be directly determined by “binning” the heave-data collected in map-scale and long cliff-scale sections, as these lines record the regional fault-strain.

The contributions of faults and veins below the resolution of regional sections can be estimated from shorter cliff-scale and bed-scale sections that provide local samp-
ples of the regional deformation. As discussed in section 2.3, it is usually possible to distinguish different "deformation types" within a study-area. Most commonly these are "damage-zones" in which most of the regional extension is taken up, compared to "background-strain zones" that accommodate significantly less extension. The strain accommodated by small structures (faults and veins) within a certain heave-interval can be estimated in three ways:

1. The first method assumes that damage and background strain zones have been sampled in approximately the right proportions to represent the regional strain heterogeneity. This can readily be checked by summing the total distance sampled within higher and lower strain zones and comparing it with the regional distributions derived from map-scale and long cliff-scale sections. In this case the data from all shorter cliff-scale and bed-scale sections can be added to generate one “representative sample” of small-scale damage. The mean extensions accommodated by structures within heave-intervals can then be extracted from this data-set.

2. The second method is similar to the first one but instead of generating one representative sample of small-scale deformation, higher and lower strain zones are analysed separately. This is the better option where higher and lower strain zones have not been sampled in the right proportions.

3. The third method is particularly useful where all of the small-scale damage is accommodated within (narrow) damage zones associated with larger faults. In that case a “scaling law” can often be established that relates the amount of extension accommodated within a damage zone to the displacement (heave) of the associated fault. The total extension accommodated by small-scale structures is then expressed as a fraction of the fault-heave.
Total regional extension

From the “extensions per heave interval” it is straight forward to determine the overall brittle extension in the study area. The total extension accommodated by faults can readily be determined by summing of the fault interval-strains. The contributions of smaller structures are added depending on the above methods. In case:

1. the interval-strains can simply be summed up, together with the fault-interval strains to give the total extension.

2. the regional extent (proportions) of higher and lower strain zones are estimated from regional lines (map and cliff-scale). By multiplying the small-scale interval-strain with these proportions and adding the overall fault-strain the total extension is determined.

3. the extension accommodated within damage-zones is added directly to the fault interval-strains before these are summed to give an estimate of the total regional extension.

Likewise it is also possible to determine bulk strains for damage-zones and background-strain zones (or other zones of interest) separately.
Chapter 3

Kimmeridge Bay Field Study

3.1 Introduction

The coast around Kimmeridge Bay in Southern England (Fig. 3.1) provides a perfect cross-section through a system of extensional north-south trending normal faults and associated fractures (Donovan and Stride, 1961). The generally east-west trending coastline intersects the extensional faults and fractures at a high angle, with cliffs providing two continuous sections of several kilometres length (Fig. 3.2a). The mud-stone dominated stratigraphy of the Kimmeridge Clay is interrupted by numerous competent carbonate beds with thicknesses of 0.1 m to 2 m causing a strong mechanical layering of more ductile mudstones and brittle carbonates. Hunsdale and Sanderson (1998) have studied scaling relationships of the exposed fault population and used a high-resolution seismic reflection line to examine offshore faulting (Hunsdale et al., 1998). The available seismic data in addition to the well exposed cliff sections and the well known stratigraphy (Morgans-Bell et al., 2001) with numerous distinct marker beds make it possible to examine the extensional fault system around Kimmeridge Bay over a displacement-scale range of 4 orders of magnitude. Including bed-scale veining associated with normal faulting extends the data-set to more than 6 orders of magnitude and permits the detailed examination of spatial and scaling relationships between seismically resolvable faults and sub-seismic-scale structures, which is the aim of the present study.
3.2 Geological overview

The type-section of the Upper Jurassic Kimmeridge Clay at Kimmeridge Bay (Fig. 3.1) consists mainly of mudrocks with some intercalated white coccolithic limestones and minor grey and yellow limestones and dolostones (Morgans-Bell et al., 2001). The shale/carbonate ratio of the exposed section is about 13/1. The rocks form part of the Wessex Basin which underwent north-south extension from Permian to mid-Cretaceous times. This extension phase lead to the development of a se-
Figure 3.2: Outcrop pictures from the study area. (a) Cliff-section to the east of Kimmeridge Bay, view towards west. (b) Regularly spaced thin veins in carbonate bed. (c) Small pull-apart within carbonate bed connecting shear-fractures in the shales above and below. (d) Damage zone with Mode I failure (veins) in the carbonate bed and distributed shear failure in the under and overlying shales. (e) Small fault with prominent pull-aparts in carbonate beds.
ries of east-west trending normal faults, sometimes reactivating basement structures (Chadwick, 1985; Karner et al., 1987; Lake and Karner, 1987; Underhill and Stoneley, 1998). Active crustal extension ceased in the Aptian, and the region underwent widespread subsidence, presumed to result from thermal relaxation after the Jurassic and Early Cretaceous crustal stretching (Chadwick, 1993). Across the Wessex Basin a thick, flat-lying post-rift sequence of marine sandstones, shales and chalk were deposited during Aptian to Danian time.

In the lower Tertiary there is evidence for a change of structural style in the Wessex Basin. The east-west striking normal faults were reactivated in response to north-south compression (Chadwick, 1993) leading to widespread inversion structures, such as the Purbeck Anticline that lies just to the north of the sections examined in this study (Fig. 3.1).

During Oligocene-Miocene times, the north-south compression led to the development of conjugate strike-slip faults in some areas (e.g. North Somerset - Peacock and Sanderson, 1998; Lyme Bay - Harvey and Stewart, 1998) and to conjugate, north-south trending extensional faults in the Weymouth Bay region (Donovan and Stride, 1961; Hunsdale and Sanderson, 1998). This late extensional event produced the normal faults and veins that are the subject of this chapter.

### 3.3 Data

Normal faults and tensile fractures (veins) from the same deformation event, observed in the study area, show displacements ranging over 6 orders of magnitude. Data-sets (scan-lines) of different length and resolution were used to capture this wide range of scales (Fig. 3.3). A high-resolution seismic section (Hunsdale et al., 1998) covers the regional fault-strain and two long cliff-scale sections provide the link between regional fault-strain and localized smaller-scale damage. 8 high-resolution scan-lines were used to sample bed-scale strain accommodated in damage zones (Fig. 3.2d) and background-strain zones (Fig. 3.2b). The following section describes the
Figure 3.3: Plot of sampled range of heaves (compare with Table 3.1) against traverse length for all line samples. Chirp and fault lines sample faults with $0.1 < \text{heave} < 200$ m over distances of about 2.1 to 15.7 km. Vein lines trace single beds at great distance from faults, encountering only veins with a heave-range of 0.1 to 10 mm with traverse lengths ranging from about 30 to 140 m. Integrated lines sample both veins and faults as they trace single beds through fault zones. Thin sections locally add $>2$ orders of magnitude to the lower end of the scale range.

data-sets, and discusses the observations made at the respective scale, in detail.

### 3.3.1 Regional fault-lines

A high-resolution seismic section, hereafter referred to as *Chirp line*, was acquired using a Chirp system (Hunsdale et al., 1998). This trends parallel to the Dorset Coast and is oriented at a high angle to the main fault trend (Fig. 3.1). It has a length of 15.7 km and records 153 faults. All faults with throws $\geq 1$ m are resolved in this data-set. Recorded displacements range from 221 m to 0.5 m. The vertical separation of prominent reflectors was used to determine fault-throw, which was converted to heave (bedding parallel displacement component) using the fault dip. The cumulative heave plot (Fig. 3.4a, solid line) for these data represents a total
resolved fault-strain of 6.7%. The section is dominated by a high-strain zone between 10.5 and 12 km that takes up about two thirds of the extension. The rest of the extension appears to be continuously distributed along the line at length scales $\gg 100$ m. The 10 km section to the west of the high strain zone shows an extension of about 4% whereas to the east, closer to Kimmeridge Bay, it is lower at $<2\%$. The cumulative-number graph (Fig. 3.4a, dashed line) shows that fault-spacing is relatively homogenous with no increase within the high-strain zone and an average spacing of about 100 m.

### 3.3.2 Cliff-scale fault-lines

Two cliff sections were measured over 2.14 km and 3.95 km to the west and east of Kimmeridge Bay, respectively (Fig. 3.1). These lines record virtually all faults along the sections and are hereafter referred to as *fault lines*. The continuous exposure along the sampled sections permits recording of all faults with displacements $\geq 0.2$ m. Fault-heaves were directly measured where displacements are small ($\leq 5$ m) and calculated from stratigraphic separation and fault dip where displacements are high ($\geq 5$ m). The stratigraphic separation is defined as the stratigraphic thickness between horizons on either side of a fault that were brought next to each other due to the displacement on the fault. The measuring error for the direct tape-measurements is estimated with $\leq 0.1$ m. By using the detailed graphic logs of Morgans-Bell et al. (2001) to correlate marker-bed across larger faults, stratigraphic separations and thus displacements, can be estimated with accuracies of about $\pm 0.2$ m. The two *fault lines* show lower total extension than the *Chirp line*, due mainly to them not crossing any major fault zones. The largest faults encountered have heaves of about 40 m (*Western fault line*) and 5.4 m (*Eastern fault line*), compared to the maximum heave of 192 m on the *Chirp line*. However, the average fault spacing and extensions recorded for the *fault lines* compare well with the lower strain zones (0 to 10 km and 12 km to end) surrounding the high strain zone in the Chirp data set. The *Western fault line* (Fig. 3.4b) has a length of 2.14 km and a total resolved...
Figure 3.4: Cumulative plots. Cumulative heave (solid lines) and fracture-number (dashed lines) plotted against corrected distance (traverse length) for the Chirp line (a), the 2 fault lines (b, c) and the three integrated lines (d to f). Left vertical axes show the cumulative heave, right vertical axes show the cumulative number of encountered structures along each line.
fault-strain of 4.3%, which is comparable to the western portion of the Chirp line (0 to 10 km). It encounters 16 faults over a distance of about 2.14 km, i.e. an average fault spacing of 134 m. Both strain distribution and fault-spacing (Fig. 3.4b) appear heterogeneous at this scale of observation with one fault in the centre of the line accounting for about 2/5ths of the total resolved extension.

The Eastern fault line (Fig. 3.4c) encounters 22 faults over a distance of 3.95 km, i.e. an average fault spacing of 180 m. The total resolved extension along this section is only 0.8%. Both, strain and fault-spacing are fairly homogenously distributed along the line.

### 3.3.3 Bed-scale vein-lines

Eight data-sets were collected by tracing single beds over distances of 30 to 150 m, sampling all structures observable with the naked eye (thickness ≥ 0.1 mm). To cover the entire variability in strain, both low-strain zones (background-strain vein-lines) deformed almost exclusively by veins, and damage-zones that comprise both faults and veins (damage-zone vein-lines), were sampled. Vein apertures were measured using a 10x magnifier with a scale graduated to 0.1 mm.

**Damage-zone vein-lines**

The three damage-zone vein-lines (Fig. 3.4 d to f) cover a total length of about 350 m. All of them trace single beds across small fault zones with heaves between 1.1 m and 2.5 m. The wall-rock deformation adjacent to the faults is represented by zones of slightly increased strain compared to the background extension and can be seen as a change in slope of the heave-distance graph. The increase in strain is mainly accommodated by greater dilation of veins rather than by nucleation of additional veins as can be seen from the cumulative number versus distance graphs. Line D-3 shows a damage zone with a total width of about 50 m, with the footwall showing more intense deformation than the hanging wall (Fig. 3.4d). Line D-2 shows asymmetric wall-rock deformation with most of the fault-related strain accommodated in
its footwall within a zone of about 15 m (Fig. 3.4e). Line D-1 displays symmetric wall-rock deformation extending about 15 m into hanging wall and footwall of the fault (Fig. 3.4f).

The fracture spacing along each line is fairly constant as can be seen in the cumulative number versus distance graphs that are close to straight-line trends (Fig. 3.4d to f, dashed lines). The extensional fractures in these lines display a positive correlation between fracture-spacing and layer-thickness (Table 3.1). The average fracture spacing increases from 0.17 m in a 0.1 m thick bed (line D-3) to 0.4 m in a 2 m thick bed (line D-2).

**Background-strain vein-lines**

The five background-strain vein-lines (Fig. 3.5) sample a total length of about 340 m. They trace single beds within un-faulted regions at distances of at least 50 m from the closest fault. Extension along these lines is accommodated by veins with apertures between 0.1 mm and 10 mm, but most commonly these are between 0.1 mm and 0.5 mm. Line B-3 (Fig. 3.5a) shows the widest range of vein thicknesses ($10^{-4}$ to $10^{-2}$ m) and relatively high variation in fracture spacing. All other lines show more regular distributions of both extension and fractures, with average fracture spacing between 0.1 m and 0.3 m and cumulative extensions between 0.07% and 0.24%. No direct relationship between fracture spacing and layer thickness was found. The observed variation may be related to varying distances from the closest fault zone and/or may be due to different lithologies (different rheological behaviour) of the sampled beds.

**3.3.4 Micro-strain (Thin sections)**

Twelve samples for thin sectioning were taken from the bed-scale sections, both within damage zones and far from faults. The aim was to investigate the contribution of micro-fractures (with thicknesses $\leq 0.1$ mm). No fractures with apertures $\leq 0.1$ mm that could not be resolved with naked eye were observed in thin section. Thus,
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<th>sample size [1]</th>
<th>min heave [m]</th>
<th>max heave [m]</th>
<th>Σ heaves [m]</th>
<th>bed thickn.</th>
<th>av. fract. spacing</th>
<th>pos. start</th>
<th>pos. end</th>
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<td>2.22</td>
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<td>137.0</td>
<td>493</td>
<td>1.00E-04</td>
<td>1.00E-03</td>
<td>9.51E-02</td>
<td>0.3</td>
<td>0.257</td>
<td>E: 392424,N: 77577</td>
<td>E: 392232,N: 77642</td>
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<tr>
<td>B-3</td>
<td>70.6</td>
<td>167</td>
<td>1.00E-04</td>
<td>1.00E-02</td>
<td>1.34E-01</td>
<td>0.2 (0.4)</td>
<td>0.423</td>
<td>E: 394141,N: 77293</td>
<td>E: 394077,N: 77224</td>
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<tr>
<td>B-4</td>
<td>60.2</td>
<td>321</td>
<td>1.00E-04</td>
<td>4.00E-03</td>
<td>5.59E-02</td>
<td>0.55-0.65</td>
<td>0.188</td>
<td>E: 399785,N: 76954</td>
<td>E: 399726,N: 76969</td>
</tr>
<tr>
<td>B-5</td>
<td>29.2</td>
<td>160</td>
<td>1.00E-04</td>
<td>2.00E-03</td>
<td>6.92E-02</td>
<td>0.35 (0.55)</td>
<td>0.184</td>
<td>E: 398456,N: 798427</td>
<td>E: 398427,N: 76629</td>
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</tbody>
</table>

Table 3.1: Summary of field observations and positions of the 11 line samples. Coordinates refer to the British National Grid and were determined using an EGNOS-enabled hand-held GPS with accuracies estimated to be $\leq 8$ m. Bed thicknesses in parentheses are cumulative thicknesses of two to three closely spaced beds where some of the veins are confined by the entire package rather than by a single bed.
Figure 3.5: Cumulative heave (solid lines) and number (dashed lines) plotted against corrected distance (traverse length) for the 5 background-strain vein-lines. Left vertical axes show the cumulative heave, right vertical axes show the cumulative number of encountered structures along each line.
it is concluded that veins with apertures <0.1 mm are very rare in the Kimmeridge Clay, and that field observations in good outcrops record all of the significant brittle deformation. This result is supported by the absence of deformation of ammonites, even close to faults. If there was a measurable “micro-scale” component of extension it would be recorded by shape changes of bedding-parallel ammonites. Based on these observations a lower limit for the displacement scale range in the Kimmeridge area can be given with 0.1 mm thickness (heave).

3.4 Discussion of results

3.4.1 Fracture distribution

The following section describes the spatial and size-distribution of faults and tensile fractures (veins) in the study area.

Fracture spacing

Fracture spacing at Kimmeridge Bay has been analysed using two independent methods. The first one is the determination of the Coefficient of Variation of fracture spacing \( C_V^* \). The second method determines the heterogeneity of fracture-distribution by comparing the cumulative frequency plot for each data-set with the uniform cumulative distribution by applying Kuiper’s test. Both methods are described in detail in Chapter 2 and the results are discussed below.

Coefficient of Variation:

The values for \( C_V^* \) are listed in Table 3.2 and plotted against the heave-range covered in each data-set in Fig. 3.6a. All lines apart from the Chirp line show \( 0.7 < C_V^* \leq 1 \), suggesting that the spatial distribution of tensile fractures and smaller normal faults is random or slightly anti-clustered. The fairly constant values of \( C_V^* = 1 \) show that the standard-deviation is proportional to the mean and thus that \( C_V^* \) scales with the structure-size. Only at the higher end of the sampled scale-range (Chirp line) a weak clustering is observed with a \( C_V^* = 1.4 \).
### Table 3.2: Summary of statistical data of the 11 line samples.

<table>
<thead>
<tr>
<th>NAME OF LINE</th>
<th>Sample Size</th>
<th>BULK EXTENSION [%]</th>
<th>Cv*</th>
<th>VF’</th>
<th>VF*</th>
<th>Stat. sign. of Vf</th>
<th>VS’</th>
<th>VS*</th>
<th>Stat. sign. of Vs</th>
</tr>
</thead>
<tbody>
<tr>
<td>chirp faultE</td>
<td>153</td>
<td>6.7</td>
<td>1.4</td>
<td>0.123</td>
<td>1.53</td>
<td>n.s.</td>
<td>0.459</td>
<td>5.68</td>
<td>***</td>
</tr>
<tr>
<td>faultW D-1 D-2 D-3 B-1 B-2 B-3 B-4 B-5</td>
<td>22 16 267 363 572 396 493 167 321 159</td>
<td>4.3 1.12 0.82 0.185 0.194 0.74 1.94 2.15 2.6</td>
<td>0.77</td>
<td>0.091</td>
<td>0.057</td>
<td>0.356 0.211</td>
<td>0.981</td>
<td>1.919</td>
<td>0.88 0.122 0.277</td>
</tr>
<tr>
<td></td>
<td></td>
<td>0.84</td>
<td>0.185</td>
<td>0.091</td>
<td>0.057</td>
<td>0.356 0.211</td>
<td>0.981</td>
<td>1.919</td>
<td>0.88 0.122 0.277</td>
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<td></td>
<td></td>
<td>0.82</td>
<td>0.194</td>
<td>0.091</td>
<td>0.057</td>
<td>0.356 0.211</td>
<td>0.981</td>
<td>1.919</td>
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<tr>
<td></td>
<td></td>
<td>0.13</td>
<td>0.241</td>
<td>0.077</td>
<td>0.047</td>
<td>0.047 0.047</td>
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<td>0.047</td>
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<td></td>
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<td>0.71</td>
<td>0.261</td>
<td>0.136</td>
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<td>0.07</td>
<td>0.311</td>
<td>0.172</td>
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<tr>
<td></td>
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<td>0.07</td>
<td>0.311</td>
<td>0.172</td>
<td>0.172</td>
<td>0.172 0.172</td>
<td>0.172</td>
<td>0.172</td>
<td>0.172 0.172 0.172</td>
</tr>
</tbody>
</table>

*CV* is the modified Coefficient of Variation of fracture spacing after Ackermann et al. (2001) and Gillespie (2003). V’ and V* are the test results for Kuiper’s non-parametric test applied to the cumulative number and heave data respectively. V’ is a measure of the departure from a uniform distribution with 0 < V’ < 1. Stars indicate statistical significance of the derived heterogeneities at probabilities of 0.005 (***) 0.01 (**) and 0.05 (*) respectively. ‘n.s.’ stands for “not significant”. Extensions accommodated within heave-intervals for each data-set are in % extension. ‘x’ means that the scale range of the respective data-set does not cover this heave-interval. The cumulative extensions accommodated by veins and faults were derived by weighting the average interval-extensions with regards to the length of the respective sample-line.
Figure 3.6: Heterogeneity analysis of fracture spacing. (a) Coefficient of Variation ($C_V^*$) versus heave-range. (b) Frequency-heterogeneity ($V'_F$) versus sample-size. The three curves in the diagram represent the critical values for rejection of the null hypothesis of the uniform distribution at confidence levels of 90%, 95% and 99.5% respectively. (c) Frequency-heterogeneity ($V'_F$) versus heave-range included in each data-set.
Kuiper’s Test:
As discussed in Chapter 2, Kuiper’s method provides a measure to determine the statistical significance of the derived heterogeneity values. In Fig. 3.6b the determined spacing-heterogeneities ($V_F'$) are plotted against sample-size; the three curves in the diagrams represent the critical values for rejection of the null hypothesis of the uniform distribution at confidence levels of 90%, 95% and 99.5% respectively (Stephens, 1965). It can be seen that most of the data-sets are only weakly heterogeneous and some are not significantly different from a uniform fracture distribution. The same data is plotted with respect to the heave-range sampled in each data-set in Fig. 3.6c. The heterogeneity ($V_F'$) is low in all data-sets ($0.07 \leq V_F' \leq 0.19$) indicating close-to-uniform (random) distribution which agrees with the derived Coefficients of Variation.

Both tests show that fracture spacing is close to random over the sampled scale-range with virtually no clustering of smaller fractures around larger ones. This suggests that tensile fractures formed largely independent of the normal faults. Only the higher Coefficient of Variation for the Chirp-line indicates that there is some clustering of smaller faults around the largest faults in the system.

Damage zones around faults

Field observations and the damage-zone vein-lines (Fig. 3.4d to f) show that extension increases adjacent to most faults. In this study damage zones are defined as regions of increased strain ($\gg 0.1\%$) compared to the observed background extension of $\approx 0.1\%$. Damage zones are found to extend typically for 10-20 m on either side of the fault (Fig. 3.7). The width of the damage-zone appears to be largely independent of the displacement on the fault for heaves between 0.1 m and 40 m. Within the damage zones virtually no increase in fracture density is observed compared to the frequencies within background-strain zones. Throughout the study area the fracture spacing is between 0.17 m and 0.4 m, and is largely independent of the local strain.
Figure 3.7: Schematic cross-section, showing the typical extent of damage zones and background-damage zones in the study area. The average fault spacing (derived from Chirp and fault lines) is about 115 m, with damage zones typically extending for about 15 m into footwalls and hanging walls, leaving a zone of background deformation of about 85 m in between two faults.

Thus, the higher extensional strains in the damage zones are largely accommodated by increased opening of existing fractures rather than by the nucleation of new fractures. In the carbonate beds, this opening often leads to the development of pull-aparts (Peacock and Sanderson, 1995) linked to small faults splaying from the main fault (Fig. 3.2c,e).

**Size-distribution of fractures**

The number of structures within half-order-of-magnitude-bins is best plotted against the heave-intervals covered in each data-set (Fig. 3.8). Fracture-frequency (number of fractures/m) shows a rapid decrease from several fractures per metre for fractures with mm-scale heaves, to <1 fault per 100 m for faults with heaves of m-scale (Fig. 3.8a). These frequencies suggest that about 500 extensional fractures (veins) can be expected for one observed fault. If only large faults are considered, such as would be observable on a commercial seismic line (say with throws $\geq 10$ to 20 m), one may expect 10 - 30 “sub-seismic-resolution faults” and 5000 to 15,000 extensional fractures for each seismically resolvable fault in the study area.

The frequency-data appear to conform to a power-law distribution as reported widely.
Figure 3.8: Frequency plots. (a) Frequency (number/m) of structures (+) plotted against heave in half-order-of-magnitude bins in log-linear space. In (b) the average frequency for each heave-interval is plotted. The 3 solid lines are best fit power-law trends through the data for different scale ranges.
for faults and fractures (Walsh et al., 1991; Jackson and Sanderson, 1992; Marrett and Allmendinger, 1992; Pickering et al., 1994). In Fig. 3.8b the average number of fractures in each bin is plotted on a log-log scale. The larger faults (≥ 10 m heave) conform to a power-law distribution, as do the veins (< 0.1 m aperture), both having scaling exponents of ≈0.95, with correlation coefficients $R^2 > 0.98$. Note that these plots are for discrete bins and not the more conventional cumulative plots where the ranking of the data always generate high $R^2$ values (Pickering et al., 1995).

However, the veins and faults are clearly separated by a transitional region at heaves of between 0.1 and 10 m (Fig. 3.8b) and thus do not form part of the same power-law distribution. A similar transition was found by Hunsdale and Sanderson (1998) in their study of the scaling of fault frequencies at Kimmeridge, but this was between faults cutting many beds and micro-faults confined to a single carbonate bed (the Whitestone Band). The vein population in the present study represents opening mode fractures that are developed in most of the carbonate beds and have a higher scaling exponent than that of the layer-confined micro-faults (Hunsdale and Sanderson, 1998).

Extrapolation of the fault-frequencies to the mm-scale would predict about 10 fractures per meter. Such high fracture-densities are not generally observed in this study area, even within damage zones, and deformation at this scale is accommodated by 1-3 veins per metre.

### 3.4.2 Strain distribution

The following section describes the spatial heterogeneity and size-distribution of brittle extension in the study area.

**Spatial heterogeneity of extensional strain**

As discussed in Chapter 2, Kuiper’s method is applicable to analyse the heterogeneity of strain-distributions and it also provides a measure of the statistical significance of the derived heterogeneities. Thus, before discussing the results of variation
in strain-heterogeneity, it should be established that the derived heterogeneities are statistically significant with respect to the magnitude of $V'_S$ and the sample size, $n$. Values of $V'_S$ are listed in Table 3.2 together with an indication of their statistical significance. The statistical significance is also shown in Fig. 3.9a for the 90%, 95% and 99.5% probability levels. Only the Eastern fault line and line B-4 are not significantly different from a uniform distribution (at the 0.05 level), with all other lines showing a significantly heterogeneous strain distribution.

Having established the statistical significance of the determined heterogeneities the results of Kuiper's test applied to the 11 cumulative heave curves (Fig. 3.9b) can now be discussed. $V'_S$ values for all lines are plotted against the maximum heave included in each data-set (Fig. 3.9b, circles). In the background-strain vein-lines, extension is accommodated by opening of tensile fractures (veins), with apertures in the range 0.1 mm to 10 mm. The heterogeneity is weak ($V'_S < 0.2$), but the large sample sizes make these values statistically significant (Table 3.2, Fig. 3.9a).

In those vein-lines which cross damage-zones of one or more small faults, strain is mainly accommodated by small faults and pull-aparts with heaves of 10 mm to 100 mm. These sections show much higher heterogeneity ($V'_S > 0.5$).

The heterogeneity for each line has also been analysed by successively removing the largest faults from the data-set in half-order-of-magnitude steps. The derived $V'_S$ values are displayed as a series of 'tails' to the left of the complete data-sets (Fig. 3.9b). One interesting feature of this plot is that the 'tails' tend to link the different types of sample lines. Thus, elimination of the larger faults in the Chirp-line reduces the heterogeneity towards that of the fault-lines and elimination of the faults in the damage-zone vein-lines reduces the heterogeneity towards that of the background-strain vein-lines. There is even some suggestion that elimination of larger faults in the fault lines leads to increasing heterogeneity (towards that of the damage-zone vein-lines), but the reduction in sample size from the already small numbers of the fault-lines increases the uncertainty in the values of $V'_S$. The general dependence of strain-heterogeneity and fault-size is indicated by the grey shaded arrow in Fig.
Figure 3.9: (a) Heterogeneity of strain distribution ($V'_S$) versus sample-size. The three curves in the diagram represent the critical values for rejection of the null hypothesis of the uniform distribution at confidence levels of 90%, 95% and 99.5% respectively. (b) $V'_S$ values plotted against the maximum heave for each sample line. The values for complete data-sets are shown as circles. The ‘tails’ to the left of each circle were derived by successively removing the largest structures (heaves) from the complete data-sets in half-order-of-magnitude steps ($\approx \log_{10} 3$). The grey arrow emphasises the observed scale dependence of heterogeneity.
3.9b.

In the Chirp line and the two fault lines most of the strain is accommodated on faults with heaves >1 m (Figs. 3.10, 3.11). Strain accommodated along these lines shows some heterogeneity (0.2 < $V'_S < 0.5$), but the small sample size of the fault lines makes these estimates less certain and in particular for the Eastern fault line a uniform strain distribution cannot be rejected. There is some suggestion that heterogeneity may be increasing at the higher end of the sampled scale-range (>100 m heave), with localisation of strain onto the largest faults (Fig. 3.9b).

Strain scaling

*Bin-strain:*

The sample-lines were designed to record structures at different scale-ranges, allowing the strain, accommodated in different heave-intervals, to be determined. To assess the proportion of the total strain that is accommodated by structures of a certain size range, cumulative heave versus distance plots for all data-sets were constructed with successive removal of the higher-heave structures. Examples of these diagrams are shown for the Chirp line and the two fault lines in Fig. 3.10. The uppermost curve in each diagram includes all sampled structures. The other lines are cumulative-heave plots excluding faults above certain displacement values in steps of half-an-order of magnitude. The left-hand vertical axes show the cumulative heaves, whilst the right-hand vertical axes show the total resolved extensions. The successive removal of the larger-heave faults significantly reduces the overall extension, especially down to about 10 m. Consequently, the cumulative curves also become smoother showing that the extension accommodated by smaller-scale structures is more homogenously distributed as discussed in the previous section (Figs. 3.9b, 3.10).

The extensions for each half-order-of-magnitude bin are listed in Table 3.2 and plotted against heave-intervals in Fig. 3.11 (crosses). This shows the range of observed extension values accommodated by structures of different size. The solid lines in
Figure 3.10: Cumulative heave versus corrected distance plots for the Chirp (a) and fault lines (b, c) with successive removal of the largest fault-heaves in half-order-of-magnitude steps ($\approx \log_{10} 3$). The uppermost curve in each diagram includes all sampled faults of the respective line. The right hand vertical axes show the bulk extension accommodated along the lines. Labels on the graphs indicate the included heave-ranges.
Figure 3.11: Extensions for half-order-of-magnitude bins ($\approx \log 10^3$) plotted against heave-intervals in log-linear space (a) and log-log space (b). The solid lines represent the average extension values plotted as a histogram in (a) and as a moving average in (b).

Fig. 3.11 represent the average extension values per bin. The plots show that the extension in each heave-interval is low ($\approx 0.1\%$) up to a heave of about 0.1 m and is accommodated entirely by the opening of veins. The data are scattered, reflecting the varying contributions of veins to the background strain and to the development of damage zones around faults. Above about 0.3 m the extension is accommodated by faults and increases in each bin with increasing heave. The data are much less scattered, indicating that the faults show a fairly well-ordered scaling, with larger faults contributing more to the overall strain. An interesting observation from the Chirp and fault lines is that almost independent of the total strain in each section, faults with heaves $<3$ m accommodate $<1\%$ extension. On the other hand, the larger the encountered faults in any section, the higher the total extension. This clearly shows that extension is localized by concentrating slip on large faults rather than by increasing the number of small fractures. This has also been observed for faults on seismic sections (e.g. Meyer et al., 2002).

**Total brittle extension:**

Based on the observations of the distribution and scaling of fractures, it is now possible to analyse the relative contribution of smaller and larger structures to the total brittle strain of the region. A key to achieving this is to assess if faults, damage
zones and background strain zones have been sampled in a way that is representative of the entire population.

The fault-data \textit{(Chirp and fault lines)} cover a total sampled distance of 21.9 km. Based on these data the average fault spacing at Kimmeridge Bay is about 115 m and is similar on both types of lines (Fig. 3.7). The \textit{Chirp line} and the two \textit{fault lines} also cover a similar heave-range and do not overlap. Hence, each of these data-sets can be treated as an independent sample of the faulting and they can simply be added together to provide estimates of the regional fault-density and fault-strain.

Field observations along the cliff sections and \textit{damage-zone vein-lines} (Fig. 3.4d to e) show that wall-rock deformation adjacent to faults typically extends for about 10 m to 20 m (average observed extent: 15 m) into the footwall and hanging wall and that the extent of these deformation zones is largely independent of the fault size (Fig. 3.7). Based on these observations, the approximate ratio of damage zones to background-strain regions can be estimated as 30 m : (115 - 30 m) \approx 1 : 3 (Fig. 3.7). Within the Kimmeridge area, the small-scale deformation (veins and joints) was sampled in 8 traverses over a total distance of 690 m. This comprises about 180 m within damage zones with extensions >0.1\% and 510 m with background strains (\leq0.1\% extension), giving a ratio of damage zones : background strain of \approx 1 : 3. Thus, at least approximately and quite fortuitously, high-strain and background-strain regions have been sampled in about the correct proportions.

Having established that the 11 sample-lines as a whole are a representative sample of the entire scale-range of brittle extension in the study area, the average extensions per heave-interval (Table 3.2) can be used as a basis to estimate the relative contributions of larger and smaller structures to the total brittle extension. To do this, the extensions per heave-interval were weighted based on the length of the sample-line and thus based on the sample size. The estimated cumulative extensions are given in Table 3.2. Veins (heave <0.1 m) accommodate about 0.5\% extension. Small faults with heaves between 0.1 and 10 m take up another 2\%. Thus, structures of size below seismic resolution account for a total extension of about 2.5\%.
Larger faults, with heaves >10 m, accommodate about 4.5% and thus account for the largest portion of the total extension which is about 7%.

### 3.4.3 Evolution of early-stage extension-related damage

Based on the field observations and analysis of fracture and strain distributions a conceptual model of the early-stage extensional structures can be defined (Fig. 3.12). Early extension is accommodated by distributed tensile failure of the carbonate beds throughout the entire region whilst the shales appear to accommodate deformation by shear failure on many planes (Fig. 3.12a). Increased extension localises displacement onto a few regularly spaced minor faults and opens veins to form pull-apart structures (Figs. 3.12b, 3.2c). Finally some fault planes break through and further increase in strain is accommodated dominantly by slip on these planes (Figs. 3.12c, 3.2e).

### 3.5 Summary and Conclusions

Spacing analysis based on the Coefficient of Variation and Kuiper’s Test shows that veins and smaller faults are randomly distributed in the study area over 6 orders of displacement-size magnitude. Only at the high end of the scale range some clustering of damage around the largest faults in the region can be observed. This study yields three different patterns of strain distribution:

1. Tensile fractures are widely distributed with fracture spacing in carbonate beds varying between 0.1 m and 0.4 m with an average of about 0.3 m. The variation is probably caused by differences in lithologies and thicknesses between the beds. Opening of these fractures produces thin veins, usually ≤0.5 mm in aperture, accommodating a background extension of about 0.1% across the area. This strain can be neglected in terms of the overall extension, but may be of interest for interpretation of the early deformation history and the permeability of the rock units. Extension is fairly homogenously distributed
Figure 3.12: Conceptual model of the early-stage evolution of extension-related damage in the mudstone-dominated rocks around Kimmeridge Bay. The sketches show the evolution from low (a) to higher extension (c).
(0.1 ≤ \(V_s'\) ≤ 0.3) across the 0.1 to 10 mm range of vein opening.

2. Extension increases in the damage zones around faults with only a small increase of fracture density. The damage zones typically extend 10-20 m into the wall-rock to either side of the fault. The widths of the damage zones appear to be largely independent of the displacement on the fault. Deformation in these zones is mainly achieved by opening (dilation) of fractures, to produce veins with apertures of 0.1 to 10 mm, rather than the opening of additional fractures. Pull-apart structures are common in the damage zones and produce veins with up to 100 mm thickness. Due to the inter-layering of mudstones and limestones, the interaction of faulting and opening-mode fractures produces very heterogeneous strain distributions (\(V_s' > 0.5\)) within damage zones around faults. This behavior may not apply in other lithologies, such as porous sandstones where strain hardening may lead to nucleation of new faults rather than to opening of veins.

3. Faults with heaves of 1-100 m are developed at a spacing that usually exceeds 100 m. They produce only weak to moderate heterogeneity (0.2 ≤ \(V_s'\) ≤ 0.4) in the strain distribution. There is some evidence of heterogeneity increasing with increasing fault heave, suggesting that strain is localized onto large faults at the basin-scale.

Thus the brittle deformation in the studied area occurred over a wide range of scales and was accommodated by both, veins and faults. Power-law scaling of the frequency of both veins and faults is found, but these do not share a single power-law relationship. A major change in scaling occurs at the transition between extensional (Mode I) and shear (Mode II/III) deformation. Thus, it is not possible to simply extrapolate extensions and fracture frequency from one scale-range to the other across structures with differing spatial organization and mechanical significance.

A conceptual model for the early-stage evolution of extension-related damage in this interbedded mudstone-carbonate sequence can be described as follows: The
mud-stone dominated sequence accommodates initial deformation in a distributed manner which produces a uniform low background-strain throughout the region. As extension increases, displacement localises onto fewer, regularly spaced, shear planes and opens veins to form pull-apart structures and minor faults. Further increase in strain is almost exclusively accommodated by slip on these faults.

Fractures are responsible for producing a total brittle extension of about 7% in the rocks at Kimmeridge Bay which is accommodated by slip on faults and opening of extension fractures to produce veins. The relative contribution of these structures to the overall extension in this inter-bedded mudstone-limestone sequence varies, as follows:

- Large faults, with heaves of >10 m, accommodate about 65% of the total extension, which would be resolvable by high-quality commercial seismic surveys.
- Smaller faults, with heaves 0.1-10 m, accommodate about 28% of the total extension.
- Opening mode fractures (veins) are very numerous, but account for only 7% of the overall extension.
Chapter 4

Kilve Field Study

4.1 Introduction

The Somerset Coast provides a world-class natural laboratory for studying the geometry and evolution of brittle structures. The large tidal range has produced a wide (>100 m) wave-cut platform providing a horizontal cross-section in addition to a vertical cliff-section. This unusual outcrop situation provides detailed three-dimensional information on the geometries of brittle structures. Due to the fast erosion of the relatively soft rocks the outcrops are always fresh and provide excellent conditions to measure displacements from vein to fault-scale.

Most of the Bristol Channel Basin records a poly-phase deformation history with N-S extension followed by N-S compression. The exposed brittle structures were produced by the basin-forming N-S extension during the Mesozoic and by the subsequent inversion during the Alpine Orogeny. The region therefore provides an excellent example of the history of a small-scale sedimentary basin. Due to the exceptional quality and accessibility of the exposure, the Liassic inter-beded limestones and shales around the Bristol Channel are amongst the most studied rocks in the world and are ideal rocks in which to conduct a detailed scaling study, as they provide:

- Excellent exposures in the cliffs and wave-cut platforms with several kilometres
of continuous exposure.

- Abundant faults, veins and fractures whose geometry and deformation history have been well documented
- High resolution stratigraphy and correlation of beds that allow accurate estimates of displacements from 0 - 100 m (Whittaker and Green, 1983).

The ca. 2 km long section between Kilve and Lilstock (Fig. 4.1), which was chosen for this study, records dominantly the earlier extensional phase with little indication of reverse reactivation of the Mesozoic structures. The section is a relatively high-strain zone within the Bristol Channel Basin and the observed structures range from ≪mm thick veins to faults with ≫10 m heave, providing a data-set of close to 5 orders-of-displacement-scale magnitude. The aims of this study are to analyse the spatial distribution of normal faults and associated veins, in the inter-bedded mud-stones and carbonates, over the entire observed scale-range. Specifically are examined:

- The relative contributions of faults and veins to the total brittle extension
- The spatial distribution of faults and in particular the distribution of faults with size below and close to seismic resolution
- The spatial relationship between faults and veins
- The spatial distribution and heterogeneity of brittle extension

4.2 Geological overview

The sedimentary sequence exposed in the cliffs and the inter-tidal platform (Fig. 4.2a) around Kilve (Fig. 4.1) consists of Triassic marls and Jurassic limestones, shales and marls (Whittaker and Green, 1983). Palmer (1972) divided the sequence based on lithological differences. Most prominent is the limestone/shale interbedded Blue Lias in the studied sequence, above and below which are the mudstone
dominated units of the St. Audrie’s and Kilve Shales, respectively (Fig. 4.3). The shale/carbonate ratio of the examined section is about 5/1.

The carbonate beds within the stratigraphic successions provide useful marker-beds that can be traced and correlated across the entire study area. Fault-heaves were directly measured where displacements are small (≤5 m) and calculated from stratigraphic separation and fault dip where displacements are high (≥5 m). The stratigraphic separation is defined as the stratigraphic thickness between horizons on either side of a fault that were brought next to each other due to the displacement on the fault. The measuring error for the direct tape-measurements is estimated with ≤0.1 m. By using the detailed graphic logs (Fig. 4.3) of Whittaker and Green (1983) to correlate marker-bed across larger faults, stratigraphic separations and thus displacements can be estimated with accuracies of about ±0.2 m. Associated veins have openings that can be measured to ≪1 mm in the carbonate marker-beds.

The study area is located on the southern margin of the Bristol Channel Basin (Fig. 4.1) that is part of the large Wessex Basin in Southern England. The Wessex Basin underwent north-south extension from Permian to mid-Cretaceous times leading to the development of extensional faults and veins which are subject of this study (Chadwick, 1985; Karner et al., 1987; Lake and Karner, 1987; Underhill and Stone-
Figure 4.2: Outcrop pictures: View from the top of the cliff onto the tidal platform (at low tide), towards NW (a) and NE (b) respectively. (c) Fault 59, down-throwing towards N, brings marker bed [126] (Blue Lias) next to bed [65] (Top of St. Audrie’s shales), which gives to a stratigraphic separation of 36.3 m. (d) Damage zone adjacent to normal fault with cm-thick veins and some normal fault-drag. (e) Close up on intensely veined carbonate bed in damage zone. (f) Millimetre-resolution of fault-displacement on the tidal platform (faulted ammonite). (g) Panorama-view from the tidal platform onto part of the cliff-section (fault-line).
Figure 4.3: Stratigraphy of the studied section, modified after Palmer (1972) and Whitaker and Green (1983). Prominent carbonate beds are labelled with their bed-numbers, starting with 1 at the base (bottom left). The most distinctive bed-sequences are highlighted in colour. Dashed hatching or brown shading: shales, Brick hatching or blue shading: carbonates.
ley, 1998). Major basin-bounding and intra-basinal faults strike east-west. Active crustal extension ceased in the Aptian and the region underwent widespread thermal subsidence (Chadwick, 1993). There is evidence for subsequent wide-spread inversion in the Wessex Basin during the lower Tertiary (Chadwick, 1993; Dart et al., 1995) which also reactivated earlier normal faults along some parts of the Somerset Coast.

The coastal section between Kilve and Lilstock was chosen for this study because none of the Mesozoic (pre-Aptian) extensional structures appear to have been reactivated whilst several faults to the east and west of the studied section have been inverted during the Tertiary contraction.

Based on seismic data the Bristol Channel Basin has been interpreted as a Mesozoic half-graben formed above a major south-dipping normal fault which is suggested to represent a reactivated Variscan thrust (Brooks et al., 1988). Peacock and Sanderson (1999) contradict this interpretation suggesting a hierarchical graben model with two major E-W trending faults south of the channel, plunging to the north. For the present study this disagreement is of minor importance as it concerns the basin-bounding structures rather than the sub-seismic resolution (cliff-scale) faults and fractures which are the subject of this chapter.

4.3 Data

Brittle deformation in the studied section is restricted to faults (Fig. 4.2b,c) and narrow damage-zones (Fig. 4.2d), surrounding these faults, with no observable extensional structures in between damage-zones. For this reason the data-sets required to record the entire variability of brittle damage in the study area are one cliff-scale fault-line, and a number of bed-scale sections. The cliff-scale fault-line covers the entire section and records the total fault-strain. The bed-scale scan-lines measure the extent and magnitude of damage in the wall-rock adjacent to faults of different size. Fig. 4.4 shows the line-lengths and heave-ranges covered in each data-set. Errors due to oblique sampling were corrected for by applying the methods described
by Terzaghi (1965) and Peacock and Sanderson (1993). A summary of the recorded data for all sample-lines is given in Table 4.1.

### 4.3.1 Cliff-scale fault-line

The *fault-line* (Fig. 4.1) records fault-displacements along the 1970 m long cliff-section between Kilve and Lilstock. The corrected length (i.e. perpendicular to fault strike) of this line is 1611 m. It records 75 faults with throws between 5 cm and 61 m. Due to the exceptional outcrop quality all faults with throws ≥10 cm are recorded along this section. The total fault-strain of this section is 24.75% which appears to be fairly homogenously distributed over the entire distance (Fig. 4.5a).

### 4.3.2 Bed-scale vein-lines

Given that there is no observable brittle damage outside of damage-zones, all *bed-scale vein-lines* sample the wallrock damage adjacent to, or across, faults. Most sections trace single beds either within the footwall or hanging wall of a fault and, where possible, single beds were sampled across both the hanging wall and footwall.
Figure 4.5: Cumulative heave/aperture (solid lines) and number (dashed lines) plots for (a) fault-line, (b) vein-line B1B2, (c) vein-line C1C2, (d) vein-line D2D1, (e) vein-line G3G2, (f) vein-line G5G2. Dotted lines: Uniform distribution. Grey-shaded sections: Damage zones.
Table 4.1: Properties of the 14 scan-lines. Coordinates refer to the British National Grid and were determined using an EGNOS-enabled hand-held GPS. Accuracies are estimated to be ±6 m. For short lines only one coordinate point is given, generally taken at the position of the largest sampled structure. Maximum heave values in parentheses are fault-heaves of the single fault included in short lines which cross a fault zone.

<table>
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<th>bed thickn. [m]</th>
<th>sample size [m]</th>
<th>min heave [m]</th>
<th>max heave [m]</th>
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of the same fault. Care was taken to sample the same marker beds (Fig. 4.1) along the entire cliff-section to avoid variation in the data due to differences in lithology and thickness. Damage zones were sampled spanning the entire range of fault-sizes from cm-throw faults (Lines A1 to A3), m-throw faults (Lines B1B2, B3, D2D1 and G4) up to the largest faults in the section with throws between 15, 20 m (Lines C1C2, E1 and E2) and 61 m (Lines G3G1, G5G2, G6 and G7).

Damage around cm-throw faults

Lines A1, A2 and A3 (Fig. 4.6a,d,g) are three short parallel sections with lengths between 7 and 10 m crossing a small normal fault zone that consists of several overlapping cm-throw faults with small relay ramps (Peacock and Sanderson, 1991) in between. A1 crosses the zone close to its tip, accommodating all of its extension (0.23%) by tensile veins within a narrow damage zone of only 0.6 m width. A2 crosses two cm-throw faults embedded in a 2.7 m wide damage-zone with a total extension of 1.01%. A3 crosses the centre of the fault-zone with a damage-zone width of 2.3 m and a total extension of 1.8% with tensile veins accounting for about \( \frac{1}{4} \) of the total extension.
Figure 4.6: Cumulative heave/aperture (solid lines) and number (dashed lines) plots for (a) vein-line A1, (b) vein-line E1, (c) vein-line G4, (d) vein-line A2, (e) vein-line E2, (f) vein-line G6 (g) vein-line A3, (h) vein-line B3, (i) vein-line G7. Dotted lines: Uniform distribution. Grey-shaded sections: Damage zones.
Damage around m-throw faults

Lines B1B2 and B3 are sampling the damage zones around and in between fault 3 and fault 4 of the cliff-scale fault-line. The two faults display throws of 2.9 m and 2.5 m, respectively. Line B1B2 (Fig. 4.5b) traces bed [165] from the footwall of fault 3 all the way into the hanging wall of fault 4, and encounters a total of 188 fractures. Line B3 (Fig. 4.6h) samples the footwall damage of fault 3 along bed [161], recording 78 fractures. The total extension accommodated along section B1B2 with and without the two faults is 78.7% and 2.32% respectively with veins (≤3 cm thickness) accommodating 1.52% extension. Line B3 records a total strain of 3.76%, with 2.76% accommodated by veins (≤3 cm thickness). The widths of the footwall damage zone of fault 3 and the hanging wall damage zone of fault 4 are 1.8 m and 2.8 m, respectively.

Line D2D1 (Fig. 4.5d) traces bed [165] across a fault with a throw of 3.8 m and records a total of 41 fractures over a distance of 15 m. The widths of the footwall and hanging wall damage zones are 4.7 m and 1.8 m respectively. The total extension accommodated along this section with and without the fault is 10.30% and 0.14% respectively.

Line G4 (Fig. 4.6c) starts at fault 26 (2.2m throw) and samples bed [77] over a distance of 16 metres into the hanging wall of this fault. There is virtually no deformation, apart from 2 veins within the first 0.4 m from the fault. The total strain accommodated along this section is 0.19%.

Damage around >10m-throw faults

Line E1 (Fig. 4.6b) starts at a fault with 15.6 m throw and samples bed [157] in the hanging wall of this fault. This line records 44 fractures, all of which are located within the damage zone. Line E2 (Fig. 4.6e) starts at the same fault and traces bed [126] over a distance of 7.55 m into the footwall of the fault and further into an adjacent damage zone, recording a total of 46 fractures, 15 of which belong to the footwall damage zone. The extents of hanging wall and footwall damage zones...
are 4.98 m and 3.28 m respectively. The hanging wall damage-zone accommodates a total strain of 6.64% whilst the footwall damage zone shows an extension of 3.9%.

*Line C1C2* (Fig. 4.5c) starts in the footwall of *fault 10* (18.3 m throw) of the cliff-scale fault line, tracing bed [147] into the fault and bed [165] into the hanging wall out of the fault. The total length of the line is 41.8 m. The footwall damage zone is about 9 m wide and contains 55 fractures. The first 70 cm from the fault record a strain of 8.13%, the remaining 8.3 m show a low strain of 0.15%. The hanging wall damage zone has a width of >7.5 m (end of line) and consists of 195 fractures. The highest strain (10.4%) is accommodated within the first 60 cm from the fault and a lower strain (1.39%) over the remaining 7 m.

*Line G3G2* (Fig. 4.5e) samples the damage zones around *faults 24* and 26 of the cliff-scale fault line. It starts at *fault 24* and traces bed [69] across the footwall damage zone of this fault before it crosses *fault 26* and records both, footwall and hanging wall damage of the latter. *Fault 24* is the largest fault in the Kilve cliff-section with a throw of 61 m. *Fault 26* is a smaller fault (2.2 m throw), associated with *fault 24*. The immediate footwall damage of *fault 24* has a width of 2.18 m in which 39 fractures accommodated an extension of 1.61%. *Fault 26* has very little associated damage: 8 fractures in the footwall within a zone of 2.4 m and 4 fractures in the hanging wall within a narrow zone of 20 cm from the fault.

*Line G5G2* (Fig. 4.5f) samples bed [75] and is an equivalent of *line G3G2*. As the latter it starts at *fault 24* and crosses *fault 26*. It records a 1.66 m wide footwall damage zone with 38 fractures accommodating an extension of 5.5% of *fault 24*. The damage around *fault 26* is very minor with 7 fractures in the footwall and 11 fractures in the hanging wall, accommodating strains of 0.7% and 3.5% in 80 cm and 90 cm wide zones from the fault respectively.

*Line G6* (Fig. 4.6f) traces bed [77] from *fault 24* to *fault 26* and records the footwall damage of *fault 24* and the hanging wall damage of *fault 26*. The footwall damage zone of *fault 26* consists of 27 fractures accommodating a strain of 2.21% over a distance of 2.4 m from the fault. The hanging wall damage of *fault 24* is
accommodated by 25 fractures over a distance of 3.6 m producing a strain of 1.95%.

Line G7 (Fig. 4.6i) traces bed [192] from fault 24 towards S recording the hanging wall damage of this fault. The damage zone extends 10.3 m into the hanging wall, consists of 74 fractures and accommodates an extension of 3.1%.

4.4 Discussion of results

4.4.1 Fracture distribution

The following section discusses the spatial- and size-distributions of faults and tensile fractures (veins) within the study area. A summary of the statistical data and heterogeneity analysis is given in Table 4.2.

Fracture spacing

Coefficient of Variation:

Determination of the \( C_{V^*} \) for fracture spacing allows discrimination between regular (anti-clustered), random and clustered distributions of fractures as discussed in Chapter 2. Fig. 4.7a shows the derived \( C_{V^*} \) values with respect to the structure-size range sampled in each data-set (min. heave - max. heave). At the lower end of the scale range \( (10^{-4} - 10^{-2} \text{ m heave}) \) the data show significantly clustered veining \( (1.4 \leq C_{V^*} \leq 5.40) \) compared to the spacing of faults \( (10^{-4} - 10^{2} \text{ m heave}) \) that shows somewhat less clustering \( (C_{V^*} = 1.65) \).

Kuiper’s Method:

In Fig. 4.7b the determined values of \( V_F' \) are plotted against sample-size; the three curves in the diagrams represent the critical values for rejection of the null hypothesis of the uniform distribution at confidence levels of 90%, 95% and 99.5% respectively (Stephens, 1965). It can be seen that all data-sets are significantly different from a uniform distribution.

Fig. 4.7c shows the derived heterogeneities \( (V_F') \) with respect to the structure-size range sampled in each data-set (min. heave - max. heave). It can be seen that at
Figure 4.7: Heterogeneity of fracture spacing. (a) Coefficient of Variation ($C_V^*$) versus heave-range recorded in each data-set. (b) Plot of spacing heterogeneity ($V'_F$) with respect to the sample size (n) of each data-set. The three curves labelled 0.995, 0.95 and 0.90 are the critical values for Kuiper’s test (Stephens, 1965) for the probabilities with 90%, 95% and 99.5% confidence that a data-set is significantly different from a uniform distribution. Values below the lines are not significantly heterogeneous whilst values above the curves are significantly heterogeneous. (c) Spacing heterogeneity ($V'_F$) versus heave-range covered by each data-set.
Table 4.2: Results of the heterogeneity analysis. The second column gives the average fractures spacing for each line, the third column gives the bulk extensional strain measured on each transect. Columns 4 to 10 list the results of the heterogeneity analysis for frequency ($V_F'$) and strain ($V_S'$) with indication of the statistical significance of their deviation from a uniform distribution (stars). The stars indicate the statistical significance at probabilities of 0.005 (***), 0.01 (**) and 0.05 (*) respectively. 'n.s.' stands for "not significant". $C_V^*$ is the modified Coefficient of Variation of fracture spacing.

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<th>stat. sign. of $V_F'$</th>
<th>$C_V^*$</th>
<th>$V_S'$</th>
<th>$V_S'$</th>
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<td>[%]</td>
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a heave scale-range between $10^{-4} - 10^{-2}$ m (thin veins) the data show moderate to high spatial heterogeneity ($0.37 \leq V_F' \leq 0.88$). At the fault-scale ($10^{-1} - 10^2$ m heave) the heterogeneity is much lower ($V_F' = 0.24$) indicating that the faults between Kilve and Lilstock are only weakly clustered.

Based on the two heterogeneity tests it can be said that the fracture population at Kilve displays a scale-dependence of fracture-spacing with veins being clustered around normal faults which themselves are randomly distributed.

**Size distribution of fractures**

**Faults:**

The cliff-scale *fault-line* records all faults with throws $\geq 10$ cm and samples the entire studied section between Kilve and Lilstock (Fig. 4.1). It covers structures of close to 3 orders of displacement magnitude. Fig. 4.8 shows the cumulative number of fractures in heave-bins of half-an-order of magnitude ($\cong \log_{10} 3$). 35 faults with heaves between 0.1 m to 1 m are recorded in the section. The interval of 1 m to 10 m heave includes 23 faults and 10 faults with heaves $> 10$ m are recorded. The largest structures (heaves $> 30$ m) are probably under-represented due to the length...
Figure 4.8: Log-log plot of estimated cumulative numbers of veins and faults per interval. Filled squares: Veins, circles: Faults, filled circles represent unbiased values, open circle may be biased because of under-sampling of the largest faults due to the section length < 2 km. Solid lines are best-fit power-law trends, D are the respective power-law exponents.

of the fault-line being too small to provide a representative sample of these faults.
The data appear to conform to a power-law scaling over the observed scale-range if the largest bin (fault-heaves of 30 to 100 m) is excluded to avoid bias due to under-sampling (Fig. 4.8). The scaling-exponent for faults with heaves between 0.1 m and 30 m is surprisingly low (-0.41) suggesting that the smallest faults in the section are of minor importance. However it has to be kept in mind that most of the data are below seismic resolution and very few scaling exponents have been published to date for this scale-range.

*Damage zones around faults:*
Fractures with heaves <0.1 m are mostly veins and pull-aparts (Peacock and Sanderson, 1995). As discussed in the previous sections, these fractures are located exclusively in damage zones adjacent to (larger) faults. Centimetre-throw faults are typically associated with about 10 fractures of heaves <0.1 m. Metre-throw faults are surrounded by about 50 small fractures. Damage zones adjacent to faults with throws >10 m consist of an average of about 120 fractures with heaves <0.1 m.
The majority (>90%) of fractures in all observed damage zones consist of veins with apertures ≪0.01 m. This gives a rough estimate of a total number of about 2700 fractures with heaves <0.1 m within the studied section. About 160 of these have heaves between 1 cm and 10 cm, 540 fractures lie in the range of 0.1 cm to 1 cm aperture and the majority of about 2000 fractures are thin veins with thicknesses <1 mm. Although these values are only estimates it can be seen (Fig. 4.8) that the derived numbers extend the power-law scaling relationship from the fault-scale all the way into the scale-range of tensile fractures.

4.4.2 Strain distribution

The following section describes the spatial heterogeneity and size-distribution of brittle extension in the studied section.

Heterogeneity of extensional strain

In Fig. 4.9a the determined values of $V_S'$ are plotted against sample-size; the three curves in the diagrams represent the critical values for rejection of the null hypothesis of the uniform distribution at confidence levels of 90%, 95% and 99.5% respectively (Stephens, 1965). It can be seen that the strain distribution is significantly different from a uniform distribution in all but one data-set. Only for the fault-line a uniform strain distribution cannot be ruled out.

In Fig. 4.9b $V_S'$ values for all scan-lines are plotted against the maximum heave included in each data-set. Diamonds represent the heterogeneity of the complete data-set whilst the tails to the left show the heterogeneity of the data after removal of the largest structures in half-order-of-magnitude steps ($\approx \log_{10} 3$). A grey arrow indicates the general trend of heterogeneity for the study area. It can be seen that $V_S'$ is strongly scale-dependent for all data-sets.

At the lower end of the scale range ($10^{-4} - 10^{-2}$ m heave) strain is accommodated by veins and shows high heterogeneity ($V_S' \geq 0.5$) indicating localized strain. At the fault-scale ($10^0 - 10^2$ m heave) fault-strain is only weakly heterogeneous ($0.1 <$
Figure 4.9: Heterogeneity of strain-distribution (V_S'). (a) Plot of strain heterogeneity (V_S') with respect to the sample size (n) of each data-set. The three curves labelled 0.995, 0.95 and 0.90 are the critical values for Kuiper’s test (Stephens, 1965) for the probabilities with 90%, 95% and 99.5% confidence that a data-set is significantly different from a uniform distribution. Values below the lines are not significantly heterogeneous whilst values above the curves are significantly heterogeneous. (b) Strain heterogeneity (V_S') versus heave-range covered by each data-set. Filled diamonds represent the complete data-sets, tails to the left of each diamond show the changes in heterogeneity after subsequent removal of the largest structures (heaves) out of the data-set in steps of half-order-of-magnitude (∝ log_{10} 3).
Strain scaling

Bin strains:
The extension accommodated by faults of a certain size-range in the study-area can easily be determined from the fault-line data for the heave-range from 0.1 m to 100 m. Fig. 4.10 shows the cumulative heave of the fault-line with subsequent removal of the largest faults (heaves) from the data-set in steps of half an order of magnitude ($\cong \log_{10} 3$). It can be seen that within the studied section faults with heaves between 10 m and 30 m take up 13.2% of extension. This is more than half of the total extension accommodated by faults (Fig. 4.10). Faults with heaves larger than 30 m are important but are probably somewhat under-represented due to the relatively small section length of less than 2 km. These largest faults in the section take up 6.7% of extension. Faults below the resolution limits of high-quality seismic data (heaves $< 10$ m) accommodate another 5.6% of extension.

Damage zones around faults:
Damage-zone widths in the hanging walls and footwalls of faults range from 9 cm to 10.3 m and lie most commonly in the range of 3 to 6 m. Tensile fractures accommodate extensions of between 0.3% and 15.4% within these damage zones. Neither the number of fractures within, nor the widths of the damage zones seem to scale with the fault-displacements. There is however a correlation between the heave of a fault and the amount of extension accommodated within its damage zone (Fig. 4.11b). Based on this relationship it is possible to estimate the contribution of small-scale structures to the total extension with respect to the heave of the fault they surround. Obviously, the smaller the fault, the larger the relative importance of its associated small-scale structures: For faults with heaves of 0.1 m to 1 m, the cumulative vein-thickness in the associated damage zones add about 10% to the fault-strain. In the heave-range 1 m to 10 m the veins increase the extension by about 3%. For the largest faults ($>10$ m heave) in the studied section veins add
Figure 4.10: Cumulative heave versus distance plots of the fault-line with successive removal of the largest faults (heaves) in half-order-of-magnitude ($\approx \log_{10} 3$) steps. The upper-most curve includes all sampled faults. The right-hand vertical axis shows the bulk extension accommodated by each (sub) data-set. Labels on the graphs indicate the included heave-range.
only about 1% of extension to the fault-heave (Fig. 4.11).

*Total brittle extension:*

Having established the extensions accommodated by faults and their damage-zones, the total extension of the studied section can be estimated with 25.2%. Compared to the fault-strain of 24.7% it can be seen that the small-scale damage accounts for only 0.5% of strain.

### 4.4.3 Evolution of early-stage extension-related damage

Fig. 4.12 compares the observed structures for different stages of extension in the study area. In the stiffer carbonate layers early extension is highly localised in narrow zones (typically 1 to 5 m wide) preserving large proportions of virtually undeformed rocks in between (Fig. 4.12a). Tensile deformation around the tip zones of propagating faults (process-zone) appears to be the dominant bed-scale deformation mechanism in the carbonate beds (Schoepfer et al., 2006). The more ductile mudstone-layers accommodate initial deformation in a more distributed manner. As extension increases, small displacements are accommodated by bending of the carbonate beds and opening (thickening) of veins (Fig. 4.12b) until the fault breaks through to facilitate slip (Fig. 4.12c). Once a network of faults is formed, further increase in strain is accommodated dominantly by slip on the fault system.

### 4.5 Summary and Conclusions

*Fault and vein distribution and damage zones*

Faults are clustered ($C_V^* = 1.65, V'_F = 0.24$) in the studied section and are surrounded by, typically 1 to 5 m wide, damage zones. The widths of these damage zones appear to be largely independent of the displacement on the fault. The damage is preserved as densely spaced veins and pull-apart structures. As these fractures are localised within narrow zones, leaving large zones of virtually undeformed host-rock in between, they show high spatial heterogeneity with $0.37 < V'_F < 0.89$. 

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Figure 4.11: (a) Cumulative extension for half-order-of-magnitude ($\approx \log_{10} 3$) heave-intervals. The values consist of fault-extensions and associated damage-zone vein-strain. Faults accommodate an extension of 24.7% whilst veins account for only 0.5%, adding up to a total brittle extension of 25.2%. (b) Log-log plot of extension accommodated in damage-zones versus fault-heave. Although the data are sparse, it can be seen that the damage-zone extension seems to scale with fault-heave obeying a power-law relationship.
Figure 4.12: Conceptual model of the early-stage evolution of extension-related damage in the interbedded carbonates and mudstones at Kilve. The sketches and outcrop pictures show the evolution of a localised damage zone from low strain, accommodated mainly by opening of tensile fractures (veins) and bending of the carbonate beds (a), to higher extension (b) producing a fault (c).
Strain distribution at different scales

Even though the faults within the studied section are clustered, the extension that is accommodated by these faults, is fairly uniformly distributed ($V_s' = 0.14$). The extension accommodated by small-scale structures (veins and pull-aparts) on the other hand is heterogeneous ($V_s' \geq 0.5$) due to the localisation of these fractures within narrow damage zones. Removing the largest structures from the fault-line data increases the heterogeneity, indicating that the fault-population as a whole accommodates a constant regional strain.

The amount of extension accommodated within a damage-zone is found to scale with the displacement on its associated fault. For faults with heaves of 0.1 m to 1 m, the cumulative vein-thickness in the associated damage zones adds about 10% to the fault-strain. In the heave-range 1 m to 10 m the veins increase the extension by about 3%, and for faults with heaves >10 m veins add about 1% of extension to the fault-heave.

Conceptual model of the early-stage evolution of extension-related damage

Early extension is highly localised in narrow zones preserving large proportions of virtually undeformed host-rock in between. Initial deformation is taken up by distributed deformation within the more ductile mudstone layers, whilst it causes tensile failure in the stiffer carbonate beds. Small vertical displacements are accommodated by bending of the carbonate beds associated with tensile failure and opening of veins. Eventually the fault breaks through and facilitates slip. Once a network of faults is established, further increase in extension is accommodated dominantly by slip on the fault system associated with some thickening of veins in the surrounding damage zones.

Total extension

Brittle structures accommodate a total brittle extension of about 25.2% within the
studied section. About 80% of this extension are taken up by seismically resolvable faults (heaves $\geq 10$ m) whilst the remaining 20% are accommodated by fractures below seismic resolution. Most of this “sub-seismic” extension is taken up by small faults (heaves $< 10$ m) whilst veins only accommodate a total cumulative extension of about 0.5%.
Chapter 5

Maltese Islands Field Study

5.1 Introduction

The Maltese Islands (Malta, Gozo and Comino) in the Mediterranean Sea provide exposures of pelagic carbonates that have been deformed in an extensional system since the Miocene. Large portions of the resulting Horst and Graben system can be studied in detail in excellent exposures along the coast lines of the islands. In addition the geological maps of the Maltese Islands, mapped at 1:25,000 scale, are of exceptional accuracy, permitting the outcrop data to be placed in a regional tectonic framework. These conditions make the Maltese Islands a great outcrop analogue for many (offshore) fractured carbonate platform reservoirs, both in extent and geological setting.

The region is a perfect place to examine how faults and extension are distributed in a carbonate platform. In particular it is possible to quantify displacements on, and extension accommodated by, faults over a wide range of scales. These data can be used to assess the amount of information lost due to technical limitations in seismic surveys and to analyse fault distributions below seismic resolution. Detailed information on higher and lower strain zones within the graben system allow examination of differences in spatial and size distributions of faults.

The aim of this study was to use data from exposures and maps of the Maltese Islands to evaluate interpretation of seismic reflection profiles, treating the geologi-
cal maps as equivalent to data from seismic and the exposure studies as providing information at sub-seismic resolution. The following objectives were defined:

1. Evaluation of the nature and distribution of structures developed due to extension.

2. Determination of the relative contributions of larger and smaller faults (above and below seismic reflection resolution) to the total strain of the exposed graben system.

3. Examination of the spatial organisation of faults with size and throw below and close to seismic reflection resolution.

4. Analysis of systematic differences in fault distributions in higher and lower strain zones within a graben system.

5.2 Geological overview

The Malta Graben System (Fig. 5.1) is one of several troughs within the Strait of Sicily rift system at the northern edge of the African plate. The Straight of Sicily rift system extends the central Mediterranean Pelagian Platform that represents the passive continental margin of the Alpine orogeny. The tectonic style characterising the Strait of Sicily rift system is remarkably symmetrical and has been attributed to “McKenzie-style” pure shear extension with an upper brittle layer overlying a ductile lower layer, producing a symmetrical lithospheric cross-section (Ben-Avraham et al., 2006). The stretched continental crust thins to less than 20 km (Colombi et al., 1973) beneath the 100 km wide Pantelleria Trough located SW of Malta (Reuther and Eisbacher, 1985). The sedimentary cover within the Pelagian platform consists of Meso-Cenozoic carbonates that extend from Sicily across the Mediterranean Sea to Tunisia and Libya in the south. The Malta islands (Malta, Gozo and Comino) are one of few “off-shore” locations within the platform that can be studied on land.
The Malta Platform is located about 100 km south of Sicily. It forms the north-eastern shoulder of the Pantelleria Rift (Reuther and Eisbacher, 1985) and is dissected by the ENE - WSW - trending North Gozo and North Malta Graben (Dart et al., 1993). Both rift systems were active during the Miocene but reached a peak in activity during Plio-Quarternary times (Illies, 1981; Bosence and Pedley, 1982; Finetti, 1984; Jongsma et al., 1985).

The main focus of this study lies on the normal fault system associated to the North Malta Graben that separates the Maltese Islands from each other. The extent of the North Malta Graben is defined by two major faults (South Gozo Fault and Victoria Lines Fault) flanking the graben to the North and South respectively. Both of these graben-bounding faults are exposed on the islands (Fig. 5.1).

5.3 Data

Virtually all of the brittle extension on the Maltese Islands is accommodated by faults with displacements ranging from cm to >100 m. Tensile fractures are very rare and certainly do not significantly contribute to the regional extension. The high accuracy of the 1:25,000 geological map of the Maltese islands (Pedley et al., 1993) allows production of map-scale cross-sections that include all faults with throws ≥10 m. These sections are thus equivalent to high-quality seismic reflection lines. Exceptionally well exposed cliff sections across segments of, and at high angles to, the fault-system permit the collection of high-resolution data below seismic resolution. In total, about 56 km of line-samples from Malta and Gozo have been analysed for this study (Table 5.1). 34 km of which were derived from the geological map (3 cross-sections: Malta A, B and Gozo A) and one 11 km section (Malta C) is based on a detailed section published by Dart et al. (1993) and improved during the field work for the present study. The remaining approximately 11 km consist of cliff-scale sections, sampling both, higher and lower strain zones within the fault system. The locations of all data-sets are shown in Fig. 5.1 and a summary of the physical properties of each data-set is given in Table 5.1.
Figure 5.1: Map of the Maltese Islands showing sample locations, major faults and higher strain zones. The thick dashed line is a section published by Dart et al. (1993) on which section Malta C was based for the present study. The inset shows the Maltese Islands in the regional tectonic context.
Table 5.1: Summary of field observations and positions the 16 line samples. Coordinates refer to WGS, UTM projection and were determined using an EGNOS-enabled hand-held GPS with accuracies of about ± 8 m.

It can be seen from the map (Fig. 5.1) that most of the collected data sample the exposed areas of the North Malta Graben. *Malta A* and *C* record the map-scale fault-strain within the graben. *Malta D* to *G* sample the cliff-scale fault-strain within the same section. *Gozo C* and *D* sample the equivalent cliff-scale fault-strain on the Gozo side of the graben. *Gozo A* records the map-scale fault-strain all the way from the virtually undeformed *Gozo Horst* into the strongly extended northern shoulder of the graben. *Malta H* samples a higher strain zone south of the main graben system and *Malta B* samples the map-scale fault-strain within the *Malta Horst*. In total fault-displacements over close to four orders of throw-magnitude have been sampled in sections ranging from several 10s of metres to >10 km length (Fig. 5.2). In the following sections the data-sets are described in more detail. For this purpose the data are separated by origin (Malta or Gozo) and by type of section (map-scale or cliff-scale).

Given that the NW-SE extension commenced in Miocene times (Dart et al., 1993) and is still ongoing, the top of the Lower Coralline Limestone that pre-dates the extension was used as the reference level for determining fault-displacements (Fig. 5.3) at the map-scale. For the cliff-scale sections a persistent marker-layer - a hardground at the base of the Globigerina Limestone - which pre-dates the extension
was chosen (Fig. 5.3). This marker bed crops out widely on the Maltese Islands and the 12 cliff-scale sections trace it over distances between 24 m and 6 km. The bed is easy to identify and allows fault-displacements to be accurately measured (Fig. 5.4).

5.3.1 Malta

Map-scale sections

*Malta A* and *B* (Fig. 5.5a, b) are the N-S trending cross-sections *A-A’* and *B-B’* from the geological map of Malta (Pedley et al., 1993). These lines have lengths of 16 km and 8.3 km respectively and are considered to resolve all faults with throws ≥10 m. *Malta A* includes large portions of the *North Malta Graben* and records a bulk extension of 2.22% compared to 3.02% within the graben. The spatial distribution of the faults and their accommodated strain appears to be fairly uniform with some localisation of strain onto the *Victoria Lines Fault* (Figs. 5.1, 5.5a). *Malta B* is
Figure 5.3: Stratigraphic correlation between the structural elements of the North Malta Graben System. The Gozo and Malta Horsts are separated by the North Malta Graben, which is bound by the South Gozo Fault to the North and the Victoria Lines Fault to the South. Note the approximate position of the marker bed used in this study. Modified after Dart et al. (1993).

located outside the graben system, in a lower strain zone in the east of the island, and records an extension of only 0.13% accommodated by a single fault with throw $>10$ m. Together these two map-scale sections provide a good regional cross-section of the entire island, showing that the seismically resolvable fault-strain on the island can be divided in the higher-strain North Malta Graben System in the NW and a lower-strain horst zone to the SE.

*Malta C* (Fig. 5.5c) trends sub-parallel to *Malta A*, and thus duplicates the latter partly (Fig. 5.5a). However, as *Malta C* is based on a detailed cross-section published by Dart et al. (1993) and has been refined during the field work for the present study, the resolution of this section is higher than the ones derived from the geological map. Although this section is based on the Dart et al (1993) section, the starting and end point of the original section were connected by a straight line for this study, rather than using 4 segments with different trends as done by Dart et al. (1993) (Fig. 5.1). *Malta C* has a length of 11 km and includes all faults with throws $\geq 5$ m, recording an extension of 2.04%. This is somewhat lower than the
Figure 5.4: Outcrop pictures from the study area. (a) and (c) to (e) show normal faults displaying the marker bed (black arrows) with throws from cm (d) to 15 m (e), all photographs taken along line Malta D’. (b) Is a view along the NW coast of Malta towards SE. The major fault which causes a step in the coast line is the Quammieh Fault with a throw of ca. 120 m. For a scale note the two people in the circle.
Figure 5.5: Cumulative heave (solid lines) and fault-number (dashed lines) plotted against corrected distance (traverse length) for lines: *Malta A* (a), *Malta B* (b), *Malta C* (c), *Malta D* (d), *Malta D’* (e), *Malta D”* (f). Left vertical axes show the cumulative heave, right vertical axes show the cumulative number of encountered structures along each line.
strain recorded within the graben along Malta A. The difference is probably due to Malta C not covering the entire extent of the footwall damage zone of the Victoria Lines Fault (Fig. 5.1) which splays into several segments towards the West.

Cliff-scale sections

Malta D (Fig. 5.5d) has a length of almost 6 km and resolves all faults with throws ≥1 m. It starts at the Victoria Lines Fault in the North, crosses a 2.1 km wide damage-zone with a resolved extension of 1.17%, and traces the marker-bed over another 3.9 km within a low-strain zone (0.05% extension). The great width of this “footwall-damage zone” is due to the Victoria Lines Fault breaking up into several footwall splays in this region, distributing some of its displacement over a more than 2 km wide zone (Fig. 5.1). Within this damage-zone the outcrop-quality is even higher than throughout the rest of the section, allowing for two high-resolution sub-data-sets. One of these, Malta D’ (Fig. 5.5e) starts at the same point as Malta D but with a resolution of 0.5 m over a distance of 1600 m, recording a fault-extension of 1.52%. Within Malta D’, a 961m long sub-section (Malta D”, Fig. 5.5f) provides even higher resolution. The section follows a wide ledge on top of the Lower Coralline Limestone which forms a balcony above the sea and is displaced by the faults which are the subject of this study. It allows all faults with a resolution ≥0.1 m to be recorded and the outcrop-quality is magnificent (Fig. 5.4a,c,d,e).

Malta E and F (Fig. 5.6a,b) are located in the NW of Malta. Malta E has a length of 903 m at a resolution of 0.5 m and samples mainly the footwall of the major Quammieh Fault (Fig. 5.4b) which it crosses about 200 m before the end of the section. Most of the 2.18% extension recorded for this section is accommodated by the Quammieh Fault. The resolved extension without this major fault is only 0.14%. The trend of the line however is highly oblique to the principal stretching direction. Malta F is a short (195m), high-resolution (≥ 0.2 m throw) section just north of Malta E which crosses a small (66m wide) higher strain zone, recording a total extension of 2.38%.
Figure 5.6: Cumulative heave (solid lines) and fault-number (dashed lines) plotted against corrected distance (traverse length) for lines: Malta E (a), Malta F (b), Malta G (c), Malta H (d). Left vertical axes show the cumulative heave, right vertical axes show the cumulative number of encountered structures along each line. Sampling direction is from N to S in all lines.

Malta G (Fig. 5.6c) has a length of 416 m at a resolution of 0.1 m. It is located in the NE of Malta, just W of St. Paul’s Island. This section samples the hanging-wall of the major St. Paul’s Fault and ends at this fault. The extension accommodated along this section with the major fault is 9.82%, but only 0.83% outside the main fault zone.

Malta H (Fig. 5.6d) has a length of 2411 m and a resolution of 1 m. It is located in the SW of Malta and is the southern extension of Malta D with a low-strain gap of about 2 km in between the two sections. This section crosses a higher-strain zone (3.28% extension) within the Malta Horst with two major faults accommodating most of the strain.
5.3.2 Gozo

Map-scale section

Gozo A (Fig. 5.7a) is the NW-SE trending cross-section A-A’ from the geological map of Gozo (Pedley et al., 1993). It has a length of about 9.4 km and a resolution of 10 m. Most of this section is located on the Gozo Horst and only the last few kilometres cross the northern margin of the North Malta Graben. The first 7 km show no fault-strain (≥10 m throw) whilst the final 2 km record an extension of 3.24%, giving an average extension of 0.69% for the entire section.

Cliff-scale sections

Gozo B (Fig. 5.7b) is located in the SE of Gozo, has a length of 528 m and a resolution of 1 m. It samples the footwall of a major fault within the North Malta Graben and ends at that fault. The extension accommodated in this section with and without the major fault is 13.56% and 3.00% respectively. Gozo B’ (Fig. 5.7c) is a high resolution (≥0.03 m) sub-sample of Gozo B with a length of 24 m, located in the immediate footwall of the major fault close to the end of the section.

Gozo C (Fig. 5.7d) is located in the S of Gozo, just W of the harbour (Mgarr). It has a length of 205 m and shows a slight increase in strain from N to S. The total resolved extension, at a resolution of 1 m, is 2.09%. Gozo C’ (Fig. 5.7e) is a 68 m long sub-sample of Gozo C with a resolution of 0.1 m.

Gozo D (Fig. 5.7f) is located at Xlendi Bay in SW Gozo. It has a length of 288 m and a resolution of 0.03 m. The extension of 0.48% is fairly homogenously distributed across the section and is entirely accommodated by faults with throws ≤0.65 m.

5.4 Discussion of results

Integrating the different line-samples facilitates spatial and size-distribution analysis of the Malta Fault System covering faults with a range of throws of close to 4 orders of magnitude (Fig. 5.2). The map-scale cross-sections are used to determine the
Figure 5.7: Cumulative heave (solid lines) and fault-number (dashed lines) plotted against corrected distance (traverse length) for lines: *Gozo A* (a), *Gozo B* (b) *Gozo B’* (c) *Gozo C* (d), *Gozo C’* (e) *Gozo D* (f). Left vertical axes show the cumulative heave, right vertical axes show the cumulative number of encountered structures along each line. Sampling direction is from N to S in all lines.
distribution of, and the strain accommodated by, seismically resolvable faults. The cliff-scale sections are used to constrain distribution and strain accommodated by sub-seismic scale faults, both within higher and lower-strain zones.

For the purpose of the statistical analysis the cliff-scale data-sets were grouped in three categories: i) higher strain zones, ii) lower-strain zones and iii) mixed data. Higher strain zones are defined as areas in which the observed strain is significantly (typically an order of magnitude) higher than in the surrounding areas. In Fig. 5.8, higher strain zones are shaded in dark grey and separated by light-grey-shaded lower-strain zones. Mixed data are sections that include both higher strain zones and sections of lower strain. This simple separation permits comparison of trends in fault distribution and scaling that may differ in higher and lower strain zones.

5.4.1 Spatial distribution of faults and strain

Fault-spacing

Fault spacing on the Maltese Islands has been analysed using two independent methods. The first one is the determination of the Coefficient of Variation of fracture spacing ($C_V^*$). The second one determines the heterogeneity of fracture-distribution by comparing the cumulative frequency plot for each data-set with the uniform cumulative distribution by applying Kuiper’s test. Both methods are described in detail in Chapter 2 and the results are discussed below. Due to the localisation of extension on the Maltese Islands in higher and lower strain zones (Fig. 5.8), with the latter showing virtually no brittle deformation, the following heterogeneity analysis focuses on the higher-strain zones.

Coefficient of Variation:

The values for $C_V^*$ are listed in Table 5.2 and plotted against the throw-range covered in each data-set in Fig. 5.9a. It is seen that the map-scale data from within the North Malta Graben (Malta A graben and Malta C), and the cliff-scale damage-zones have spacings with Coefficients of Variation around $C_V^* = 1$, suggesting a random fault distribution in the higher strain zones. A vague trend within these
Figure 5.8: Map of the Maltese Islands showing the distribution of higher strain zones and lower strain zones. The NW-SE sections across the Maltese Islands which are used for the estimation of the total extension (see section 5.4) are shown as bold black lines (North Malta Graben) and dashed grey lines (Gozo and Malta Horsts). Sampling direction is from N to S in all lines.
higher-strain data (thick lines in Fig. 5.9a) suggests a somewhat more regular fault-spacing at the map-scale (larger faults) than at the cliff-scale (smaller faults) with $C_V^* \leq 1$ for the larger faults and $C_V^* \geq 1$ for the smaller faults. This could be due to a weak clustering of smaller faults in higher strain zones surrounding larger faults. As discussed above, the lower-strain data are more difficult to interpret as the sample size - in particular the number of larger faults - in these lines is very low (close to zero). However, faults in lower-strain zones, if there are any, appear to be weakly clustered at the cliff-scale and clustered at the map-scale (e.g. Gozo D, Malta B).

Kuiper’s Test:

As discussed in Chapter 2, Kuiper’s method provides a test that allows determining whether a data-set is significantly different from a uniform distribution, based on the sample size $n$ and the heterogeneity measure $V'$. Fig. 5.9b (crosses) and Table 5.2
Figure 5.9: (a) Coefficient of Variation ($C_V^*$) of fault-spacing versus throw-range for all data-sets. (b) Heterogeneity of fault spacing ($V_{F'}$) and of strain distribution ($V_{S'}$) versus sample-size of all data-sets. The three curves in the diagram represent the critical values for rejection of the null hypothesis of the uniform distribution at confidence levels of 90%, 95% and 99.5% respectively.
display the results of this test for the fault-spacing ($V_F'$). It can be seen that less than half of the data-sets show a significantly heterogeneous fault-spacing at the 0.1 level. In particular, damage-zones and map-scale sections display a fault distribution that is not significantly different from a uniform distribution. This suggests that map-scale faults and all faults within higher strain zones are randomly distributed. Fig. 5.10a shows the same results ($V_F'$) against the throw-range included in each data-set. This plot highlights the very low heterogeneities ($0.19 < V_F' \leq 0.27$) of map-scale fault-spacing within the graben (e.g. Lines Malta C and Malta A graben) and the low to moderate heterogeneities ($0.28 < V_F' < 0.36$) within cliff-scale damage-zones (e.g. Malta D”, Malta D’, Gozo C, Gozo B).

Spatial heterogeneity of extensional strain

The statistical significance of the derived strain-heterogeneities ($V_S'$) are listed in Table 5.2. The heterogeneities are plotted against sample size in Fig. 5.9b (squares), together with three curves representing the critical values for rejection of the null hypothesis of the uniform distribution at confidence levels of 90%, 95% and 99.5% respectively (Stephens, 1965). The majority of the data show significantly heterogeneous distributions of extension at the 0.1 level. However, extension on all of the map-scale fault-lines within the graben is not significantly different from uniform distributions, and thus map-scale fault-strain appears to be uniformly distributed across the North Malta Graben.

Fig. 5.10b shows the results of Kuiper’s test applied to the cumulative heave distributions ($V_S'$) against the throw-range covered in each data-set. It can be seen that strain is moderately heterogeneously distributed ($0.5 > V_S' > 0.3$) at the cliff-scale (lower end of the scale-range), both in higher and lower-strain zones (e.g. Malta D” and Gozo D). At the map-scale (higher end of the scale-range) there is an apparent difference between horsts (e.g. Malta B) showing high heterogeneity ($V_S' close to 1$) and graben (e.g. Malta A and C) with $V_S' < 0.5$. However, because the lower-strain zones show no or only sparse map-scale faulting, the resulting high heterogeneities
Figure 5.10: Heterogeneity of fault spacing ($V_F'$) (a) and of strain distribution ($V_S'$) (b) for all data-sets, plotted against the throw-range covered by each data set.
of these areas (e.g. Malta B) may be a result of the small sample size rather than of high localisation (Fig. 5.9b). Mixed data, sampling both, higher- and lower-strain zones, show heterogeneities that plot in between the higher- and lower-strain data ($0.9 > V'S' > 0.4$). As the sample-size of these data-sets is large enough, to detect significant deviations from uniform (or random) distributions, it appears that the heterogeneity of extension is indeed somewhat higher within the horsts than within the graben.

5.4.2 Fault and strain scaling

Fault frequency in the North Malta Graben System

Separating the data-sets as discussed previously in higher-strain and lower-strain data (Fig. 5.8) allows the examination of systematic differences in fault scaling between higher and lower strain zones. Fault frequencies for half-order-of-magnitude throw-intervals were determined for different data-sets sampling higher and lower strain zones (Table 5.3). Only the most representative sections, in terms of position, length and resolution, were considered for this analysis. The position of a sample-line (i.e. within or outside a higher strain zone, within or outside the graben) determines which deformation type it samples. The length of a section is important because short lines (say <100 m length) yield samples which are strongly dependent on short-wavelength variations in small-scale damage. Long cliff-scale sections (say >1 km) on the other hand may sample both higher and lower strain zones and thus yield samples which do not represent either type of deformation. Resolution limits of each data-set have to be considered to avoid sampling artefacts (i.e. under- or over-sampling of structures of a certain size).

The 11 data-sets chosen for the analysis are thought to best represent the differences in deformation between horsts and graben at the map-scale and between higher-strain and lower-strain zones at the cliff-scale. Table 5.3 lists the observed frequencies per throw-interval of faults in higher and lower strain zones. The mean fault-frequencies for each interval are plotted as cumulative-frequencies against throw in
Table 5.3: Representative fault-frequencies per throw-interval for higher and lower strain zones. A: Probably under-sampled due to resolution limits; B: Probably under-sampled due to insufficient traverse length.

Fig. 5.11. The data for both higher and lower-strain zones appear to obey power-law distributions over a throw-range of 2 to 3 orders of magnitude. However, the data do not share a single scaling-relationship. The lower-strain areas show a steeper slope ($D = -1.26$) than the higher-strain-zones ($D = -0.84$) with a division at a scaling exponent of $D \approx -1$. This implies that at the limit of seismic resolution (throw $\approx 10$ m) there is an order of magnitude difference in fault frequency between higher and lower strain zones. The different scaling-exponents indicate that in background-strain zones most of the extension is accommodated by small faults, whilst in higher strain zones most strain is taken up by the largest faults.

Similar differences in fracture-scaling between higher and lower strain zones have been found in vein-populations by Roberts et al. (1998, 1999). This has been explained as due to increased localization and fracture-linkage as the overall strain increases. Based on field observations (e.g., Fossen and Rornes, 1996; Wilson, 2001; Moriya et al., 2005) and numerical models (e.g., Spyropoulus et al., 1999; Hardacre and Cowie, 2003) similar trends have been suggested for fault populations: Higher-strain regions may show a kink in a size-frequency log-log plot with a smaller slope (exponent) for smaller (earlier) faults and a higher slope for larger (later) faults.
Figure 5.11: Cumulative fault-frequency (number/m) plotted against fault-throw in half-order-of-magnitude bins in log-log space. Note the different slopes of the power-law trends (dashed lines) for higher and lower strain zones.
Figure 5.12: Table representation of the distribution of extension within the exposed areas of the North Malta Graben and the Gozo and Malta Horsts. Most of the total extension is accommodated within higher strain zones, both, within the graben and the horsts. The North Malta Graben consists of 75% higher strain zones, separated by 25% of background strain zones. The horsts north and south of the graben consist of about 38% of higher strain zones embedded in 62% of background strain. Faults with throws >10 m accommodate 2 to 3% of extension and smaller faults take up another 1 to 2.5% within the higher strain zones.

**Total brittle extension**

From the most representative data-sets that were already used for the frequency analysis in the previous section (Table 5.3) it is straight forward to determine the mean extensions accommodated by faults in higher and lower strain zones. This is done by separately calculating the cumulative fault-heave along a sample line for faults with throws ≥10 m and <10 m for each data-set. Grouping the derived values in higher and lower strain zones it can be seen that faults below seismic resolution (throw <10 m) take up between 1% and 2.5% extension within higher strain zones whilst they accommodate less than 0.5% extension in lower strain zones (Fig. 5.12). Larger faults (throw ≥10 m) account for an extension of 2% to 3% in higher strain zones but close to 0% in lower strain zones (Fig. 5.12).

Based on a NW-SE section across Gozo and Malta (Fig. 5.8) it can be estimated
that the exposed sections of the *North Malta Graben* consist of 75% of higher strain zones (dark shaded) separated by 25% of background strain zones (light shaded). The exposed areas in the horsts north and south of the graben consist of about 38% of higher strain zones embedded in 62% of background strain. Using these proportions of higher and lower strain zones and the derived strain values the mean NW-SE extension within the exposed section of the *North Malta Graben* can be estimated with 3.3%. Outside the graben the mean extension is about 1.8% (Fig. 5.12).

Thus, faults with throws <10 m take up most of the extension in lower-strain zones, whilst in higher strain zones it is the largest faults (≥10 m) that accommodate the largest portion of the strain. These results agree well with the different power-law scaling relationships found for faults in higher and lower strain zones (Fig. 5.11) as discussed in the previous section.

### 5.5 Conclusions

The studied, approximately ENE trending fault system associated with the *North Malta Graben System*, largely covers the Maltese Islands and can be divided in three structural domains, which are from N to S, the *Gozo Horst*, the *North Malta Graben* and the *Malta Horst*. The extent of the *North Malta Graben* is defined by two major faults (*South Gozo Fault* and *Victoria Lines Fault*), flanking the graben to the North and South and cropping out at Gozo and Malta, respectively.

Virtually all the brittle extension on the Maltese Islands is accommodated by faults with displacements ranging from cm to >100 m. The extension related damage is not homogeneously distributed across the islands but is localised in few >km-wide higher strain zones that are separated by virtually undeformed lower-strain regions. These higher strain zones are most prominent within the graben but also occur within the horsts.

Faults in higher strain zones generally are randomly distributed with a trend from somewhat greater heterogeneity at the cliff-scale (faults below seismic resolution)
to more uniformly spaced faults at the map-scale (faults within seismic resolution). Where there are larger faults present within horsts, these show clustered distributions. Spatial distribution of extension in (cliff-scale) higher strain zones shows moderate heterogeneity. At the map-scale extension is randomly distributed within the graben whilst it is heterogeneously distributed (localised) within the horsts. Higher-strain zones make up about 75% of the graben and exactly half of this of the horsts. Faults, both in higher and lower strain zones, display power-law throw-distributions over three orders of magnitude, but with different scaling exponents of $D = -0.84$ and $D = -1.26$, respectively. This implies that at the limit of seismic resolution (throw $\approx 10$ m) there is an order of magnitude of difference in fault frequency between higher and lower strain zones with few faults reaching a sufficient displacement to be imaged in the lower strain zones. Yet, at the cm-scale resolution similar frequencies are observed. The division of scaling exponents around $D = -1$ indicates that larger faults are dominant in higher strain zones whilst smaller faults are more important in lower strain zones.

Higher strain zones typically accommodate a total strain of 4.25%; about 2.5% taken up by larger ($\geq 10$ m throw) and about 1.75% by smaller ($< 10$ m throw) faults. Lower strain zones in between higher strain zones typically accommodate only 0.35% extension, most of which (0.25%) is taken up by smaller ($< 10$ m throw) faults. From these values it can be estimated that the fault-system as a whole accommodates a bulk NNW-SSE extension of about 2.4% across the islands. Major faults and higher strain zones account for about 95% of this extension and the North Malta Graben takes up about twice the extension of the Gozo and Malta Horsts.

Based on these results it can be concluded that a high-quality seismic reflection survey (throw-resolution $\geq 10$ m) across the North Malta Graben system would miss between 40% and 45% of the total extension. Number and size of sub-seismic faults (throw $< 10$ m) can be predicted by power-law scaling relationships from seismically resolvable faults only if care is taken to separate higher and lower strain zones for determining the scaling-exponents.
Chapter 6

Strain Tensor

6.1 Introduction

So far only the horizontal (or bedding-parallel) component of extension has been considered (Chapters 2 to 5). These extension estimates were based on the sum of recorded heaves along each line after correcting them for oblique (non-perpendicular) intersection between sample line and fault trend. Tensile fractures (veins) commonly develop parallel to $\sigma_1$ and perpendicular to $\sigma_3$ whilst normal faults usually trend approximately perpendicular to the maximum and minimum principal stretching axes (e.g. Price, 1981). Thus, under pure-shear conditions, sample-lines orientated perpendicular to the mean fault (and vein) trend can be assumed to be approximately parallel to the maximum stretching axis and thus should record the maximum extension. This assumption holds if

1. the polarity of faults in the region under investigation is statistically uniformly distributed (i.e. approximate same frequency and displacements of N - dipping and S - dipping faults in a E-W trending set of faults).

2. the strike-variation of the sampled faults is low.

3. the faults show dominantly dip-slip displacements.
Dominant dip of faults in one direction causes a vertical, rotational component of the strain (simple-shear component) implying that one-dimensional horizontal estimates will underestimate the maximum extension. Strike-variation and strike-slip components of fault displacement cause an “out-of-plane” component of strain and can rotate the principal stretching axes horizontally, which again may give rise to an underestimation of the maximum extension.

Sets of natural faults are rarely orientated perfectly parallel and can show considerable variation in orientation (i.e. strike and dip) and displacement direction. This implies that the derived one-dimensional extension values do not necessarily represent the maximum principal extension. In the present chapter the validity of the above (non rotational, plane strain) assumptions for the data-sets used in the thesis will be tested and deviations from pure-shear, plane strain conditions examined. This is done by extending the one-dimensional analysis to two and three dimensions applying a method based on position gradient tensors.

Under pure-shear, plane strain conditions the area of a cross-section parallel to the maximum and minimum stretching axes is preserved and thus, the maximum stretching in $e_1$ direction $(1 + e_1)$ is equal to the inverse of the minimum stretching in $e_3$ direction $1/(1 + e_3)$ (Fig. 6.1a).

$$(1 + e_1)(1 + e_3) = 1 \quad (6.1)$$

The directions of greatest compression and extension are constant and independent of the magnitude of the strain. The orientations of maximum and minimum stretching axes remain constant. All other lines rotate. In this case one-dimensional analysis yields the maximum principal strain.

Under simple-shear, plane strain conditions the area of a cross-section also remains constant, but only the direction parallel to the shear plane remains in its initial orientation whilst all other directions (including the principal stretching axes) rotate relative to it (Fig. 6.1b).

As discussed in Chapter 1, the term strain should only be used for continuous de-
Shear strain: 0
Shear strain: 0.5
Shear strain: 1.0
Shear strain: 1.5

Figure 6.1: Pure shear (a) and simple shear (b) deformation of a "unit-square" shown at shear strains of 0.5, 1.0 and 1.5. Note that the strain ellipses have the same shape (size) at the same shear-strain, but they show different orientations in pure shear and simple shear.

formation, whilst the term fault strain is more appropriate for brittle deformation. However, where discrete deformation (slip on faults and opening of tensile fractures) is small, compared to the extent of the area of investigation, and the deformation is distributed (as oppose to localised) across the region, the cumulative deformation can be treated as continuous (Fig. 6.2). In other words, the distinction between continuous and discontinuous deformation is a matter of scale (Gauthier and Angelier, 1985; Jamison, 1989).

6.2 Strain in two dimensions

Many data-sets (e.g. isolated 2-D seismic reflection profiles) are essentially two-dimensional. They sample vertical information (depths of reflectors) along horizontal straight lines. This type of data does not sample any “out-of-plane-of-observation” information, such as the trend of faults. For this reason the fault-heaves, determined from recorded fault-throws and dips, are not real but apparent heaves because the fault-dips cannot be corrected for oblique intersection of faults. However, as discussed above, it is possible to analyse the data in two dimensions to examine rotational components and thus deviation from pure-shear deformation.
Figure 6.2: Examples of distributed small fault displacements that convert a circle into a “broken ellipse”. (a) Parallel faults represent simple shear, (b) sets of conjugate faults represent pure shear conditions. Note the rotation of the (strain) ellipse in (a) by an angle $\theta$. 
Figure 6.3: Sketch of the components of the displacement tensor. \( \mathbf{n} \) and \( \mathbf{u} \) are unit-vectors orientated normal and parallel to the displacement direction respectively. \( \mathbf{n} \) is directed downwards (into the footwall of a fault) whilst \( \mathbf{u} \) is directed upward. This sign convention assures that resulting strains are positive for extension and negative for compression which is the standard convention in tectonic context.

### 6.2.1 Assumptions

- All sampled faults trend approximately perpendicular to the analysed section. 
  (This can often be verified from the known regional trend of large faults).
- All faults show down-dip slip.
- All faults are planar (i.e. fault dip does not change with depth).
- Fault-displacements are small and distributed along the section.

### 6.2.2 2D - Tensor

The displacement on a fault can be expressed by two unit-vectors \( \mathbf{n} \) and \( \mathbf{u} \) which are orientated perpendicular to the fault-plane and parallel to the displacement of the fault respectively (Fig. 6.3).

\[
D_{ij} = s \begin{pmatrix} n_1 u_1 & n_1 u_3 \\ n_3 u_1 & n_3 u_3 \end{pmatrix} = s \begin{pmatrix} -\sin^2 \theta & \sin \theta \cos \theta \\ -\sin \theta \cos \theta & \cos^2 \theta \end{pmatrix}
\]  

(6.2)

\( D_{ij} \) is the displacement tensor; \( s \) is the displacement on the fault and \( \theta \) is the dip of
Following Peacock and Sanderson (1993), the probability of a sample-line intersecting a fault is dependent on the orientation of the fault (i.e. faults parallel to the sample line have probability = 0 whilst faults perpendicular to the sample-line have probability = 1). In the two-dimensional case this sampling issue is of less relevance because all faults are assumed to trend perpendicular to the sample line. Thus, the correction factor only needs to take the fault-dip into account and can be expressed as a weighting factor \( W_t = \frac{1}{\sin \theta} \) (Terzaghi, 1965). The weighting factor can be directly applied to the displacement tensor of each fault before summing the components of all \( D_{ij} \) to give the cumulative displacement tensor describing the cumulative fault displacements along the sampled section.

For small strains (i.e. sum of displacements \( \ll \) length of sample line) the displacement gradient tensor \( (D_{ij}) \) is related to the Lagrangian (forward) strain matrix \( (e_{ij}) \)

\[
e_{ij} = \frac{D_{ij} + D_{ji}}{2} \tag{6.3}
\]

\( e_{ij} \) is a symmetrical matrix (i.e. the off-diagonal components are equal). Thus, it is derived by replacing the off-diagonal components of the cumulative tensor by their mean value and by normalizing each component by the undeformed length of the sample line:

\[
e_{ij} = \begin{pmatrix}
\frac{a_{11}}{L_0} & \frac{a_{13} + a_{31}}{2L_0} \\
\frac{a_{31} + a_{13}}{2L_0} & \frac{a_{33}}{L_0}
\end{pmatrix} \tag{6.4}
\]

\( a_{ij} \) are the components of the cumulative strain tensor \( D_{ij} \) and \( L_0 \) is the initial length of the sample line, i.e. the length before deformation: \( L_0 = L - \sum(heaves) \).

The eigenvalues and eigenvectors of the strain matrix give magnitude \( (e_1 \text{ and } e_3) \) and orientation \( (\theta) \) of the principal strains.

These calculations can easily be carried out in a spread-sheet type procedure for data collected along sample lines as shown in the following section.
6.2.3 Simple tests

To illustrate the methodology, the 2D-tensor was applied to two theoretical examples (Fig. 6.4). Fig. 6.4 shows sets of six normal faults, arranged as conjugate sets in the first case (Fig. 6.4a), and with constant polarity in the second case (Fig. 6.4a). Both examples show the same number of faults in the same position and with the same dip values. The difference is that in Fig. 6.4a the faults are arranged in conjugate sets resulting in a pure shear geometry, whereas in Fig. 6.4b the faults all show the same dip direction, resulting in a rotational strain that is approximately simple shear. In both cases the sum of heaves gives an extension of 4% extension in the horizontal direction. Table 6.1 shows the spreadsheet used for the tensor calculations for the two examples.

It can be seen that in the first case (Fig. 6.4a) \( \theta = 0 \), which means that the direction of maximum stretch is orientated parallel to the sample-line (horizontally) and thus that the sampled deformation conforms to a pure-shear geometry. For this reason the maximum extension can be expressed as \( e_1 = \frac{L - L_0}{L_0} = \frac{(1000 - 961.2)}{1000} = 0.040 \), and is the same as the extension obtained by summing the heaves (i.e. a one-dimensional analysis).

In the second example (Fig. 6.4b) \( \theta = 15.61^\circ \), i.e. the faults produce a maximum extension that is not horizontal, but at about \( 45^\circ \) to the mean fault dip, as expected for an incremental (small) simple shear strain. The maximum principal strain in this case is \( e_1 = 0.047 \), which is larger than that obtained by summing the heaves and in the pure shear case (Table 6.1).

These simple examples show the potential error that can arise from significant simple-shear components of strain. The horizontal extension is the same in both examples, but where there is a rotational strain or vorticity, the one-dimensional strain analysis underestimates the maximum extension. In the example (Fig. 6.4b), the “extreme case” of all faults dipping in the same direction, representing simple shear, causes an error of about 18% of the one-dimensional compared to the two-dimensional estimate of maximum extension.
a) "Pure shear test"

b) "Simple shear test"

Figure 6.4: Simple tests for the two-dimensional strain tensor. 6 faults with throws between 5 and 20 metres and dips between 50° and 70°, displace a marker layer (green). In (a) the faults are arranged as 3 conjugate sets with opposing dip-directions, and thus represent a “pure-shear geometry”. In (b) the conjugate faults are arranged parallel so that all fault dip in the same direction. This represents a “simple-shear geometry”.

Table 6.1: 2D-tensor spreadsheet for the example shown in Fig. 6.4. In both cases six normal faults displace a marker layer by a total heave of 38.76 m. In case (a) the maximum stretch (e₁) is orientated horizontally (theta = 0°) due to the conjugate orientation of the faults. In case (b) theta = 15.61° and thus the strain-ellipse is rotated by this angle with respect to the horizontal marker layer. Grey fields: Input data; yellow field: Cumulative strain tensor, orange field: cumulative strain matrix, box: magnitude and orientation of resulting principal strain.
6.2.4 Application to real data

The strains recorded along map-scale and cliff-scale sample lines from the three study areas (Chapters 3 to 5) were re-examined in a similar way to the examples above. This was done to test the validity of the pure-shear assumption and to estimate the errors due to simple shear strain-components in the study areas. Some results of this two-dimensional strain analysis are compared with one-dimensional strain estimates in Table 6.2. Two examples are discussed in more detail due to the particular significance of 2D strain estimates in these cases.

**Chirp line Weymouth Bay**

The data derived from a high resolution seismic reflection line (Hunsdale et al., 1998), at Weymouth Bay in Southern England (Fig. 3.1), were presented in Chapter 3 (e.g. Fig. 3.4). Because it is an isolated seismic line (i.e. no parallel sections that could be used to infer fault trends), this section represents a “real two-dimensional data set”. The orientation of the section was chosen perpendicular to the regional fault-trend mapped by Donovan and Stride (1961). The mapped faults (Fig. 3.1) trend approximately North-South and are distributed with only slight clustering (Fig. 3.6a) along the entire section.

Two-dimensional analysis of the data yields almost identical values for the maximum extension ($e_1 = 6.51\%$) as the one-dimensional estimate ($e_1 = 6.52\%$) (Table 6.2). This is due to East and West dipping faults occurring in nearly equal proportions, giving the regional deformation an approximate pure-shear geometry ($\theta=1.4^\circ$). In this example one-dimensional, horizontal strain estimates give the correct maximum extension.

**Cliff-section Kilve-Lilstock**

The approximately 2 km long cliff-section between Kilve and Lilstock at the southern bank of the Bristol Channel (Fig. 4.1) shows three characteristics which differ significantly from all other examined sections throughout the thesis. These are
Table 6.2: Summary of 2D strain analysis. It can be seen, that the error of one-dimensional strain estimates compared to the 2D estimates is very small (< 1%) for \( \theta < 5^\circ \) and remains below 5% for \( \theta \) up to about 10°.

1. a higher extension, accommodated by faults, recorded along the line (about 24% in 1D estimate),

2. a dominance of N-dipping faults compared to conjugate S-dipping ones, and

3. a stratigraphic separation between start and end of the section that does not represent the cumulative fault-throws along the section.

The first observation implies that the "small strain assumption" (sum of displacements \( \ll \) length of section) that is used in the presented tensor-method may not be valid for this line. This will be discussed in more detail at the end of the section.

The second observation suggests that the deformation may not represent pure-shear conditions but may involve a significant simple-shear component.

2D analysis of the fault-displacements along the section yields strains that are about 8% higher than 1D estimates (Table 6.2) which is due to a simple shear component of \( \theta = -9.3^\circ \).

However, considering a simple shear geometry (Fig. 6.5a) or a combination of simple shear and pure-shear (Fig. 6.5b) deformation, leads to the third observation that needs to be discussed in some detail. Summing the recorded fault-throws along the section in accordance to their polarity (with or against sampling-direction), would predict (based on a general shear model) marker-beds, which are observed at sea-level at the start of the section, to be more than 200 m below surface at the end of the section (Fig. 6.5d, circles). The observed situation however is quite different; the marker-beds are seen almost 50 m above sea-level at the end of the section (Fig. 6.5d, squares).
Figure 6.5: (a) to (c): Sketches of brittle deformation of a block with rectangular cross section by a set of 6 faults. (a) All faults orientated parallel producing a "simple-shear geometry". (b) 4 faults orientated parallel, 2 faults orientated with reverse dip-direction forming conjugate sets and thus a combination of simple and pure-shear geometry (general shear). (c) All faults orientated parallel, domino-style block rotation. (d) Position of a marker-bed along the cliff-section between Kilve and Lilstock. The bed is observed at sea-level at the start of the section and about 50 m above sea level at the end (black squares). Most faults down-throw to the N. A general shear model as shown in (b) would predict the position of the marker-bed to be about 200 m below surface at the end of the section (circles). A domino-style model (crosses) seems to agree better with the observed geometry.
This indicates that extension is not only accommodated by fault-displacements but also by associated domino-style block-rotation (Fig. 6.5c). This "rigid-body rotation" cannot directly be resolved by the presented tensor method. Considering the geometry of a domino model (Fig. 6.5c), it can be seen that the horizontal extension in this model \( e_{domino} \) is related to the horizontal extensions of simple, pure and general shear models \( e_{generalshear} \) (simple and pure shear) by the mean bed-dip because the faults rotate by the same angle as the bedding.

\[
e_{domino} = \frac{e_{generalshear}}{\cos(bed - dip)}
\] (6.5)

Observed bed-dips along the section vary between 0° and 25° with a mean of about 9°. Using the simplified model in Fig. 6.5c and the mean dip of bedding, a rough estimate on the magnitude of strain, accommodated by block rotation, can be made. A rigid body rotation of about 9° accommodates an extra about 1.2% of horizontal extension \( (1 - \cos(9°)) \) compared to the non-rotational pure-shear model.

### 6.2.5 Results and Discussion

The presented examples show that in regions where the deformation deviates significantly from pure-shear geometry, one-dimensional estimates may significantly under-estimate the maximum extension. Typical sections from the three study areas show that the simple shear component, and thus the deviation from pure-shear geometry, is generally low (theta < 10°). The error made by using one-dimensional, compared to two-dimensional, estimates is \( \leq5\% \) in all but one section.

A larger error (8%) was found for one-dimensional strain estimates from the Kilve-Lilstock section. This higher strain sample (about 26% extension) shows a significant simple-shear component (theta = -9.3°) and a domino-style rigid body rotation of about 9° as observed from bedding dips.

Domino-style rigid body rotation is not accounted for in the presented tensor method.
It is however indirectly represented in both, one-dimensional and two-dimensional extension estimates due to the faults rotating with the blocks they bound. This reduces the initial fault-dip which has two effects:

1. It increases the observed fault-heave and thus the one-dimensional estimate of extension, and

2. It counteracts the simple-shear related rotation of the strain-ellipsoid, which means that the simple-shear component is effectively reduced.

As discussed previously, the presented tensor method is based on the "small-strain theory" (sum of displacements $\ll$ length of section) and it was stated that it should only be applied for sections with low (<10%) strain. The Kilve-Lilstock section shows much higher extension (26%), which means that the derived extension estimates have to be interpreted with care. However, to allow for cases with higher strains, the small-strain theory was slightly modified for the presented 2D tensor method by taking the horizontal change in length ($\Sigma$(heaves)) into account, rather than assuming that the change in length of the sampled section is insignificant.

### 6.3 Strain in three dimensions

#### 6.3.1 Expanding 2D to 3D

Orientation measurements on faults along a section provide three-dimensional information on the regional extension. The methods adopted here follow those of Peacock and Sanderson (1993) who modified earlier applications of displacement gradient tensors to analyse field data (Jamison, 1989; Wojtal, 1989; Marrett and Allmendinger, 1990) and to summation of seismic moments (Molnar, 1983; Marrett and Allmendinger, 1990).

Provided the following conditions are fulfilled, the cumulative brittle strain can be represented by a strain ellipsoid (Jamison, 1989):
• The section is long enough to provide a representative sample of the regional deformation.

• Displacements on the recorded faults are small compared to the length of the examined section (small-strain theory).

• Spacing of the faults is small compared to the length of the section.

This ellipsoid, and the associated strain tensor that describes the deformation of a unit-sphere to the finite strain ellipsoid, are the sum of the effects of the individual faults (Fig. 6.2). Obviously this only works under the stated conditions above, that the strain is homogeneously distributed and small, say <10%.

6.3.2 3D - Tensor

Based on the above discussion on brittle strain in two dimensions, only few additional parameters are needed to treat strain in three dimensions. First of all, faults are not assumed to be perpendicular to the sample line any longer. This requires the dip-direction (azimuth) of the fault in addition to the dip (plunge) to fully describe the fault-orientation vector \( \mathbf{n} \).

\[
\mathbf{n} = (-\cos(\phi)\sin(\theta), -\sin(\phi)\sin(\theta), \cos(\theta))
\]  

(6.6)

Where the fault dips at an angle \( \theta \) towards \( \phi \). \( \mathbf{n} \) is still orientated perpendicular to the fault plane and points downwards into the footwall.

The slip (displacement) along a fault is now permitted to take on any direction within the fault-plane, from horizontal (strike-slip fault) to dip-slip (normal fault). Thus two additional orientation angles are needed to describe the orientation of the slip-vector.

\[
\mathbf{u} = (\cos(\delta)\cos(\xi), \sin(\delta)\cos(\xi), \sin(\xi))
\]  

(6.7)

Where the displacement direction plunges at an angle \( \xi \) towards \( \delta \). As in the two-
dimensional case, the displacement vector $\mathbf{u}$ must lie within the fault-plane and has a negative sign for normal faults (see Peacock and Sanderson, 1993, for a detailed discussion of this). This allows slip-direction measurements (e.g. from slickenside lineations) to be taken into account.

Apart from these additional components and the modification that the displacement-gradient tensor $D_{ij}$ consists of 3x3 components, instead of 2x2 as in the two-dimensional case, the strain analysis follows exactly the same procedure for determining the components of the strain matrix $E_{ij}$, which is derived from $\frac{D_{ij} + D_{ji}}{2}$ for all faults, and incorporates a weighting factor based on the angle of the fault to the line-section. The determination of eigenvalues and eigenvectors for the orientation and value of the three principal strains ($e_1, e_2, e_3$) from the three-dimensional strain matrix is somewhat more demanding than in the two-dimensional case. This can however be done by applying numerical methods such as the Jacobi method (Jacobi, 1846). The Jacobi method (e.g. Bronshtein and Semendyayev, 1997) is a method for solving matrix equations that can be applied to any matrix that does not contain zeros in its main diagonal. Each diagonal element is solved for, and an approximate value plugged in. The process is then iterated until it converges.

As for the two-dimensional strain tensor, the method was implemented in a spreadsheet (Table 6.3) in which standard macros (Volpi et al., 2003) were used for matrix calculations.

### 6.3.3 Simple tests

Many simple two-dimensional and three-dimensional examples were used to test the 3D tensor template. To illustrate the methodology and the testing, one of the tests is presented here in detail.

Fig. 6.6 shows a theoretical example of a sample line of 100 m length, intersecting 2 sets of conjugate faults. The line trends E-W and the faults are orientated at angles of 45 degrees to the line. Each fault has a displacement of 1 m. The data-input and results can be seen in Table 6.3. Due to the two conjugate fault sets
Figure 6.6: Simple test for the three-dimensional strain tensor. 4 faults with displacements of 1 m and dipping at $60^\circ$, form two conjugate sets. The sample-line has a length of 100 m and is orientated horizontally. Due to the orientations of the two conjugate fault sets which trend perpendicular to each other, $(e_1 = e_2 = -\frac{e_3}{2})$.

trending perpendicular to each other, and the displacements being symmetrically distributed with respect to the section, the finite strain orientation is very simple: $e_1$ is orientated parallel to the sample line, $e_2$ is perpendicular to $e_1$ and horizontal, and both are extensions. $e_3$ is orientated vertical and represents a compression of twice the magnitude of $e_1$ and $e_2$ ($e_1 = e_2 = -\frac{e_3}{2}$). Thus, the deformation represents “uniaxial flattening”.

### 6.3.4 Application to real data

Analogue to the procedure for the above example, three-dimensional strain analysis was carried out for map-scale and cliff-scale sections from the three study areas. Table 6.4 gives an overview of 1D, 2D and three-dimensional strain estimates for different data-sets, together with an estimate of the error between the different estimates.

The cliff-sections from around Kimmeridge Bay show perfect plane-strain conditions with $e_2 \approx 0$. All other samples represent approximate plane strain conditions of deformation with the out-of-plane component typically being one order of magnitude lower than the maximum principal extension ($e_2 \ll e_1$) (Table 6.4). For this reason the three-dimensional strain estimates do not differ significantly from the two-dimensional estimates.
### Table 6.3: 3D-tensor spreadsheet for the example in Fig. 6.6

The eigenvalues and eigenvectors (values and orientations of principal strains) were calculated by applying a macro of Volpi (2003) based on the Jacobian method. Grey fields: Input data; yellow field: Cumulative strain tensor, orange field: cumulative strain matrix, boxes: magnitudes and orientations of finite principal strains.

Most sections show an approximately vertical orientation of $e_3$, suggesting that simple shear components are small. Only one data-set from Malta (Malta C) and the Kilve-Lilstock section show significant deviations from pure shear conditions with rotations in the order of 10° (Table 6.4). The domino-style simple shear deformation at Kimmeridge has already been discussed in some detail in the previous section.

Bedding was observed to be approximately horizontal (no rigid body rotation) across the Maltese Islands. Thus the simple-shear component, observed in the Malta C section, represents a different structural style than that of the Kilve-Lilstock section. Malta C is a map-scale section sampling the southern portion of the North Malta Graben (see Chapter 5). Given that the North Malta Graben is a symmetric full graben system, Malta C represents a sample of the asymmetric southern half of this graben. Thus, it is not surprising, that there is some significant simple-shear component.

Due to the small out-of-plane components of strain in all sections, the differences between three-dimensional and two-dimensional strain estimates are low ($\leq 5\%$). The apparently high error of 17% for both, one-dimensional and two-dimensional estimates for Malta D are not significant because the recorded extension along this...
Table 6.4: Summary of 3D strain analysis. All data-sets approximate plane-strain conditions of deformation with $e_2 \gg e_1$. The magnitudes ($e_1$, $e_2$, $e_3$) and orientations of the maximum principal extensions are listed together with an error estimate for 1D and 2D strain analysis compared to 2D analysis.
section is low (0.5%) and thus close to the measurement error for sample-line data.

6.4 Conclusions

The difference between strain estimates derived from 1D and 2D analysis is directly related to the rotation of the principal stretching axes by the angle $\theta$. Thus, for pure-shear geometry (no rotation) the 1D and 2D results are equal. The larger $\theta$, and thus the simple shear component of the deformation, the larger the error of 1D strain estimates. However, it seems that for $\theta \leq 10^\circ$ the difference is low (<10%), which is the case for all analysed sections.

Three-dimensional strain analysis takes variations in fault trend and slip-direction into account. The difference between strain estimates derived from 2D and 3D analysis is directly related to the out-of-plane component of strain. For sets of dip-slip faults which trend parallel to each other (plane strain conditions: $e_2=0$) the two-dimensional strain estimates are equal to the results of the three-dimensional analysis. However, where the trend of faults is heterogeneous (i.e. several sets of faults with different orientations, or high variability in fault trend of a single fault set), three dimensional analysis is important.

Analysis of cliff-scale and map-scale sections from the three study areas shows that the out of plane component of strain in all sampled sections is either very small ($e_2 \approx 0$), or at least one order of magnitude smaller than the maximum principal extension ($e_2 \ll e_1$). Thus, the error due to the out-of-plane component can be disregarded for the data used in the thesis.

These results validate the one-dimensional approach chosen for the statistical analysis in Chapters 3 to 5 and suggest that the error of one-dimensional strain estimates is generally $\ll 10\%$ of the total extension and is largely dependent on the simple-shear (rotational) component of strain.
Chapter 7

Fault drag

7.1 Introduction

Chapters 3 to 5 deal with extensional strain under plane-strain assumption; Chapter 6 extends this one-dimensional analysis to three dimensions and, at the same time, validates the one-dimensional approach for the analysis of strain-distributions. The present chapter focuses on a rotational, and thus not one-dimensional, component of extension, which is often associated with normal faulting.

The rotation is observed as a bending of material lines (in 2D) or planes (in 3D) in the proximity of faults. In the literature these geometries are referred to as drag folds (e.g. Hamblin, 1965) or flanking structures (e.g. Passchier, 2001). In upper crustal deformation of sedimentary rocks these structures are most commonly observed as the bending of bedding surfaces and occur at a wide range of scales. Depending on the polarity of this bending, which can be convex or concave in the direction of slip, the structure is referred to as normal or reverse drag respectively (Fig. 7.1).

A classic explanation for these structures is frictional resistance on the fault-plane “which causes bedding surfaces near faults to curve in the direction of motion of the opposite fault block” (Twiss and Moores, 1992). This implies that normal faults should be associated with normal drag.

Analogue and numerical models on the other hand suggest that slip on any normal fault should cause reverse drag on both sides of the fault plane with an amplitude
Figure 7.1: Sketch of drag geometries associated with normal faults. (a) Convex bending of bedding planes in the direction of slip is called "normal drag". (b) Concave bending of bedding planes in the direction of slip is called "reverse drag".
that is related to the displacement along the fault and the friction on the fault plane (e.g. Reches and Eidelman, 1995; Grasemann and Stuewe, 2001; Passchier, 2001; Exner et al., 2004). Field observations appear to contradict the model results and add some complexity to the issue:

- The most commonly observed drag structure related to normal faulting is normal drag and not, as predicted by model results, reverse drag.

- Combinations (or super-positions) of normal and reverse drag are a frequently observed feature that has not been explained satisfactory to date.

In this chapter exposed drag structures in inter-bedded carbonates and shales from North Somerset, UK, are examined and a qualitative model is derived to explain the observations. This model is then tested and validated with a simple numerical model. The main objective is to understand how the observed geometries form and to test whether super-imposed normal and reverse drag-folds can be produced during extension, or whether they indicate compression subsequent to extension.

### 7.2 Background

Existing literature on drag-folds can be divided into two groups. The first group concentrates on fault-related folding under upper-crustal conditions and within sedimentary rocks (e.g. Hamblin, 1965; Barnett et al., 1987; Ferill et al., 2005; Porras et al., 2003; Corfield and Sharp, 2000; Berg and Skar, 2005; Fodor et al., 2005). In particular the more recent of these studies were motivated by the high potential for hydrocarbon-traps within drag-folds. Many different models for the occurrence of fault-drag have been proposed, explaining these structures as due to:

- Frictional slip on the fault plane,

- Segmentation of faults along dip or strike,

- Changes in fault dip, producing listric and anti-listric geometries, and
Reactivation of basement faults beneath sedimentary basins.

The second group of publications arose from studies on ductile shear zones. The main motivation for these works is to use drag-folds, mostly referred to as “flanking structures”, as shear-sense indicators. This type of drag-folding has been observed and modelled occurring adjacent to cross cutting elements (such as pinned cracks, veins or dikes) under viscous flow conditions. In these, mostly experimental studies, the fault (or any other cross-cutting elements such as veins or dikes) rotates passively, generally under simple shear conditions, producing normal and reverse drag structures due to viscous flow of the matrix around the fault (e.g. Reches and Eidelman, 1995; Grasemann and Stuewe, 2001; Passchier, 2001; Exner, et al., 2004).

Hamblin (1965) explains reverse drag as due to rollover-folding in the hanging-walls of listric normal faults, Barnett et al. (1987) argue that reverse drag can occur on planar normal faults (Fig. 7.2). They suggest that displacements associated with faulting decrease systematically with increasing distance normal to the fault-surface, which is seen as reverse drag in both hanging wall and footwall of a normal fault. Recent analogue and numerical models support the ideas of Barnett et al. (1987) and suggest that displacement gradients along and normal to a fault are responsible for producing drag-structures (e.g. Reches and Eidelman, 1995; Grasemann and Stuewe, 2001; Paschier, 2001; Exner et al., 2004). Slip along a fault produces a heterogeneous stress and displacement field in the surrounding rock (Pollard and Segall, 1987). In the case of an isolated single fault the slip distribution is elliptical with a maximum at the centre of the fault and zero displacement at the tip-line. Grasemann et al. (2005) show that near the centre of a fault the sense of fault drag is mainly a function of the angle between the marker and the fault plane, with low angles favouring normal drag whilst high angles favour reverse drag.
Figure 7.2: Sketch of reverse drag along a planar normal fault due to the displacement gradients i) within the fault-plane and ii) normal to the fault-plane. Redrawn after Fig. 1b of Barnett et al. (1987).
7.3 Field observations

The observations described in this section were made during field work at North Somerset. The detailed location of the examined section between Kilve and Lilstock, as well as a geological overview of the study area, can be found in Chapter 4.

7.3.1 Normal drag

Normal faults are often flanked by normal drag structures (Fig. 7.3c) and frequently display strong bending of bedding surfaces close to their tip-line (Fig. 7.3a,b,e). In inter-bedded shales and carbonates this bending causes high localized strains that are accommodated by plastic deformation in the softer mudstones (Fig. 7.6f) and by tensile failure, producing veins, in the more competent carbonate beds (Fig. 7.3d). Amplitude and wavelength of the normal drag appear to be largely independent of fault size and displacement but governed by rheology and thickness of the deformed beds. Thin beds and mudstone layers show tighter bending than thicker layers and carbonate beds. Generally the wavelengths of observed normal drag structures, associated with m to 100-m-throw faults, lie between 1 and 10 metres with amplitudes ranging from a few cm up to several metres.

7.3.2 Combination of normal and reverse drag

Combinations of normal and reverse drag structures are another common observation in inter-bedded sedimentary sequences (Figs. 7.4a to c). These structures usually display normal drag close to the fault with a reverse bending located farther from the fault. Unlike normal drag structures, these combinations of normal and reverse drag are usually not symmetrically developed in hanging wall and footwall. They can occur on both sides of a fault-plane but are commonly developed only in one or the other (Fig. 7.4a to c).
Figure 7.3: Field photographs of normal drag structures from the Somerset Coast, UK. (a), (b), (e) Bending of bedding surfaces with "normal polarity" close to the tip of faults. (c), (f) Normal drag, (d) Bending and tensile failure (veining and formation of pull-apart structures) of a carbonate bed close to the upper tip of a normal fault.
Figure 7.4: Field photographs of combinations of normal and reverse drag structures in the hanging wall (a), (b) and in the footwall side (c) of normal faults. Pictures taken at the Somerset Coast, UK.
7.3.3 Small-scale strain associated with drag-folds

The inter-layered shales and carbonates at Kilve accommodate the drag-related strain in different ways. The carbonate layers deform by tensile failure producing veins, whilst the softer shales appear to be “squeezed” in between the stiffer carbonate beds, accommodating deformation in a viscous-flow-like manner.

**Brittle damage in carbonate layers due to bending**

As discussed in Chapter 2 it is possible to determine the brittle extension in carbonate beds by measuring the bedding-parallel thicknesses of veins. Fig. 7.5 shows examples of cumulative vein-thickness and vein-frequency per 0.5 m intervals in carbonate beds deformed by fault drag. The observed damage-zones are between 1 and 10 m wide (distances measured from the fault-plane into the footwall or hanging wall) with most of the bed-scale extension being accommodated within 1 to 2 metres. It can be seen that bed-scale extension often does not simply decrease from a maximum close to the fault to background-strain at greater distance as in Fig. 7.3a, but that there can be several strain-maxima within the damage zone (Fig. 7.3b to e). Typically the highest interval-extensions (up to >20%) are recorded within 0.5 m to 1 m from the fault, with the 2\textsuperscript{nd} and 3\textsuperscript{rd} maxima (1% to 10% extension) being spaced with 1-2 m wide lower-strain intervals in between.

**Plastic damage in mudstone layers due to bending**

The mudstone layers accommodate strain due to bending by distributed (viscous) rather than by discrete (fracturing) deformation. This means that the extension cannot directly be measured by summing fracture apertures as in carbonate beds. However, the exposed shale layers at the Somerset coast contain numerous ammonites, which have been deformed together with their host rock and thus provide samples of the local strain.

Ammonites, as well as many other molluscs, build hard parts in a spiral form, often very close to a logarithmic spiral (Mosely, 1838; Thompson, 1942). Rocha and
Figure 7.5: Frequency and thickness of tensile fractures (veins) observed in carbonate beds which are deformed by fault-drag. The plots show the frequency and thickness per 0.5-m-intervals. The veined zones are between 1 and 10 m wide (distances measured from the fault-plane into the footwall or hanging wall) with most of the bed-scale extension being accommodated within the first 0.5 m from the fault. Examples (b) to (e) show two to three deformation (strain) maxima with a typical spacing between 1 to 2 m.
Dias (2005) developed a template method that makes it possible to get strain estimates even from fragments of deformed ammonites. For this study two drag-folded mudstone beds, containing deformed ammonites, were sampled all the way from the un-rotated hostrock through normal-drag zones to the fault (Fig. 7.6a,b). Photographs of about 50 ammonites were taken, together with bedding-plane orientation and position measurements of each fossil. The pictures were analysed following the procedure described by Rocha and Dias (2005) and the strain was determined for each sample.

The measured extensions range from a few percent far away from the fault up to more than 60% close to the fault planes (Fig. 7.6c,d,f). A clear indication that most of this strain is caused by the bending due to drag is the strong correlation between bedding-dip and recorded extension. Figures 7.6c and d show measured strains plotted against bedding-dip together with the theoretical curves for a simple-shear model of fault-drag. In this plane-strain simple-shear model (Fig. 7.6e) the longitudinal stretch is a function of fault-dip and bedding-dip. The observed data agree well with the measured fault-dips of 80° and 70° at locations 1 and 2 respectively.

As mentioned above, both drag-related bending and extension are confined to a narrow zone adjacent to the faults. The bedding-dip decays from more than 40° close to the fault to horizontal (or background-dip) within few metres from the fault (Fig. 7.6a,b). Similarly, the extension due to bending (drag) decays from >60% close to the fault down to <10% background strain within about 1 metre from the fault (Fig. 7.6f).

7.4 Interpretation of field observations and Qualitative Model

As discussed in Chapter 4, there is almost no evidence for reverse reactivation of the exposed normal faults in the examined section between Kilve and Lilstock at the Bristol Channel (UK). For this reason, the observed drag-structures associated with
Figure 7.6: Distributed (viscous) strain due to fault-drag in mudstone beds. (a),(b) show two sample locations on the tidal platform close to Kilve, Somerset, UK. (c) and (d) are plots of extensions versus bed-dip recorded in deformed ammonites at the two sample locations. Solid lines are predicted extensions (stretches) for different fault dips based on the simple shear model in (e). (e) Simple shear model of fault drag (see text for details). (f) Recorded ammonite strain versus horizontal distance from the fault for location 2.
normal faults, are assumed as having been produced during the basin-forming Mesozoic extension of the Bristol Channel. From the described observations it is possible to derive a qualitative model for the origin of normal and reverse drag structures associated with normal faults (Fig. 7.7).

Initial bending, with normal polarity, forms in the process zone of a (propagating) normal fault (Figs. 7.7a, 7.3a). This deformation stage is expressed by kink-like bending of carbonate beds associated with opening of some tensile fractures (veins) and is similar to the damage in shear-bands described by Reches and Eidelman (1995) and Grasemann et al. (2005). As no discrete displacement (slip) can be observed, this phase is referred to as the ”blind-fault stage”. Bending of carbonate beds, which are stiff compared to the surrounding mudstones, is limited by their flexural rigidity, which is dependent on layer thickness and rheology. The bending causes highest stresses in the regions of highest curvature, which, at a critical value, leads to tensile failure and development of veins (Fig. 7.7b). As the fault accumulates more displacement, the thicker carbonate beds in the process zone (Cowie and Shipton, 1998) start to break. Further strain is accommodated by thickening of veins and opening of pull-apart structures (Figs. 7.7b, 7.3d) until the bed is fully breached. At this stage the fault has propagated through and displaced the carbonate bed (Fig. 7.3e). This produces a normal fault with normal drag (Figs. 7.7c, 7.3c). Following Stearns (1978) this phase is called the “forced faulting stage” here.

The interpreted geometries and their evolution from the Blind Fault to the Forced Fault stage, all with normal polarity of the drag structures, are well supported by field evidence.

The final stage produces reverse drag due to slip on the fault. Due to the preserved earlier normal drag geometry this causes a combination (or superposition) of normal and reverse drag (Fig. 7.7d). At this point this stage is only a hypothesis, as there is no direct field evidence for this model. Due to restricted cliff height it is not possible to observe both zones of normal drag and zones of combined normal and reverse drag along a single fault plane. However, Porras et al. (2003) show seismic
Figure 7.7: Qualitative model for the origin of normal and reverse drag structures associated with normal faults. (a) "Blind-fault stage": Initial bending, with normal polarity, forms in the process zone of a (propagating) normal fault. This deformation stage is expressed by kink-like bending of carbonate beds associated with opening of some tensile fractures (veins). (b) Increase in extension leads to thickening of veins and small displacements may be accommodated by formation of pull-apart structures. (c) "Forced faulting stage": As the fault accumulates more displacement, the thicker carbonate beds in the process zone start to break giving rise to a "normal-fault-with-normal-drag" geometry. (d) Finally, slip may cause bending with reverse polarity which may be superimposed on the earlier normal drag.
images of normal faults from the extensional Eastern Venezuela Basin, displaying all the described drag-geometries (normal, reverse and combined normal and reverse drag-structures) (Fig. 7.8). These images display both, normal and super-imposed drag-folds in the hanging wall of a single normal fault (Fig. 7.8c). To test whether it is possible to produce super-imposed normal and reverse drag during ongoing extension, or whether these structures indicate compression subsequent to extension, a numerical model was set up and is described in the following section.

### 7.5 Numerical Model

#### 7.5.1 Brief review of existing work

Drag structures have been modelled by a number of workers during the last 20 years. Here a brief review of the limitations of these studies is given before describing the setup of the elastic model used in this study.
The key problem with all existing models of drag-folding is that “pinned cross-cutting elements” (CCE), representing fractures, dikes or veins, are used in the model setup. These cracks then rotate passively in a matrix that undergoes pure-shear or simple-shear deformation. No matter how high the strain, the cracks are not permitted to propagate. This approach may be valid for CCE, such as dikes or veins that do not change their shape or size during deformation and are embedded in a matrix that viscously flows around the CCE. However, for drag associated with (brittle) faulting, it appears to be as important to study the deformation occurring around the tips of faults as at their centres.

The field observations presented above indicate that the main cause for normal drag along normal faults is shear-band like deformation within the process zone of a (propagating) fault. This can subsequently be superimposed by slip-related reverse drag. Odonne (1990) deformed a wax plate with a pre-existing fracture in uniaxial compression (Fig. 7.9a). The marker through the centre of the fault (Fig. 7.9a, blue line) shows reverse drag adjacent to the fracture. However, another feature of Odonne’s model, which was neglected in his discussion, is even more interesting for the present study. The two markers close to the tip of the fault (Fig. 7.9a, orange lines) show slight bending with normal geometry that could be interpreted as distributed damage in the process-zone ahead of the fault. Reches and Eidelmann (1995) achieved similar results with elastic numerical models in simple shear deformation, but they concentrated only on the slip-related drag at the centre of a fault. They showed that a marker line through the centre of a pinned fracture orientated parallel to the maximum shear stress will always show reverse drag (Fig. 7.9b). Increase in friction only reduces the amplitude of this reverse drag but cannot invert its polarity. Grasemann et al. (2005) applied analogue and numerical models of viscous flow to the same problem. They showed that changes in the displacement field around a rotating fault in simple shear can cause a switch from normal to reverse drag during ongoing deformation. However, for ”realistic” dips of normal faults (say 40° to 80°) their viscous models also predict reverse drag for a marker
7.5.2 Model setup

Based on the above review a simple numerical elastic model was chosen with the setup shown in Fig. 7.10a to test the qualitative model derived from field observations. The well established code Poly3d, which is a three-dimensional boundary element code based on linear elasticity theory (Jaeger and Cook, 1979), was used. Poly3d has been applied widely to model stress distributions around different fault geometries and the mechanical interaction between faults in two and three dimensions (e.g. Willemse, 1997; Maerten et al., 1999; Kattenhorn et al., 2000; Soliva et
al., 2006). The code models displacements and stresses of earthquake-like single slip events (incremental strains). In the present study the main interest lies in the variations in displacement geometries along a single pinned fault embedded in a linear elastic medium in pure shear. To do this, the elastic bending of two marker layers (Fig. 7.10a, blue and orange lines) is analysed after one slip event on the fault. Figs. 7.10b and c show the vertical and horizontal strain distributions (normalized by maximum strain) after one incremental displacement on the fault.

7.5.3 Discussion and Interpretation of model results

The resulting vertical displacements of the two marker layers are plotted in Fig. 7.11a. The layer through the centre of the fault (Fig. 7.10a, blue line) shows displacement with reverse drag, which agrees with the results of previous studies (e.g. Reches and Eidelman, 1995; Grasemann and Stuewe, 2005; Exner et al., 2004). The layer just above the fault tip (Fig. 7.10a, orange line) shows bending with normal polarity which is in agreement with the analogue model of Odonne (1990) and with the described field observations.

The amplitude and wavelength of the reverse drag in this model is about 3x the dimensions of the normal drag. The amount of vertical displacement (amplitude of the drag structures) in the elastic model is proportional to fault size and amount of slip and is dependent on the elastic properties of the deformed material.

From the field observations it is known that, unlike in the elastic numerical models, the strain caused by drag is not accommodated in an elastic manner but causes permanent plastic deformation. This localises the strain within a narrow (few metres wide) zone adjacent to the fault rather than distributing it over great distance as in the elastic case. A second effect of the plastic deformation is that the deformed layer in the process zone (Fig. 7.10a, orange line) will not bounce back to its original horizontal position when the fault breaks through, but will remain its bending with normal polarity.

Further slip on the (now larger) fault will then produce reverse drag on the orange
Figure 7.10: Model for the pure-shear deformation of a homogenous elastic material with pre-existing fracture. (a) Model setup; (b) and (c) show the vertical and horizontal strain distributions (normalized by maximum strain) after one incremental displacement on the fault.
marker, which will be superimposed onto the earlier normal drag. Fig. 7.11b shows this situation approximated by a simple summation of the normal and reverse vertical displacements from Fig. 7.11a.

This shows that combinations of normal and reverse drag in the vicinity of normal faults can be produced during ongoing extension and fault propagation if the drag-related strain is accommodated by plastic rather than elastic deformation. In this simple model normal drag is caused by fault-propagation whilst reverse drag is produced by fault-slip.

7.6 Discussion and Interpretation

7.6.1 Model predictions

The described model predicts a particular distribution of normal and reverse drag structures along a fault (Fig. 7.12) and thus can easily be tested against observations from field and seismic studies or analogue and numerical models. Assuming that a normal fault initiated at its centre there should be very little normal drag at the centre because the initial fault was small and had little displacement. As the fault grows bigger and propagates, the normal drag produced in its process zones should remain small (or grow only a little bit bigger) as the displacement gradient in the process zone should be about constant (and small as the process zone surrounds the tip of the fault with no displacement). Friction on the fault plane may amplify the normal drag geometry. The amplitude (and wavelength) of the reverse drag should increase as the size and displacement of the fault increases. Thus, depending on the size of the fault and on the point of observation with respect to the tip line and the centre of the fault, either normal drag, reverse drag or combinations of both will be the dominant structure (Fig. 7.12).
Figure 7.11: Results of the numerical model. (a) Vertical displacements of the two marker layers (see Fig. 7.10). The layer through the centre of the fault (Fig. 7.10a, blue line) shows displacement with reverse drag. The layer just above the fault tip (Fig. 7.10a, orange line) shows bending with normal polarity. (b) Superposition of normal and reverse drag approximated as a summation of the two curves in (a).
Figure 7.12: Schematic illustration of the predicted drag geometries at different positions along a normal fault. A: Bending of marker layers with normal polarity in the process zones of a (propagating) normal fault. B: Superposition of normal and reverse drag is most likely to occur between the centre and the tip-line of a normal fault. C: Dominance of slip-related reverse drag close to the centre of a normal fault.
7.6.2 Normal drag

Normal drag is a result of "forced faulting" that causes bending of the layers in the process zone (close to the tip line) of normal faults. Bending of carbonate beds, which are stiff compared to the surrounding mudstones, is limited by their flexural rigidity that is dependent on layer thickness and rheology. Once this "critical bending" is exceeded, the layer breaks and the fault plane propagates through the layer rather than bending it further. Due to permanent deformation (plastic strain and tensile fracturing) the bending remains even after the fault has propagated through the layer. This preserves normal drag geometries adjacent to normal faults. Frictional resistance to sliding on the fault plane may amplify this normal drag but cannot initiate it.

7.6.3 Reverse drag

Reverse drag is produced by slip on a normal fault and shows the maximum amplitude around the zone of maximum displacement on the fault surface. In both, viscous and elastic models the amplitude of the reverse drag is a function of fault size and displacement. Increased friction on the fault plane reduces the magnitude of the slip, and thus the amplitude of the reverse drag, but cannot invert its polarity.

7.6.4 Combination of normal and reverse drag

When a normal fault grows in size, it propagates through its earlier process zone. As slip on the fault accumulates, normal drag, which had formed in the previous process zone, becomes superimposed by slip-related reverse drag. This process can produce complex double-bent drag structures during ongoing extension.

7.6.5 Wavelength and amplitude of drag structures

One major difference between the observed and the modelled structures is their different wavelengths. Observed drag structures typically extend only few metres from
the fault into the wallrock whilst the modelled elastic structures show a wavelength of about 100 m at comparable amplitudes. These differences are not surprising as the studied rocks are no perfect elastic materials but would better be represented as elasto-viscous or elasto-plastic materials. Four factors are likely to govern the wavelength (and amplitude) of drag-structures:

1. In the isotropic, homogenous elastic model wavelength and amplitude of the drag structure are proportional to fault size and displacement. The observed rocks on the other hand deform plastically (brittle tensile failure of stiff carbonates) and viscous (flow-like deformation of soft mudstones) which localises the deformation and drastically reduces the wavelength of the drag structures.

2. Another important factor is the flexural rigidity of the stiff layers (carbonate beds). In the elastic model a single thick layer with typical limestone properties was used. Flexural rigidity is governed by the elastic moduli and thus is strongly dependent on the layer thickness. Thus, bending (drag) will be more localized in interbedded carbonates and mudstones than in massive carbonates.

3. Friction on the fault surface may reduce both, the amplitude and the wavelength of the drag structure.

4. The shape of the fault plane may influence the geometry of the produced drag structures. In the elastic model the fault is represented as a planar surface. In nature fault planes are curved and show steps which may locally tighten the bending.

To sum up the above discussion it can be said that simple elastic models help us to understand the origin of the observed structures but are not a representation of the real processes. To better represent the deformation, more sophisticated models are needed that would allow i) fracture propagation, ii) elasto-plastic deformation, iii) elasto-viscous flow and iv) layering of different mechanical properties and v) friction on the fault plane.
7.7 Conclusions

In this chapter observed normal drag and super-imposed normal and reverse drag structures in inter-beded carbonates and shales were examined and their origin explained with the help of a simple elastic model. Normal drag is shown to be produced in the process zone (close to the tip) of normal faults in a shear-band-like manner and can become superimposed by reverse drag as the fault propagates and accumulates displacement. Friction on the fault plane can i) amplify the magnitude of normal drag structures and ii) reduce the magnitude of reverse drag but is not required for the initiation of either geometry. It has been shown that both normal and reverse drag structures can be produced during extension (propagation of, and accumulation of displacement on, normal faults). Thus, it can be concluded that, although it is tempting to use this argument in the field, super-positions of normal and reverse drag are no valid indicator for compressional reactivation of normal faults. However, compression subsequent to extension, as reported for many sedimentary basins, may amplify and tighten the originally extension-related drag-structures. In detail it has been shown that:

- Drag in the vicinity of normal faults accommodates a high localized bed-scale extension of up to >60% in mudstone layers that is accommodated by continuous simple shear. In the stiff carbonate layers drag-related extension is accommodated by brittle fracturing (tensile veining) with interval-strains of up to >20%.

- Simple elastic models show that all observed drag structures (normal, reverse and superimposed normal and reverse drag) can be produced during extension and without friction on the fault plane.

- Any normal fault may display normal and reverse drag and the combination of the two geometries at different positions relative to the tip line and the region of nucleation of the fault.

- Normal drag can be expected closer to the tip line whilst reverse drag and
superimposed reverse and normal drag can be expected closer to the centre of a fault.

- Normal drag initiates within the process zone (close to the tip line) of a normal fault due to shear-band-type bending, and subsequent brittle failure, of the beds, and propagation of the fault plane through the bent layers (forced faulting).

- Reverse drag is produced by slip on a normal fault and can be super-imposed on the earlier normal drag.

- Friction on the fault plane may amplify normal drag structures and reduce the magnitude of reverse drag structures.
Chapter 8

Discussion and Conclusions

The aims of this study (Chapter 1) were to use direct measurement of the spatial and size distributions of normal faults and associated tensile fractures in sedimentary rocks formed during extension to:

- Quantify regional extension at different scales and in particular below seismic resolution limits.
- Determine the relative importance of large and small structures in accommodating regional extension.
- Quantify the spatial heterogeneity of brittle extension at different scales and in particular relationships between large and small structures.
- Compare the scaling laws for veins and faults belonging to the same extension event.
- Understand how brittle deformation evolves in space and time during extension of a region.
8.1 Extension at different scales and importance of small structures

8.1.1 Direct measurements of extension at different scales

As discussed in Chapter 2, extension in the upper crust is largely accommodated by tensile fractures (joints and veins) and normal faults. Measurements of vein apertures and fault heaves provide simple and robust measures of extension that can be directly related to the location and orientation of the structures. This allows analysis of spatial distributions of both extension and structures. Three-dimensional strain analysis based on 2D and 3D tensor representations (Chapter 6) confirmed that one-dimensional estimates of extension (based on the sum of heaves and apertures) provide good approximations of the maximum principal strains in all studied sections.

Detailed field studies allow measurements of extensions of <1 mm on individual structures. Long continuous (cliff-) sections record faults below seismic reflection resolution and high-resolution geological maps or seismic data provide regional fault-scale information. A hierarchical sampling approach (Chapter 2) was used for the integration of information at different scales. This permits assembling a complete picture of the brittle deformation that accommodates regional extension over the entire scale range.

8.1.2 How important are fractures of size below seismic resolution?

The three field studies (Chapters 3 to 5) have shown that faults below seismic resolution accommodate significant portions of the total extension in a region (Table 8.1, Fig. 8.1). Assuming that high quality seismic data detect all faults with displacements $\geq 10$ m, between 20% (in high strain zones) and 70% (in low strain zones) of the total extension accommodated by faulting may be missed. Note that
Figure 8.1: Total extension and seismically resolved extension in the three study areas, based on a seismic resolution of fault-displacements ≥10 m. (a) Plot of actual regional extension against seismically observed extension in log-log space. (b) Proportion of seismically resolved extension (in % of total extension) against total extension in lin-log space.

A lower limit of resolution ≥10 m is a very optimistic estimate and for many (older) commercial 2D seismic lines this limit will be closer to 50 m (e.g. Brown, 2003). Comparing the seismically resolvable extension (i.e. extension accommodated on faults with displacements ≥10 m) with the total regional extension in the three study areas (Fig. 8.1a) brings forward a remarkable relationship: The seismically observed extension scales with the total extension obeying a power-law relationship with a scaling exponent of 0.8. In other words, the proportion of the total extension that is resolved in seismic reflection data systematically increases with increasing strain. This indicates that small structures are of highest importance in lower-strain regimes and that large structures dominate higher-strain regimes.

Plotting the seismically resolvable extension as a proportion of the total extension for the three study areas (Fig. 8.1b) shows that this relationship displays a logarithmic correlation. Extension estimates based on seismic data underestimate the total extension by more than $\frac{1}{3}$ at strains smaller 10% but record $>\frac{2}{3}$ of the total extension at strains higher than 10%. In the following sections the potential implications of these results with respect to crustal stretching factors, seismic moment deficit and fluid flow are discussed.
Table 8.1: Summary of extension estimates for the three study areas. The second column gives the estimated total extension for each area. The third column gives the proportion of the total extension that would be resolved in high quality seismic reflection data (fault-displacement $\geq 10$ m). Fourth and fifth columns give the proportion of the total extension that is accommodated on structures (small faults and vein) below seismic resolution.

### 8.1.3 Crustal stretching factor

White and McKenzie (1988) proposed that the amount of extension in the lithospheric mantle should be equal to that in the crust to avoid space problems. However, a common observation in studies of extensional basins and continental margins is that the amount of extension visible on normal faults (usually observed in seismic reflection profiles) is significantly lower than the amount of extension (or thinning) estimated from crustal or lithospheric thickness and thermal subsidence (e.g. Walsh et al., 1991; Reston, 2007). Two different arguments were given to explain this extension discrepancy:

1. Significant amounts of extension accommodated by small faults (i.e. faults of size below seismic resolution limits), which account for the difference (Walsh et al., 1991; Marrett and Allmendinger, 1992) or polyphase faulting accommodating extensions of $>100\%$ (i.e. $\beta > 2$) (Reston, 2005).

2. Depth-dependent stretching of the lithosphere in which the upper crust is extended and thinned less than the rest of the lithosphere (Sibuet, 1992; Driscoll and Karner, 1998; Kusznir et al., 2005).

Taking the two examples of Kilve and the Maltese Islands, it can be seen that extension estimates based on standard seismic data or geological maps will under-estimate...
the extension by about 40% (Table 8.1). Very similar results have been found by Walsh et al. (1991), Marrett and Allmendinger (1992), Pickering et al. (1996) in the North Sea Basin. These authors conclude that resolvable faults only account for about 40% to 75% of the total extension. The differences in amount of extension resolved at Malta in higher and lower strain zones (Table 8.1) suggests that a larger proportion of the total extension is missed in lower strain regimes.

This can be explained based on the evolution of extensional systems which will be discussed in section 8.4. It has been shown that extension in a region evolves from many (distributed) smaller active faults to progressive localisation of strain onto the largest faults with increasing strain. This is seen as a change from higher to lower scaling exponents for fracture frequency plots as extension increases. Thus, the higher the strain in a region the higher the proportion of large faults, which means that more faults (and strain) are detected in seismic sections. These observations partly support the idea that significant amounts of extension are accommodated by small faults. The results of this study show that this argument is valid for regions with relatively small strain (say <10%) and/or high scaling exponent (say close to or >=-1) but does not hold for higher extension regions or areas with high localisation of strain.

None of the data in the thesis allow testing of the depth-dependent stretching of the lithosphere. However, recent studies based on modern high-angle seismic techniques (e.g. Reston, 2007) suggest that deviations from uniform stretching are not common in (non-volcanic) crustal extension and certainly are much lower than what would be required to explain the extension discrepancy between measured fault-extension and crustal stretching factors.

An alternative explanation may be found in the results of Chapter 6. McKenzie’s (1978) uniform stretching model and most of its derivatives assume plane strain pure shear geometry and the quantitative description based on both fault-controlled and thermal subsidence is essentially one-dimensional. In Chapter 6 it was concluded that the one-dimensional analysis may underestimate the maximum extension in a
region significantly if the faults show considerable strike variation or oblique slip components. In this case $e_2 \neq 0$ and the strain involves an out-of-plane component. Given that most regional seismic sections are two-dimensional, this error can contribute to the under-estimation of the total extension.

### 8.1.4 Seismic moment deficit

The traditional measurement of earthquake size is *magnitude*, which is a logarithmic scale, based on the amplitude of a seismic wave measured at a particular frequency (seismogram), suitably corrected for distance and instrument response. Due to different propagation paths of earthquake waves to different stations determined magnitudes can vary considerably from one station to another. An alternative, and physically more meaningful, expression of the magnitude or size of an earthquake is the *seismic moment* (e.g. Scholz, 2002).

\[
M_{0ij} = \mu (\Delta u_i n_j + (\Delta u_j n_i)A)
\]

Where $\Delta u_i$ is the mean slip vector averaged over the fault surface $A$, with the unit normal $n_j$, and $\mu$ is the shear modulus. One of the advantages of using the seismic moment rather than earthquake magnitude is that it can be applied to studies of both active and inactive deformation.

Because $\Delta u_i$ and $n_j$ are vectors, $M_{0ij}$ is a second rank tensor with a scalar value $M_0 = \mu \Delta u A$, and the two directions that define slip and fault orientations. Thus, the seismic moment applied to seismological problems (Molnar, 1983; Marrett and Allmendinger, 1990) is similar to the displacement gradient tensor applied to field data that, as discussed in Chapter 6, can be derived from field measurements of fault geometry and displacement.

A more recent method of determining seismic moments of actively deforming regions is based on geodetic strain rate estimates. The total seismic moment of a region depends on the total area of, and displacement on, faults in an area and on the time interval over which the data are collected. Accurate geodetic data have only been

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available for about one or two decades. For this reason the determined strain rates represent a “snap-shot” in geological time and are strongly dependent on short-term variations. An advantage of this type of data is that they record both, seismic and aseismic deformation. Earthquake measurements and historic catalogues cover several decades to several hundreds of years and thus may sample several earthquake cycles. However, the resolution (size of smallest earthquakes recorded) is limited and little information on small-scale deformation and aseismic slip can be derived from these data. Geological field observations represent an “integrated sample” of past deformation over long time periods and may be expected to sample all the deformation in a region, usually with no or limited information on whether or not all of the observed structures were active at the same period. Thus, geological estimates sample the resulting finite strain over many (possibly 100s or 1000s) seismic cycles.

Considering these fundamental differences between the three methods for estimating the total seismic moment of a region with regard to the evolution of fault populations, as discussed in section 8.4, it is not surprising that the results often differ significantly from each other. In low strain settings (large scaling exponent of fault frequency) the earthquake based strain estimates can be expected to be too low due to under-sampling of small-scale deformation. Geodetic data would record both, seismic and aseismic (small-scale) deformation but may show a considerable error due to the short time-interval the estimates are based on. Geological record (provided high resolution) should yield higher strain estimates than the seismic record because it takes both, active and inactive faults into account.

Similar conclusions were also drawn by Walsh et al. (1991) who suggest from observed fault displacement populations that faults with small displacements make a greater contribution to the finite strain than they do to the infinitesimal (seismic) strain. This may appear paradoxical given that the geological faulting is presumed to be produced by repeated earthquake cycles. However, considering the evolution from initially more distributed deformation to progressive localisation of strain with increasing extension, it makes sense to argue that active fault-populations at any
time will differ from the total population that consists of active and dead faults. None of the examples in this thesis are from currently active areas, although seismicity is recorded from Malta. More work of the sort presented in this thesis in areas of recent seismicity is needed.

8.2 Heterogeneity of brittle deformation and extension

8.2.1 “Organisation of brittle deformation”

The field studies in this thesis show different organisation of extension-related deformation at several length-scales. Within the platform carbonates of the Maltese Islands for example, extension produced a horst-and-graben morphology at a scale of observation $>10$ km. This suggests a localisation of extension within the North Malta Graben that is flanked by largely undeformed Gozo and Malta Horsts. Zooming in by an order of magnitude reveals that most of the strain is accommodated within km-wide damage-zones, separated by virtually undeformed regions. The damage-zones are associated with major faults and are observed both within the graben and the horsts, but are more prominent within the graben. Virtually all of the deformation is accommodated by faults (Chapter 5).

At Weymouth Bay (Kimmeridge study area) extension is fairly uniformly distributed at both 10-km and km-scales. Zooming in even further to the scale of single beds shows that each fault is surrounded by a damage zone with tens to hundreds of thin veins (Chapter 3) and that apparently undeformed zones in between damage zones contain randomly distributed tensile fractures.

The cliff-section between Kilve and Lilstock (Somerset Coast) also shows fairly uniformly distributed extension at the km-scale. Zooming in to the scale of single beds however shows high localisation of extension within narrow zones ($<10$ m wide). This bed-scale deformation is accommodated by intense veining of carbonate beds.
and by distributed shear failure in mudstone layers. Regions in between these damage zones are virtually undeformed.

To quantify the observed heterogeneity of fracture distributions and strain, at any length scale and resolution, a new method for spatial heterogeneity analysis for line samples was developed.

8.2.2 Quantifying heterogeneity

A new method for quantifying heterogeneity of brittle deformation has been introduced in Chapter 2 and was applied to spacing and displacement data from three areas that have undergone extensional deformation (Chapters 3 to 5). The method was initially developed to provide a measure of strain heterogeneity, based on both position and displacement of individual fractures, sampled along a linear traverse. However, it can be applied to any property that shows a spatial distribution and can be sampled along a line (e.g. cumulative number of fractures, cumulative heave, seepage quantities, permeability of intervals...).

Statistically uniform (homogeneous or random) distributions of fractures and strain yield a heterogeneity measure of $V' \rightarrow 0$, whilst localisation of fractures and strain in a point along the line yields $V' \rightarrow 1$. Plotting the heterogeneity of strain distribution ($V'_S$) against the heterogeneity of fracture spacing ($V'_F$) can reveal differences in deformation patterns between different extensional regions (Fig. 8.2a). Plotting the heterogeneities from the three study areas in this way (Fig. 8.2b) shows that there is a (surprising) conformity between the data from the Maltese Islands and from the Kilve-Lilstock section whilst the data from Kimmeridge Bay show a somewhat different trend. The Malta and Kilve data show exactly the same signature, indicating that the heterogeneities of spacing and strain are equal in both areas. The Kimmeridge data on the other hand shows low heterogeneity in fracture spacing but significantly higher heterogeneity in strain distribution.

These differences can also be observed in a qualitative plot of heterogeneity against the scale of observation (Fig. 8.2c). It can be seen that at Kilve heterogeneity is
equally high for strain and fracture spacing at the bed-scale and decreases with increasing scale of observation. At Malta lower-strain zones generally show somewhat higher heterogeneity than higher-strain zones. However, in both higher and lower strain zones the heterogeneities of strain and fault spacing are linked and show little variation from the cliff-scale to the map-scale. At Kimmeridge the heterogeneities are similarly low at the bed-scale with veins and accommodated strain being randomly distributed. At the cliff-scale however, faults are still randomly spaced whilst the strain accommodated on these faults is significantly localised. At the map-scale the relationship is inverted again with the faults still showing random distribution but the strain being fairly uniformly distributed with some indication of higher localisation towards the high end of the scale-range.

These differences may be related to differences in lithology, response to early extension and magnitude of regional strain and are discussed in more detail in Section 8.4.

8.3 Comparison of scaling laws

Deformation in the layered sequences of Kimmeridge Bay and along the Kilve-Lilstock section is accommodated by both tensile fractures (preserved as calcite veins) and normal faults with observed displacements ranging from $\ll$ mm to $>100$ m. In the massive carbonates at the Maltese Islands virtually all the extension is taken up by faults with mm to $>100$ m throws. In all three study areas larger structures may exist but are not exposed onshore or sampled in seismic data.

The frequency-data for all study areas appear to conform to power-law distributions as reported widely for faults and fractures (Walsh et al., 1991; Jackson and Sanderson, 1992; Marrett and Allmendinger, 1992; Pickering et al., 1994). However, there are some significant differences in the fracture frequencies between and within study areas as can be seen in Fig. 8.3 where the average frequency of fractures is plotted on log-log scale for intervals of heave or throw. Note that these plots are for discrete bins and not the more conventional cumulative plots (Pickering et al., 1995).
Figure 8.2: (a) Schematic "map" of the relationships between the heterogeneity measures for strain (VS') and fracture distribution (VF'). (b) Plot of heterogeneity measure VS' (strain heterogeneity) against VF' (spacing heterogeneity) for the three study areas. (c) Qualitative representation of trends of spacing heterogeneity (dashed arrows) and strain heterogeneity (solid arrows) for Kimmeridge (orange), Kilve (yellow) and Malta (blue). Trends for higher and lower strain zones for Malta are shown in light blue and dark blue respectively.
At Kimmeridge Bay (Fig. 8.3a) the larger faults (≥10 m heave) conform to a power-law distribution, as do the veins (<0.1 m aperture), both having scaling exponents of ≈0.95. However, the veins and faults are separated by a transitional region, with heaves between 0.1 and 10 m, and thus do not form part of the same power-law distribution.

Fractures from the Kilve-Lilstock section (Fig. 8.3b) conform to a single power-law scaling over the observed scale-range if the largest heave-interval (fault-heaves of 30 to 100 m) is excluded to avoid bias due to under-sampling. The scaling-exponent for faults with heaves between 0.1 m and 30 m is surprisingly low (-0.41) suggesting that the smallest faults in the section are of minor importance.

Fault-frequencies at the Maltese Islands were determined for higher-strain (damage) zones and lower strain zones separately (Fig. 8.3c). The data for both, higher and lower-strain zones appear to obey power-law distributions over a throw-range of 2 to 3 orders of magnitude. However, the data do not share a single scaling-relationship. The lower-strain areas show a steeper slope \( D = -1.26 \) than the higher-strain-zones \( D = -0.84 \) with a division at a scaling exponent of \( D \approx -1 \).

### 8.3.1 Influence of mechanical layering

All limestone and dolostone beds at Kimmeridge Bay are separated by several metres of mudstone and the veins are observed in the carbonate beds only. The transition in the power-law scaling relationship of fracture frequency from Kimmeridge Bay (Fig. 8.3a) occurs at a length-scale of about 0.1 m to 2 m. This corresponds to the typical thickness of carbonate beds within the Kimmeridge Clay and suggests that the observed deviation from a simple power-law relationship is related to the bed-thickness. This agrees with a common observation that vertical confinement strongly influences spacing and size-distributions of fractures (e.g. Knott et al., 1996; Ackermann et al., 2001; Schulz and Fossen, 2002; Soliva et al., 2006). Small faults with displacements in the range of layer thickness, say <3 m, appear to be underrepresented. This may imply that small faults are suppressed or arrested at
Figure 8.3: Fault and vein frequencies in the three study areas. (a) Kimmeridge Bay, (b) Kilve-Lilstock section, (c) Maltese Islands. The solid lines are best-fit power-law trends, the scaling exponents (D) are indicated. Arrows indicate the frequency of faults with a heave (a) and (b) or throw (c) of 10 m (see text for details).
carbonate-shale interfaces (Soliva et al., 2006). Faults with displacements larger than the bed-thickness appear to have developed undisturbed by the mechanical layering. Similar observations have been made by Odling et al. (1999) who distinguish two end-member fracture systems (at the reservoir scale): i) Stratabound systems, in which fractures are confined to single layers, with regular spacing and scale-restricted size distributions, and ii) non-stratabound systems, in which fractures show a wide range of sizes, usually with power-law distributions, which may be spatially clustered and vertically persistent.

8.3.2 Influence of strain

Comparing the slopes (scaling exponents) in the fracture frequency plots from the three study areas (Fig. 8.3) shows a negative correlation between bulk extension and slope. The highest exponent \( D = -1.26 \) is observed in lower strain zones at Malta (extension <0.5%). Higher strain (damage) zones from the same area show \( D = -0.84 \) (Fig. 8.3c). Similar extension (≈7%) at Kimmeridge produces a slope with \( D = -0.94 \) for faults. The highest extension (≈25%) was recorded along the cliff section between Kilve and Lilstock (Fig. 8.3b) with a scaling exponent \( D = -0.41 \). Even though power-law scaling exponents derived for different study areas and lithologies can only be compared with caution these trends appear to support the concept of evolution of extensional systems discussed in Section 8.4.

Similar evolution of regional deformation in response to extension from more or less randomly distributed small scale (tensile) fractures to increasing localisation of strain onto the largest faults in the system has been described based on field observations (e.g. Meyer et al., 2002; Bailey et al., 2005; Walsh et al., 2003) and analogue and numerical models (e.g. Gupta and Scholz, 2000; Hardacre and Cowie, 2003).
8.3.3 Considerations on deformation

beyond limits of observation

Representativeness of the collected data for the regional deformation

The length-scales (extents) of the extensional systems in the three study areas are in the order of 25 km and 22 km for Kimmeridge Bay (E-W extent Weymouth Bay fault system) and Kilve (N-S extent Bristol Channel fault system) respectively. The North Malta Graben is about 15 km wide (NW-SE) and is separated from the adjacent horsts by the graben-bounding South Gozo Fault and Victoria Lines Fault. The length-scale of the entire system (including horsts and graben) is in the order of 40 km.

Comparing these dimensions to the total length of line-samples (Kimmeridge: $\approx 22$ km, Kilve: $\approx 2$ km, Malta: $\approx 56$ km) it can be argued that the data-sets from Kimmeridge and Malta are representative for the regional deformation whilst the Kilve data only sample a small portion of the Bristol Channel system. In terms of size and geometry, the Bristol Channel can be compared to the North Malta Graben. However, the recorded extension along the Kilve-Lilstock section ($\approx 25\%$) is about one magnitude higher than the mean extension across the Maltese Islands ($\approx 2.4\%$). As it is unlikely that basins (or grabens) of similar dimensions (length-scale and accommodation space) would show extensions that differ by an order of magnitude, it can be hypothesised that the sampled section at Kilve is not representative for the regional extension but rather represents a high strain (damage) zone. Such a high strain zone may represent a large-scale damage zone, associated with a major (basin-bounding) fault, as observed adjacent to the Victoria Lines Fault on Malta. Given the high extension accommodated it could even be argued that the Kilve-Lilstock section could represent a linkage zone (relay) in between two major (overstepping) faults.

This hypothesis contradicts the proposed half-graben model for the Bristol Channel (Brooks et al., 1988) that explains the graben as being caused by extensional reac-
tivation of a major south-dipping Variscan thrust located on the northern margin of the basin. A different model, which proposes two major (overstepping) north-dipping normal faults at the southern margin of the Bristol Channel, and thus explains the Bristol Channel as a full graben, is the one proposed by Peacock and Sanderson (1999). The postulated east-west striking faults are estimated to have throws in the order of 1000 m. The Kilve-Lilstock section with its high strain (≈25%) and domino-style deformation may represent a relay zone in between these two overstepping (and at depth potentially connected) normal faults. The domino-style deformation along this section (Chapter 6) could be related to this structural position within a km-scale relay ramp and with a weak detachment in Triassic halite bearing mudstones underlying the deformed Jurassic sequence (Stewart and Argent, 2000). Based on these hypotheses it can be predicted that there must be lower-strain zones of similar length-scales as the Kilve-Lilstock section with scaling exponents closer to $D = -1$ within the Bristol Channel Basin.

**Largest Faults in the System**

Scaling relationships (Fig. 8.3) represent the size-distributions of observed faults (and veins) in the region under investigation. For example, only one fault with a heave of 10 m is to be expected in a 1 km long section at Kimmeridge (Fig. 8.3a, arrow) and similar frequencies are observed in higher strain zones on the Maltese Islands (Fig. 8.3c, upper arrow). Along the Kilve-Lilstock section on the other hand one fault with a heave of 10 m can be expected every 100 m of section (Fig. 8.3b, arrow). In lower strain zones on the Maltese Islands only one fault of this size would be observed in a section of about 17 km length (Fig. 8.3c, lower arrow).

These relationships also have predictive power as they can be used to estimate the size of the (potentially unobserved) largest faults which are likely to occur in the system. Similar to the above considerations based on the occurrence of a 10-m-heave fault, it is straightforward to estimate what the heave (or throw) of the largest fault in a system of a certain length-scale should be. The largest observed faults in the
three study areas have throws of 221 m, 61 m and 182 m at Kimmeridge, Kilve and Malta respectively.

The length-scale (extent) of the extensional system at Kimmeridge Bay lies in the order of 25 km as discussed above. Based on the observed scaling relationships for this region (Fig. 8.3a), it can be predicted that no more than one major fault with a heave between 100 and 300 m is likely to exist in this system. Given that the largest sampled fault has a heave of 192 m it can be assumed, that no major fault exists beyond the sampled sections.

The same exercise can be carried out for the Maltese Islands (Fig. 8.3c). Given the length-scale of 15 km for the North Malta Graben and a total of 40 km for the entire horst-and-graben system, it can be seen that no major faults were missed in lower strain zones apart from potentially one additional fault in the throw interval 10 - 30 m. For higher strain zones it can be predicted that 2 to 3 faults with throws in the range 100 - 300 m and potentially one fault with a throw >1000 m may exist beyond the limits of the sampled sections. These faults may be located in the central part of the North Malta Graben between Gozo and Malta, which is the only section that is not covered by the geological map and cliff-sections because it is under the sea.

These considerations support the above statement, that the data collected at Kimmeridge and Malta are representative for the regional deformation over the entire size of faults and fractures present in these extensional systems.

For the Kilve-Lilstock section it is not as straightforward to determine the size of the largest faults in the system. As discussed above, the N-S extent of the Bristol Channel is about 22 km of which the sampled section only covers 1.6 km. Assuming that the high-strain section between Kilve and Lilstock is representative for the entire Bristol Channel would predict very unrealistic maximum fault sizes (heaves between $10^6$ and $10^7$ m) when interpolated to the size of the system. Thus, as discussed above, it is clear that the sampled section is not representative for the entire extensional system but more likely represents a large-scale damage zone.
8.3.4 Considerations on the significance of fracture frequency data

The method used for data collection throughout this work is a special case of one-dimensional sampling. Most scan-lines used in this study trace single beds or marker horizons across faulted regions and record the damage that this particular bed or horizon has experienced. Provided that the length of the line is sufficient to yield a statistically representative sample of the deformation, this method is very useful for the analysis of magnitude and heterogeneity of fracture spacing and strain as presented in Chapters 2 to 6.

However, in the strict sense, such samples are only true one-dimensional samples after restoring the fault displacements to bring the marker back to its undeformed (initial) geometry. In that respect the applied sampling method is a "Langrangian approach" as it considers what happened to the initial geometry of the marker during extension. Thus, in Chapter 6 the reference is the undeformed state and the description of deformation is a "forward" description. So far these considerations do not have much implication on the results of the study as both heterogeneity analysis and strain analysis will yield correct results.

However, when discussing the data in terms of the frequency of faults and tensile fractures (veins) there are some issues that should be considered. Veins usually are restricted to single beds, at least in sequences of inter-bedded carbonates and mudstones, whilst faults commonly cross-cut many layers. Thus, the derived fracture frequencies represent only the number of faults and veins that deform one bed (or horizon).

For many applications (e.g. fluid flow in fractured rock) it is however more important how many fractures can be expected within a volume of rock rather than in a single bed. Faults may intersect the entire volume of interest, or significant portions of it, whilst veins will occur in each bed largely independent of the deformation in the other beds within the volume. For this reason the derived vein frequencies for Kimmeridge Bay (Fig. 8.3a) and the Kilve-Lilstock section (Fig. 8.3b), both of
which are inter-bedded carbonates and mudstones, would need to be multiplied by
the number of carbonate beds within the volume of interest. In addition, depending
on the dimensions of the volume of interest, correction factors for faults that do not
intersect the entire volume, would have to be applied.

8.4 Evolution of brittle deformation in extensional
regimes

Early extension has been shown to be accommodated in different ways in the three
study areas. Differences in the lithologies appear to be the dominating factor for
caus[[...]]g early localisation or distributed deformation. The fairly homogeneous and
stiff platform carbonates at Malta accommodate early extension by randomly dis-
tributed small-scale faulting. As extension increases some favorably orientated and
located faults grow bigger and link to form larger structures whilst others become
inactive (Fig. 8.4a). Similar behaviour has been described for sandstones (e.g.
Joussineau and Aydin, 2007) and other “massive” (as oppose to layered) rocks
(e.g. Odling, 1999).

The layered but mudstone dominated sequence at Kimmeridge Bay responds to low
strains by “ductile” deformation in the shales and by randomly distributed to anti-
clustered tensile failure (veining) in the stiffer carbonate beds. Increased extension
localises displacement onto a few regularly spaced minor faults and opens veins to
form pull-apart structures (Fig. 8.4b). This behaviour has also been observed in
analogue (Horsfield, 1977; Withjack et al., 1990; Mandl, 2000) and numerical models
(e.g. Schoepfer et al., 2006) with rheological layering.

In the layered sequence at Kilve stiff carbonate beds are more prominent and may
govern the overall strength of the sequence. Early extension in this area is highly
localised in narrow zones (typically 1 to 5 m wide) in the carbonate beds, preserving
large portions of virtually undeformed rock in between. The more ductile mudstone-
layers accommodate initial deformation in a more distributed manner. As extension
Figure 8.4: Conceptual models of the early-stage evolution of brittle deformation in the massive platform carbonates at Malta (a) and the layered sequences of Kimmeridge (b) and Kilve (c) respectively. Extension increases from top to bottom. Pictures to the right of the sketches show typical field examples of the different stages of deformation.
increases, small displacements are accommodated by bending of the carbonate beds and opening (thickening) of veins in the process zones of initiating and propagating faults (Fig. 8.4c).

Increase in strain appears to have developed populations of small, randomly distributed faults from some of the early fractures in all study areas, whilst most of the remaining (small) fractures died out. This stage of deformation is preserved in the low heterogeneities (random distributions) of cliff-scale faults in all areas (e.g. Fig. 8.2c).

Further extension localised deformation on fewer but larger faults and their associated damage zones with the largest faults gaining more importance and many smaller faults becoming inactive. These observations agree with recent observations and numerical models (e.g. Cowie et al., 2007).

8.5 Future work

Discussing the outcome and results of a project and thinking about unresolved questions is a good starting point for generating ideas on where to go next. For this reason a few thoughts on potential future research that may advance heterogeneity analysis of brittle deformation are discussed below.

8.5.1 Heterogeneity studies at larger scales

It has been shown that low to moderate strain regions are particularly useful for the analysis of the evolution of deformation. At higher strains the record of localization may be lost or is too difficult to interpret (Cowie et al., 2005). All three field studies presented in this thesis were carried out in extensional systems with a length-scale between 10 km and 40 km and are based on the deformation observed in upper-crustal sedimentary rocks. It would now be interesting to extend the study to the scale of major extensional basins (i.e. length scales of >500 km) and compare deformation patterns and evolution between different (types of) basins. Depending
on the availability and resolution of data it would also be interesting to compare
deformation and strain accommodated in basement and cover sequences.

8.5.2 Heterogeneity studies on growth-faults

Extensional regions that are deformed by growth-faults (i.e. syn-sedimentary faults in settings where rate of sediment supply is equal or exceeds the displacement-rate) are a promising field for heterogeneity and evolution studies. Provided that high-resolution (seismic) data are available, such regions can be analysed by comparing older with younger horizons. In other words such data-sets provide several samples of different strain magnitude of the same extensional system. By applying backstripping methods it is possible to determine displacement histories of faults that were active across several horizons. Thus the temporal and spatial evolution of a population of faults may be directly observed from older to younger horizons (e.g. Childs et al., 1993; Walsh et al., 2003, 2004).

8.5.3 Heterogeneity parameters and tunnelling

As discussed previously, the method for quantifying heterogeneity from cumulative data-sets is not restricted to the analysis of fracture spacing and strain distributions. Any parameter that changes with distance can be sampled along lines and analysed based on Kuiper’s test.

Given the importance of fluid-flow for many applications (e.g. hydrocarbon migration and production, waste management, groundwater contamination, formation of ore deposits, risk management in underground excavation), a potential field of research is the heterogeneity of fluid flow in the crust. Great potential for this type of study lies in long (>10 km), straight sections across the European Alps as provided by recent large-scale railway tunnel projects (e.g. Gotthard Tunnel between Switzerland and Italy, Brenner Basis Tunnel between Austria and Italy, Mondane Tunnel between France and Italy). Several of these tunnels with >50 km length are under construction now. These tunnels are either excavated by blasting with
round-lengths between 1 and 10 metres or continuously by using large-scale millers (TBM). Traditionally geologists continuously record the position, orientation, thickness and fill of all observed structures, as well as lithology, rock class and amount of seepage, both at the front of the tunnel and at its walls and ceiling. This is done for every new section of tunnel (usually in 1 to 10 m intervals) before shotcrete and other ground support measures are applied. In-situ stress and strain measurements are usually taken in regular intervals along the section. Thus, the collected data provide high resolution one-dimensional samples of a variety of parameters across different tectonic (usually compressive or strike-slip) regimes. Heterogeneity analysis of these data would certainly increase our understanding of rock deformation and is bound to yield novel findings on the spatial distribution and localisation of deformation in orogens. In particular the relationships between distribution of major faults (with cataclastic fault core) and associated damage zones and seepage (fluid flow) would be of major interest in both scientific and industry communities.

8.6 Conclusions

Extension related deformation of sedimentary rocks was studied in detail over a wide range of scales in three study areas.

To record the entire extension accommodated by brittle structures (i.e. normal faults and veins), multiple scan-line data-sets of different length and resolution were collected in each study area. These lines record the most important properties of brittle deformation, which are position, orientation, and displacement of each structure.

To analyse the spatial heterogeneity of the collected data at different scales (i.e. bed-scale, cliff-scale and map-scale), a new non-parametric method based on Kuiper’s test was developed. This method compares the observed distribution of fractures or strain with a uniform distribution. The resulting measure of heterogeneity $V \to 0$ for uniform distributions and $V \to 1$ for maximum heterogeneity. The analysis,
carried out for the three study areas, shows that

1. heterogeneities of fracture and strain distributions in an area are not necessarily the same (i.e. one can be high whilst the other is low),

2. initial heterogeneities are strongly dependent on lithology and mechanical heterogeneity of the deformed rocks (e.g. layering of rocks with different stiffness), and

3. heterogeneities appear to evolve with increasing strain.

These differences are particularly well observed by comparing the deformation in the inter-bedded, but mudstone dominated, sedimentary sequence at Kimmeridge Bay with the massive platform carbonates at the Maltese Islands. Extensions in both areas are moderate (<10%), which means that early (low strain) deformation patterns are well preserved. The interbedded carbonates and mudstones along the Kilve-Lilstock section show a deformation behaviour that is in between the observed damage of the low-stiffness Kimmeridge sequence and the massive carbonates of Malta.

At Kimmeridge Bay initial deformation is accommodated by random to slightly anti-clustered tensile failure of carbonate beds and distributed shearing of the shale layers across the entire region. Displacements are small. This results in low heterogeneity of both fracture and strain distributions. Increase in regional extension leads to opening of some of the earlier tensile fractures (veins) to form pull-apart structures causing an increase in strain heterogeneity compared to a unchanged low heterogeneity of fracture spacing. Further increase in regional extension allows formation of a network of small faults, most of which form where earlier strain localisation had taken place. The fault-strain accommodated by these early faults is fairly uniformly distributed, bringing the strain-heterogeneity back down to low values. Thus, at Kimmeridge Bay an early evolution from low to higher (bed-scale) strain heterogeneity is observed which is followed by distributed (map-scale) faulting accommodating a fairly uniform fault-strain across the region.
On the Maltese Islands virtually all the brittle deformation is accommodated by faults with displacements ranging from cm to >100 m. Initial deformation produces a network of small faults that are randomly distributed or weakly clustered. Increase in extension causes early localisation of strain within km-wide zones that are separated by lower-strain zones that preserve their initial low deformation. Heterogeneity is somewhat lower in the higher-strain zones than in the lower strain zones but shows the same signature for both strain distribution and fault spacing. As extension increases further localisation of strain occurs that eventually leads to a horst-and-graben geometry with 2 major graben-bounding faults with throws >100 m.

The Kilve-Lilstock section is a higher strain example (≈25%) of extensional deformation. The sequence consists of interbedded carbonates and shales in a ratio of $\frac{1}{5}$. Bed-scale deformation in this section is significantly different from the sequence at Kimmeridge Bay with a carbonate/shale ratio of $\frac{1}{13}$. Early extension is highly localised in narrow zones preserving large proportions of virtually undeformed host-rock in between. Initial (bed-scale) deformation is taken up by distributed deformation within the more ductile mudstone layers, whilst it causes tensile failure in the stiffer carbonate beds. This is reflected by high bed-scale heterogeneity of early fracture and strain distributions. Increase in extension causes increasing localisation of strain in the shales forming small faults. Small displacements on these faults are accommodated by bending of the carbonate beds associated with the formation of more tensile fractures (veins). Eventually the faults break through to facilitate slip, generating a random network of faults. Further increase in extension is accommodated dominantly by slip on the fault system associated with minor thickening of veins in the surrounding damage zones. These (early) faults are randomly distributed and they accommodate a uniformly distributed extension across the studied section. This is reflected by low (map-scale) heterogeneities of both, distribution of, and strain accommodated by faults.

These observations suggest that the spatial organisation of brittle deformation is
strongly dominated by three factors:

1. Lithology

2. Strength heterogeneity (e.g. mechanical layering)

3. Strain magnitude.

It has been shown that extension generally evolves from more distributed to more localized deformation with increasing strain. This implies that frequency scaling-relationships also change from higher power-law exponents at low strains to lower exponents at higher strains. In other words small-scale deformation may dominate an extensional system at low strains whilst the largest faults will govern a system at high strains.

These interpretations are supported by estimates of the amount of extension accommodated by small and large structures in the three study areas. It has been shown that faults and veins with displacements <10 m can account for 20% to 70% of the total regional extension and that the relative importance of small structures in accommodating the regional extension depends on the total extension of the system. Large (seismically resolvable) faults accommodate only 30% of the total extension within low-strain zones at Malta, whilst they account for 60% within higher-strain zones. At similar strains large faults accommodate 65% of the total extension at Kimmeridge whilst large faults record 80% of the total extension in the higher-strain section at Kilve. Thus, high-resolution seismic data may resolve most of the total extension in high-strain regions but will significantly underestimate the extension in lower-strain areas. This has to be accounted for when estimating the total seismic moment, stretching factors, or size and frequency of faults (and veins) below seismic resolution.

It has been shown that one-dimensional estimates of extension are adequate for regions that approximately conform to pure-shear, plane-strain conditions. In this case the horizontal longitudinal extension perpendicular to the mean fault trend corresponds to the maximum principal extension. In regions that show plane-strain
conditions, but a preferred down-throw direction, deformation may show a signif-

cant rotational (simple-shear) component. This causes a rotation of the direction

of maximum principal extension, which means that horizontal, one-dimensional es-

timates of extension underestimate the maximum extension. In these cases it is

better, to analyse the strain in two-dimensional, vertical sections.

In regions where a significant variation in fault trends and/or slip-directions is ob-

served, strain will depart from plane-strain conditions. The out-of-plane components

of extension imply that one-dimensional, horizontal estimates will underestimate the

maximum principal extension.

A tensor method was developed that permits three-dimensional strain analysis from

data collected along straight lines. The method is very useful for assessing the “three-

dimensionality” of extension in a region and thus allows testing of one-dimensional

extension estimates. It has been shown that extensional deformation at Kimmeridge

Bay and in the North Malta Graben system conforms to pure-shear conditions,

whilst the Kilve-Lilstock section is better represented by a two-dimensional domino-

style deformation model.

Based on field-observations along the Kilve-Lilstock section a qualitative model for

fault-related folding (drag structures) was developed and tested with a simple elastic

model. It has been shown that:

1. Normal faults may display normal and reverse drag and the combination of

both geometries at different positions relative to the tip line and the position

of nucleation of the fault.

2. Normal drag is likely to occur close to the tip line as it is produced within the

process zone of a (propagating) normal fault.

3. Reverse drag and superimposed normal and reverse drag are likely to occur

closer to the centre of a fault because these geometries form in response to slip

on the fault.
Although slip on any (blind) normal fault may to some extent show drag structures, these appear to be particularly well developed in interbedded lithologies that show a (strong) mechanical layering.
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